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**Petrogenesis and Tectonic Evolution of the Lake of the Woods Greenstone Belt, Western
Wabigoon Subprovince, Ontario, Canada**

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**A thesis submitted in partial fulfilment of the requirements for the Degree of Doctor of
Philosophy to the Ottawa Carleton Geoscience Centre, Department of Geology, University of
Ottawa, Ottawa, Ontario.**

May 1998



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ABSTRACT

The Neoarchean Lake of the Woods greenstone belt is located immediately south of Kenora, Ontario within the western part of the Wabigoon Subprovince of the Superior Province. The supracrustal rocks have been subdivided into 3 major lithostratigraphic assemblages. The Lower Keewatin is the oldest assemblage with a zircon U-Pb age of 2738 ± 2.0 Ma. It consists of Mg and Fe tholeiitic metabasalts (Mg# of 30-65) with flat to slightly enriched rare earth elements (REE) $[(La/Yb)_{CN} = 0.9-1.6]$, elevated concentrations of large ion lithophile elements (LILE), and undepleted high field strength elements (HFSE). The tholeiitic rocks do not appear to have exact modern analogues but do have some similarities to primitive arc and back arc basin basalts consistent with their having been formed in an Archean destructive margin setting. The Upper Keewatin Assemblage disconformably overlies the Lower Keewatin Assemblage. It consists of several volcanic and sedimentary groups. Two geochemically distinct volcanic suites are recognised. The dominant suite consists of compositionally diverse calc-alkaline metabasalt to metarhyolite. Felsic volcanic rocks from this suite yield U-Pb zircon ages ranging from 2712-2723 Ma. The suite exhibits fractionated REE $[(La/Yb)_{CN} = 4-35]$, elevated LILE and Nb and Ta depletion anomalies which are very similar to modern island arc volcanic rocks developed in proximity to continental margins. A coeval suite of mafic to ultramafic volcanic rocks consists of tholeiitic metabasalt, komatiitic metabasalt and metakomatiite (Mg# of 45-80). It exhibits depleted LREE $[(La/Yb)_{CN} = 0.6-1.0]$, slightly elevated LILE and undepleted to slightly enriched HFSE suggesting this geochemically primitive suite was produced by partial melting of a refractory mantle source, perhaps as second stage mantle melts. The Electrum assemblage unconformably overlies the Upper Keewatin Assemblage in the northern part of the belt. It

consists of metaconglomerates, metasandstones and minor metavolcanic rocks having detrital ages indicating deposition after 2699 Ma. The volcanic rocks consist of shoshonitic meta-andesite and calc-alkaline metarhyolite with highly fractionated REE $[(La/Yb)_{CN} = 12-112]$, highly enriched LILE and pronounced Nb, Ta and Ti depletion anomalies. They are geochemically similar to modern shoshonites having a continental influence formed in the late stages of a mature island arc or a continental arc settings.

The Electrum shoshonites are part of a chemically diverse suite of potassic igneous rocks emplaced at a late stage in the tectonic evolution of the western Wabigoon Subprovince. The suite is also comprised of silica undersaturated ultrapotassic rocks of the Falcon Island pluton (2695 Ma) and silica saturated monzodiotitic intrusions such as the Don Lake stock. In general the rocks of the suite have elevated Mg#, Al, K, Sr, Rb, Ba, LREE, depleted HREE and HFSE, and are similar to late stage subduction-related potassic suites of the Proterozoic and Phanerozoic. The elevated LILE, Mg #, Ni, and Cr of the suite, in conjunction with depleted $^{143}Nd/^{144}Nd$ isotope ratios for the ultrapotassic Falcon Island pluton, indicates derivation from metasomatized mantle. Trend lines on bivariate plots provide vectors to a common parental composition most similar to the Falcon Island melasyenite and Don Lake monzodiorite. At Falcon Island mixing was between end-members with strongly to moderately depleted mantle Nd isotopic signatures, while the shoshonites and calc-alkaline volcanic rocks of the Electrum Assemblage have more enriched Nd isotopes indicating mixing could either have been between heterogeneous mantle melts with depleted to enriched isotopic signatures or by assimilation of enriched material in the crust.

The Falcon Island pluton consists of nepheline-bearing, alkali feldspar syenite and alkali feldspar melasyenite with lessor amounts of clinopyroxenite, monzodiorite, and peraluminous

alkali feldspar syenite. The magma evolved with increasing $P(H_2O)$ indicated by clinopyroxene cores mantled by hornblende, and hypersolvus perthitic alkali feldspar phenocrysts followed by subsolvus interstitial plagioclase, alkali feldspar and nepheline. An "aegirine-enrichment trend" is evident in the clinopyroxenes as progressive depletion in Mg and Ca, and enrichment in Fe and Na (both as zoning in individual grains and in the successively fractionated units).

Decreasing $Fe^{3+}/(Fe^{2+} + Fe^{3+})$ accompanied by increasing $Fe/(Fe + Mg)$, Ti, and ^{IV}Al content in the biotite of the evolved rock units indicates progressive magmatic reduction and Al enrichment of the magma with fractionation. A genetic linkage of gold mineralization to oxidized alkaline intrusions has been proposed on the basis of mineral chemical studies from the Abitibi Subprovince. The lack of spatially associated gold mineralization with the progressively reduced Falcon Island pluton, despite its proximity to a potentially favourable zone of deformation, supports this hypothesis.

Nd isotopic data from a compositional spectrum of representative rocks from the major stratigraphic units of the belt indicates a major change in the isotopic signature of mantle sources at about 2.7 Ga. The restricted range of positive ϵ_{Nd} values for tholeiites of the Lower Keewatin Assemblage, and calc-alkaline, tholeiitic and komatiitic volcanic rocks of the Upper Keewatin Assemblage (i.e., +1.1 to +2.1), indicates derivation from relatively homogeneous depleted mantle. An ϵ_{Nd} value of -0.9 for wacke, in conjunction with detrital zircon ages to as old as 3.0 Ga, indicate the provenance sources for the Upper Keewatin Assemblage turbidites involved a component from a long-term isotopically enriched "ancient crustal" source, such as the Winnipeg River Subprovince (2.7 to 3.0 Ga). Isotopic heterogeneity in the post 2.7 Ga mantle is demonstrated by enriched ϵ_{Nd} values (i.e. -0.4 & +0.4) in shoshonitic and calc-alkaline volcanic

rocks of the Electrum Assemblage and depleted values (avg. $\epsilon_{Nd} = +2.2$) in the ultrapotassic rocks of the Falcon Island pluton. Systematic increase of ϵ_{Nd} values from +1.8 to 3.1 in the successively evolved units of the pluton suggests mixing of mantle end member components with variable degrees of Sm/Nd depletion.

Pb isotopes of leached alkali feldspars from the Falcon Island pluton have a $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ isochron with an age of 2700 ± 100 Ma, an initial $^{207}\text{Pb}/^{204}\text{Pb}$ of 14.52 and an initial $^{206}\text{Pb}/^{204}\text{Pb}$ of 13.29 indicating a mantle source reservoir similarly depleted in $^{206}\text{Pb}/^{204}\text{Pb}$, but slightly more enriched in $^{207}\text{Pb}/^{204}\text{Pb}$, than mantle-derived mafic and ultramafic magmas from the Wabigoon and Abitibi subprovinces. Comparison of Nd, Pb and Sr isotope data from a number of well documented potassic suites across the southern Superior Province indicates heterogeneity was widespread in the post 2.7 Ga mantle. The correlation of this isotopic data suggests mixing between a juvenile depleted mantle component and an enriched end-member with isotopic characteristics approaching those of the ancient granitic terrain of Winnipeg River Subprovince and the Opatika and Pontiac metasedimentary subprovinces. This type of isotopic heterogeneity is most readily explained by metasomatism of a mantle wedge by the influx of fluids derived from partial melting of ancient crustal sediments subducted beneath the Neoproterozoic crust.

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Introduction

1.1 Purpose

My primary goal in writing this thesis is to better understand the petrogenesis and tectonic evolution of the Neoarchean greenstone belts of the southern Superior Province. The Lake of the Woods greenstone belt is an ideal area for this type of study because it is easily accessible and well exposed. The belt was the focus of a recent comprehensive geoscience program undertaken by the Ontario Geological Survey which included detailed mapping and regional synoptic mapping, regional structural and metallogenic synthesis and lithogeochemical and geochronological analysis (Ayer 1987, 1990, 1991, Ayer and Buck 1989, Davis and Smith 1991, Sanborn-Barrie 1991, Ayer and Davis 1997). The results of this work provide an excellent database from which to construct stratigraphic correlations, and to perform geodynamic analysis. Using this framework, the geochemical and isotopic evolution of the belt is evaluated and compared with other well documented greenstone belts in the Superior Province, and also with Phanerozoic analogues in which the plate tectonic processes and their rock products are relatively well understood. This analysis and the comparisons, provide a model for Neoarchean tectonic evolution which can be used to place the mineralizing environments into a geodynamic framework and thereby help focus exploration for new mineral deposits.

Radiogenic isotopes are particularly useful as petrogenetic tracers when the chemical composition of potential end-members reservoirs and assimilants are well characterised and

when a precise time reference is available. Thus the Nd and Pb isotopic studies of the Lake of the Woods belt provide additional insight into the characteristics of the Neoproterozoic crust and mantle and provides the basis for comparisons of crust-mantle reservoirs across the southern Superior Province. Study of the temporal and spatial variations in mantle composition permit a better understanding of crust-mantle interaction and aid in the characterisation of different tectonic environments. Similar to the approach that has led to the recognition of Phanerozoic mantle heterogeneities in terms of the radiogenic Sr, Nd, and Pb isotope compositions (Zindler and Hart 1986), the isotopic characteristics of the Archean mantle can be evaluated by analysis of directly derived rock products.

A secondary goal of the thesis is to document the petrogenesis of a coeval suite of compositionally diverse extrusive and intrusive potassic rocks and to compare them with other potassic suites of the Superior Province, the Proterozoic and the Phanerozoic. In many localities, potassic suites have important economic implications because of their close spatial and temporal association with gold mineralization. The oxidation state of the Abitibi alkaline suites is variable and is also considered to play an important role in mineralization. Comparison of the mineral chemistry of the Falcon Island pluton with that of other syenitic complexes, including oxidized complexes in the Kirkland Lake gold camp, thus provides insight into the geodynamic and mineralization processes associated with gold mineralization in the Neoproterozoic.

1.2 Organization

This thesis is presented in six chapters. Chapters 2, 3, 4 & 5 have been written as manuscripts for publication. To a certain extent these chapters overlap, particularly with regard

to geological descriptions, but considerable effort has been made to minimize this for the thesis. Chapter 2 presents the geological and geochronological framework for the greenstone belt and provides the foundation for the following chapters. This stratigraphic framework is then utilized to characterize the geochemical patterns of the major units and to develop, by comparison with modern terrains, a plate tectonic model for the evolution of the belt. Chapter 3 is an in depth analyses of the Falcon Island ultrapotassic pluton which has a large compositional range and degree of fractionation. The mineral chemistry of the rock forming minerals of the fractionated series reveals the variation in magmatic conditions and oxidation states which relate to the favourability of alkaline intrusions to generate gold mineralization. Chapter 4 expands the study of the chemically diverse suite of ultrapotassic and potassic rocks in the Falcon Island pluton, to also include coeval monzodioritic rocks in the Don Lake stock and shoshonitic volcanic rocks in the Electrum assemblage. The chapter focuses on details of the petrogenesis and explains the compositional diversity of this late-tectonic, mantle-derived suite. Chapter 5 applies the Nd isotope data for the major stratigraphic units of the belt and the Falcon Island pluton and the common Pb isotopes of the Falcon Island pluton. The isotopic data is utilized to further quantify the hypothesised igneous sources of Chapter 2 and consider the signatures and causes of secular changes in the nature of mantle within the belt, and across the southern Superior Province.

1.3 Co-authorship

Chapters 2 through 5 have been written as manuscripts for publication. A number of co-authors provided supervisory, instructional, or interpretational roles, commented on the initial drafts of the manuscripts and helped with funding of the analytical work. As senior author, the

original concepts, writing, primary interpretation of data, drafting, submission of manuscripts, editing and communication with selected journals has been my responsibility. The list of co-authors, their affiliations and the journals to which the manuscripts have been, or will be submitted, is as follows:

Chapter 2: 1997. *Precambrian Research*, v.81, p.155-178. Co-author Dr. D.W. Davis, Jack Satterly Geochronology Laboratory, Royal Ontario Museum, 100 Queen's Park, Toronto, Ontario M5S 2C6;

Chapter 3: *Canadian Mineralogist*, v.36, p.1-18, 1998. No co-author.

Chapter 4: Submitted. Co-author Dr. A.D. Fowler, Ottawa Carleton Geoscience Centre and Department of Earth Sciences, University of Ottawa, 140 Louis Pasteur, Ottawa, Ontario, K1N 6N5.

Chapter 5: Submitted. Co-author Dr. J. Dostal, Department of Geology, Saint Mary's University, Halifax, Nova Scotia, B3H 3C3.

1.4 Statement of Original Contributions

The following is my perception of the main "original contributions to knowledge" that are represented in this thesis.

- 1) This work represents the first comprehensive study on the petrogenesis of the Lake of the Wood Greenstone belt including establishment of a formalized stratigraphy with an absolute age framework based on extensive mapping and precise U-Pb zircon geochronology. It thus compliments a number of existing academic studies which cover various other aspects of selected portions of the belt (e.g. Zielke 1974, Carr 1980, Brown 1984, Stix 1985, Smith 1987).
- 2) It is also the first application of geodynamic analysis to the major suites of the Lake of the Woods greenstone belt and provides comparisons with the patterns of modern plate tectonic settings. This study is based on an extensive lithogeochemical database collected throughout the belt and representative samples selected for high precision trace element, Nd and Pb isotopic analyses.
- 3) The thesis incorporates a detailed study of a diverse suite of late-tectonic, potassic igneous rocks occurring both as volcanic and plutonic units in various parts of the greenstone belt. The suite is characterized utilizing microprobe-determined mineral chemistry, whole rock major and trace element chemistry and Nd and Pb isotopes. The study includes an in-depth analysis of the petrogenetic history, and geodynamic implications for the whole suite, and more specifically, the evolution of the oxidation state of the Falcon Island pluton magma using mineral chemistry. This has important economic implications because mineral chemical studies of similar suites from the Abitibi Subprovince suggest that a genetic association of gold mineralization with the late-tectonic potassic suites may be related to their oxidation state.
- 4) Finally, isotopic results from the Lake of the Woods greenstone belt are used to show the existence of a fundamental change in the nature of the mantle from being homogeneously

depleted in the pre 2.7 Ga igneous units to becoming a more heterogeneous, depleted to enriched mantle source, for the post 2.7 Ga potassic suite. This isotopic data is then compared with the results of isotopic studies from a number of other greenstone-, granitic- and metasedimentary-dominated subprovinces across the southern Superior Province in order to erect a plate tectonic model for mantle mixing in the Neoarchean, analogous to that documented in the modern mantle.

Chapter 2

Neoproterozoic Evolution of Differing Convergent Margin Assemblages in the Wabigoon Subprovince: Geochemical and Geochronological Evidence from the Lake of the Woods Greenstone Belt, Superior Province, Northwestern Ontario.

2.1 Introduction

Plate tectonic theory has been recognized as a major factor in the construction of Archean terranes for over two decades. Plate convergence resulting in accretion of subprovinces to form the Superior Province has been postulated to have formed island-arc complexes represented by the Uchi, Wabigoon, Wawa and Abitibi subprovinces and accretionary prisms represented by the English River and Quetico subprovinces (Krogh and Davis 1971; Langford and Morin, 1976; Blackburn, 1980; Card, 1990). These interpretations were based on subprovince scale synthesis. Recent, more detailed evaluations are beginning to show complex histories, with for example, the Abitibi and Wawa subprovince "arcs" representing a broad diversity of geodynamic settings (e.g., Jackson et al., 1994; Sage et al., 1996).

Intra-subprovince diversity is also evident in the Wabigoon Subprovince. This chapter will illustrate that a variety of different assemblage types occur within the Lake of the Woods greenstone Belt on the northwestern margin of the Wabigoon Subprovince (Fig. 2.1). In addition, western Wabigoon Subprovince volcanic sequences such as the Upper Keewatin, the

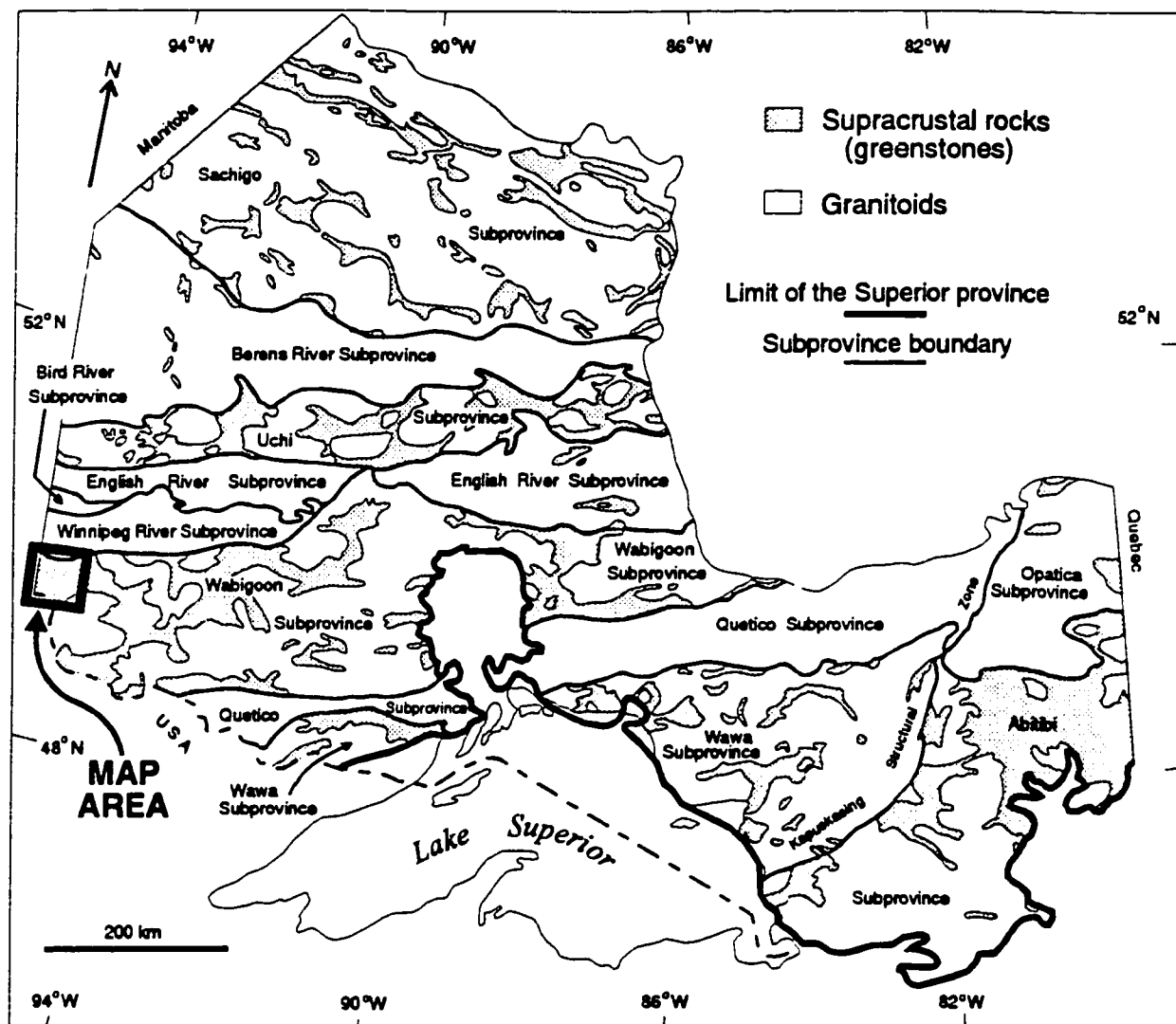


Fig.2.1. Geological map of the Superior Province showing distribution of subprovinces and location of the map in Figure 2.2.

Kakagi Lake group, the Lower Wabigoon group, the Neepawa group and the Handy Lake group (Blackburn et al., 1991) contain island-arc-like, calc-alkaline fractionation series rock suites including andesites with plagioclase, pyroxene and hornblende phenocrysts, similar to Phanerozoic analogues. The abundance of these Archean andesites, which have been poorly documented to date, is significant because their characteristics and distribution in greenstone belts can be correlated with similar Phanerozoic rocks.

The Lake of the Woods greenstone belt is a good starting point for this type of analysis because it has been the focus of a comprehensive program of detailed and regional mapping and lithogeochemical sampling projects (Johns, 1987; Ayer et al., 1987, Ayer, 1987; Ayer, 1990; Ayer, 1991; Ayer and Buck, 1989; Smith, 1987; Morrice, 1988; Morrice, 1989), structural studies (Sanborn-Barrie, 1991) and geochronological sampling (Davis and Smith 1991; this chapter). These projects have provided an excellent framework for the stratigraphic and geodynamic interpretation of this structurally complex area.

2.2 Analytical methods

The geochemical database for the area includes over 500 whole rock analyses of the major element oxides and selected trace elements in the above cited reports. Wherever possible samples of least altered flows with a minimum of phenocrysts were selected to ensure that the compositions are as close as possible to the original magmatic liquids. The samples were screened for high volatile content and alkali alteration by rejection of samples outside of the igneous spectrum of Hughes (1972) or an alteration index $[A.I. = (K_2O + MgO)/(K_2O + Na_2O + CaO +$

MgO)] x 100; Date et al., 1983) of less than 25 or greater than 65 for non-ultramafic samples with the exception of the Zigzag formation samples (discussed further in the geochemistry section below). Details of the techniques, precision and accuracy are indicated in Appendix I. Duplicates and standards submitted with the samples indicate precision and accuracy values were generally within 5% (at 2 σ) for values greater than 10 times detection limits.

Results of the U-Pb analyses on zircons and titanites from selected samples and the resultant ages are presented in Table 2.2. Analyses were carried out by the isotope dilution method. Details of the procedures are summarized in Davis and Smith (1991).

2.3 Geological setting

The Lake of the Woods greenstone belt (LWGB) is situated along the northwestern margin of the Wabigoon subprovince, a northeast-trending metavolcanic-plutonic belt about 150 km wide exposed over a length of about 900 km in the Superior Province (Fig 2.1). The subprovince is a collage of metamorphosed, supracrustal assemblages intruded by granitic batholiths. A central granitic area divides the subprovince into eastern and western regions (Blackburn et al., 1991). It is bounded to the north by the Winnipeg River and English River Subprovinces, and to the south, by the Quetico Subprovince.

Over the past century a variety of stratigraphic models and nomenclature have been proposed in the LWGB. Lawson (1885) defined the Keewatin Series as a series of volcanic rocks, later modified (Thomson 1937) to a two fold subdivision, an older Keewatin Series of

predominantly metavolcanic rocks unconformably overlain by a younger series of predominantly metasedimentary rocks equated with the Timiskaming of the Abitibi subprovince.

The belt is divided into three supracrustal assemblages: (1) the Lower Keewatin Assemblage, (2) the Upper Keewatin Assemblage and (3) the Electrum Assemblage. Assemblages are defined (Jackson et al., 1994) as regional map units which share some or all of: lithology, geochemistry, facies associations, geophysical signatures, structural style, and age. The units contained within an assemblage need not be stratigraphically related, and an assemblage may be in either fault or depositional contact with other assemblages.

Figure 2.2 outlines the distribution of major stratigraphic subdivisions for the LWGB. Stratigraphic nomenclature, rock units, interpreted depositional environments and U-Pb ages are listed in Table 2.1.

Although all the rocks in the area have been metamorphosed, their original character is commonly discernable on the basis of primary structures, petrography and geochemistry, thus in the interest of brevity the prefix meta has been omitted.

2.3.1 Lower Keewatin Assemblage

Mapping and geochronology demonstrates the Lower Keewatin Assemblage occupies the lowermost stratigraphic position within the LWGB. The assemblage consists of five geographically isolated mafic volcanic groups (Fig. 2.2). They are correlated on the basis of stratigraphy, lithological and geochemical similarities and geochronology. The maximum preserved thickness is about 10 km. However, the assemblage was originally thicker because the

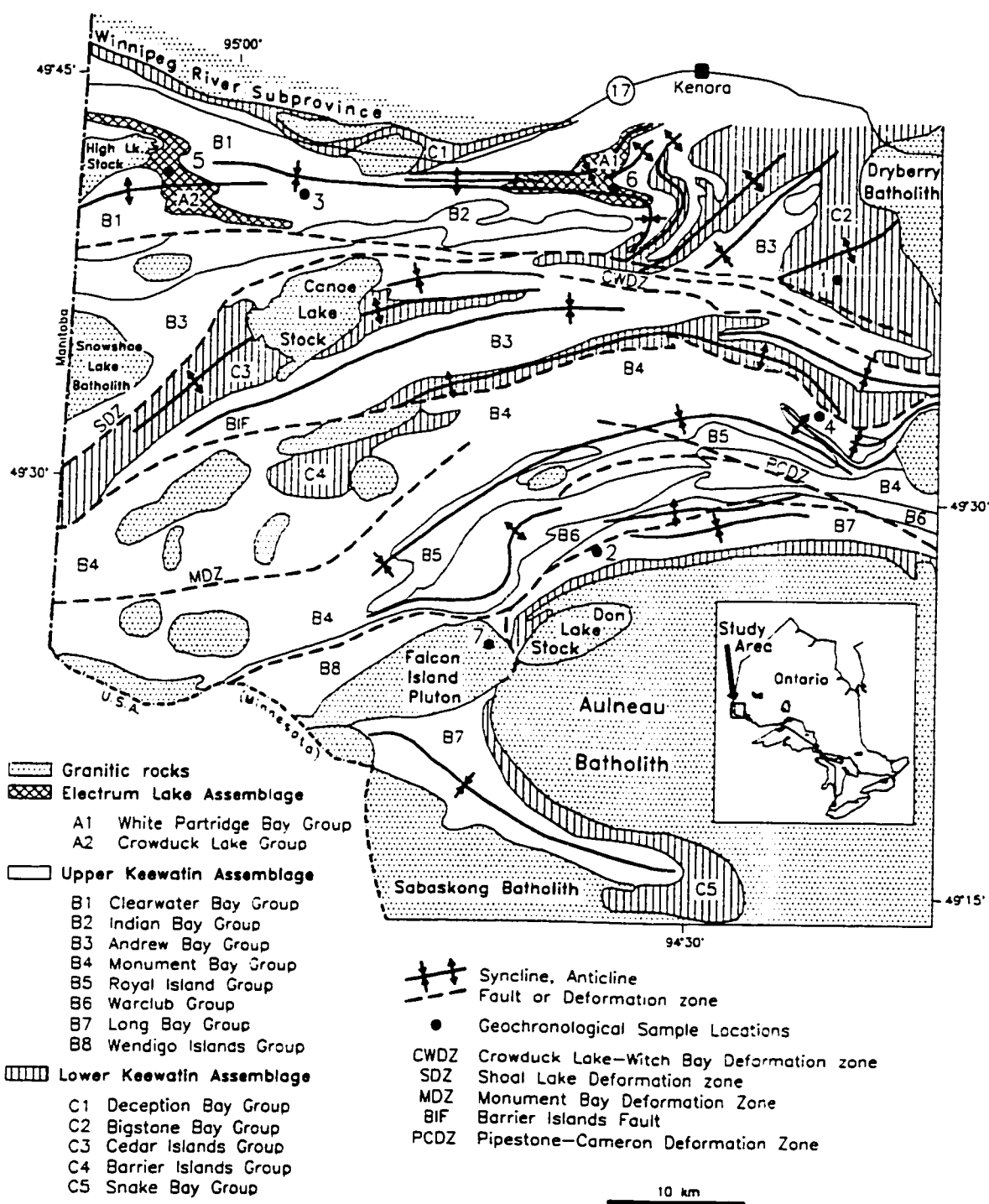


Fig. 2.2. General geology of the Lake of the Woods greenstone belt showing geochronological sample locations

Table 2.1. Lake of the Woods greenstone belt stratigraphic units (geochronology by D. Davis see 'Blackburn et al. 1991, ²this paper).

<u>Stratigraphic Unit</u>	<u>Dominant Lithologies</u>	<u>Subordinate Lithologies</u>	<u>Deposition Environment</u>	<u>Age (Ma)</u>
ELECTRUM ASSEMBLAGE				
Crowduck Lake Group	Conglomerates, arenites, wackes, siltstones.	Calc-alkaline mafic to felsic flows and pyroclastics.	Proximal subaerial to submarine.	<2699 ²
White Partridge Bay Group	Turbidites, conglomerates, mudstones, siltstones.	Felsic pyroclastics.	Distal to proximal submarine.	<2709 ²
UPPER KEEWATIN ASSEMBLAGE				
Clearwater Bay Group	Calc-alkaline mafic to felsic flows & pyroclastics.	Clastic & chemical sediments, tholeiitic mafic flows.	Shallow submarine to subaerial.	2719 ²
Indian Bay Group	Tholeiitic mafic flows	Calc-alkaline felsic volcanics.	Shallow submarine	2723 ¹
Andrew Bay Group	Calc-alkaline mafic to felsic volcanics.	Clastic & chemical sediments & tholeiitic mafic flows.	Shallow submarine	
Monument Bay Group	Calc-alkaline mafic to felsic volcanics.	Tholeiitic to komatiitic flows, conglomerates, turbidites, & chemical sediments.	Shallow submarine to subaerial	2715 ¹ 2719 ²
Royal Island Group	Turbidites.	Conglomerates & mafic to felsic volcanics.	Distal to proximal submarine.	
Warclub Group	Turbidites.	Tholeiitic to komatiitic flows & clastic sediments.	Distal to proximal submarine.	
Long Bay Group	Calc-alkaline mafic to felsic volcanics.	Clastic and chemical sediments.	Shallow submarine.	2738 ²
Windigo Islands Group	Calc-alkaline mafic to felsic volcanics.			
LOWER KEEWATIN ASSEMBLAGE				
Deception Bay Group	Tholeiitic mafic flows.	Komatiitic flows & calc-alkaline felsic volcanics.	Deep submarine.	>2732 ¹
Bigstone Bay Group	Tholeiitic mafic flows.	Felsic volcanics.	Deep submarine.	
Cedar Island Group	Tholeiitic mafic flows.		Deep submarine	
Barrier Islands Group	Tholeiitic mafic flows.		Deep submarine	>2732 ¹
Snake Bay Group	Tholeiitic mafic flows.		Deep submarine	

present basal contacts are either intrusive (in domal structures wrapping around the Aulneau and Dryberry batholiths for the Snake Bay and Bigstone Bay groups respectively), or are in fault-bounded culminations (e.g., the Deception Bay group).

The assemblage consists of pillowed to massive mafic flows which are predominantly aphyric with local plagioclase megacrystic basalt units high in the section. Variolitic flows, vesicular flows, and columnar-jointed mafic flows also occur locally in the upper stratigraphic levels. A general upward change from non-vesiculated flows to vesiculated flows in the upper stratigraphic parts suggests either shoaling upwards into shallower water or secular variation in magma volatile content.

It is commonly difficult to constrain the precise age of mafic volcanic rocks because of their scarcity of zircons. A coeval pegmatitic gabbro yielded a U-Pb zircon and baddeleyite age of $2732 \pm 4/-3$ Ma from the Katimiagamak formation in the Kakagi Lake greenstone belt southeast of the study area (Davis and Edwards 1986). The Katimiagamak formation is interpreted to be correlative with the Snake Bay group, located in the southern part of the LWGB (Fig 2.2).

A sample of a medium-grained mafic flow in the upper section of the Bigstone Bay group (DD90-28, Table 2.2; location #1 on Fig. 2.2) contained a few tiny zircons. Analyses of two fractions give data points which are concordant and agree within error at 2737.9 ± 2.0 Ma (Fig. 2.3a). The $^{208}\text{Pb}/^{206}\text{Pb}$ ratios in the zircons of this mafic volcanic sample are distinctly higher than are evident in the zircons from the felsic volcanic rock samples in the LWGB (Table 2.2). This is a typical characteristic of zircons which have crystallized from mafic rocks because of their higher Th/U ratios (Heaman et al. 1990) and suggests that the grains are not xenocrystic. The age is therefore considered to be representative of the Bigstone Bay group magmatism. The above ages

Table 2.2. U-Pb isotopic data for zircon and titanite

Number of grains	Weight (mg)	U (ppm)	Pb _{com} (pg)	Measured		Corrected		²⁰⁷ Pb/ ²⁰⁶ Pb age (Ma)
				²⁰⁷ Pb/ ²⁰⁴ Pb	²⁰⁸ Pb/ ²⁰⁶ Pb	²⁰⁵ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	
DD90-28 Bigstone Bay group, massive flow								
3 ab	0.001	235	3	588	0.918	0.52662	13.761	2738.0±2.6
1 ab	0.001	184	3	475	1.490	0.53174	13.892	2737.5±3.5
DD90-33 Sunset Channel formation, felsic tuff								
4 ab rnd	0.025	51	11	747	0.173	0.52310	13.517	2719.4±2.2
8 ab euh	0.032	53	8	1327	0.169	0.52149	13.464	2718.1±3.2
7 euh	0.050	57	32	566	0.184	0.52042	13.443	2719.0±1.9
DD90-42 Zigzag Island formation, felsic flow								
1 ab euh clr	0.002	204	11	237.6	0.171	0.52430	13.540	2718.5±2.7
1 ab rnd brn	0.001	841	7	718	0.199	0.52080	13.442	2717.6±2.6
6 rnd brn	0.020	650	9	8682	0.188	0.51712	13.319	2714.0±1.6
DD90-30 Shore Island formation, felsic tuff								
16 ab euh	0.025	46	12	664	0.115	0.53570	14.343	2778.0±4.8
2 ab euh	0.013	45	9	465	0.122	0.54120	14.588	2789.0±3.3
1 ab cir	0.002	89	2	539	0.233	0.54742	15.085	2825.0±3.4
DD90-40 Crowduck group, felsic tuff								
1 ab euh	0.015	67	3	2467	0.007	0.52046	13.286	2669.4±2.2
1 ab rnd	0.004	215	4	1436	0.061	0.53922	14.582	2794.2±2.6
1 ab euh	0.005	55	4	511	0.150	0.54378	14.737	2797.7±2.6
DD90-39 White Partridge Bay group, felsic tuff-breccia								
1 ab frag	0.001	252	3	543	0.072	0.52050	13.351	2707.3±3.8
1 ab frag	0.006	32	1	966	0.137	0.51914	13.329	2708.9±3.4
1 ab rnd	0.002	155	3	636	0.387	0.51863	13.321	2709.±2.3
1 ab euh	0.002	15	1	231.4	0.074	0.51951	13.394	2716±13
1 ab frag	0.002	69	3	331	0.157	0.52062	13.436	2717.5±5.6
1 ab euh	0.010	37	25	115.2	0.176	0.53637	14.430	2785.6±3.7
DD90-34 Falcon Island stock, monzonite								
11 ab clr ti	0.096	40	537	58.3	1.481	0.51869	13.194	2693.5±3.6
6 ab brn ti	0.111	48	1034	46.4	1.378	0.52582	13.382	2694.3±4.6

ti-titanite; ab-abraded; euh-euhedral; rnd-rounded; frag-fragment; clr-clear; brn-brown; Pb_{com}-common Pb, including blank.

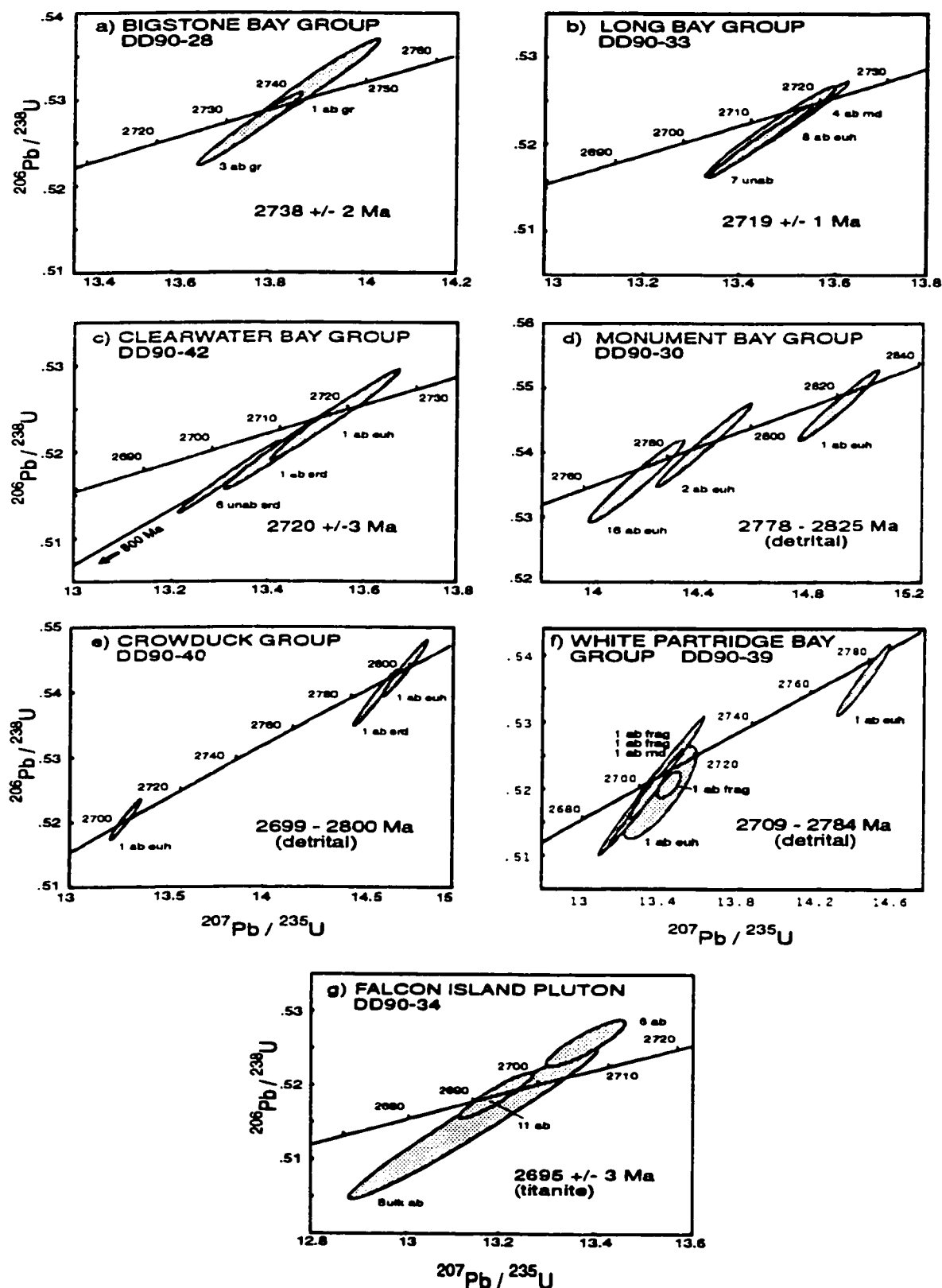


Fig. 2.3. Concordia diagrams with U-Pb analyses for zircon and titanite. Ellipses indicate the 2σ uncertainty. Abbreviations as in Table 2.2.

are similar within the range of error and therefore confirm that the physically separated groups of the Lower Keewatin Assemblage represent the oldest units within the LWGB and are most likely correlative.

2.3.2 Upper Keewatin Assemblage

The Upper Keewatin Assemblage is the most extensive assemblage within the LWGB (Fig. 2.2). The assemblage comprises a diversity of stratigraphically and structurally interdigitated rock types having an estimated maximum stratigraphic thickness of about 10 km. Three distinctive rock associations comprise the assemblage: (1) mafic to felsic volcanic rocks of calc-alkalic affinity; (2) ultramafic to mafic volcanic rocks of komatiitic to tholeiitic affinity; and (3) turbiditic sedimentary rocks. The contact between the Upper Keewatin Assemblage and the underlying Lower Keewatin Assemblage is marked by thin units of graphitic mudstone and fine-grained clastic sediments, which because of an age difference of about 15 million years (see below), is interpreted to be disconformable.

Calc-alkaline volcanic rocks are commonly porphyritic and may contain phenocrysts of plagioclase, clinopyroxene, hornblende and quartz. Volcanological features such as intercalations of epiclastic sediments and highly vesicular pillowed flows suggest shallow submarine eruption and deposition. Thick interbedded debris flows and pyroclastic flows with localized cinder cone and welded tuff deposits indicate localized subaerial eruption.

Five U-Pb zircon ages have been determined from felsic volcanic rocks of the Upper Keewatin Assemblage. Tuff breccia from the lower portion of the Andrew Bay group in the Shoal

Lake area yielded an age of 2723 ± 2 Ma (Davis and Smith, 1991). Zircons in a tuff from the lower part of the Long Bay group (sample DD90-33, location #2 on Fig. 2.2) give data points which are concordant and agree within error, defining a crystallization age of 2719 ± 1 Ma (Fig. 2.3b & Table 2.2). Flows from the Zigzag formation of the Clearwater Bay group (sample DD90-42, location #3 on Fig. 2.2) yielded two zircon populations with an age of $2719.6 \pm 3.2/-2.7$ Ma (Fig. 2.3c & Table 2.2).

Turbiditic units consisting of partial to complete Bouma sequences up to several metres thick (Ayer and Buck 1989) occur in the Warclub and Royal Island groups (Fig. 2.2, Table 2.1) in the uppermost part of the assemblage. Zircons from a felsic debris flow in the Shore Island formation of the Monument Bay group (sample DD90-30, location # 4 on Fig. 2.2), produced data points which are concordant, but which disagree in age (Fig. 2.3d and Table 2.2). The ages of 2778 to 2825 Ma are older than any recognized igneous ages in the LWGB and are most likely from detrital zircons. This observation is supported by the geological evidence in that the Shore Island debris flows conformably overlies turbidites of the Royal Island group and contains clasts of wacke and argillite (Ayer and Buck, 1989).

Calc-alkaline felsic volcanic rocks locally occur within the Warclub and Royal Island groups. Zircons from felsic pyroclastic rocks of the Berry River formation, within the Warclub group just east of the study area, have a U-Pb age of 2712 Ma (Davis and Edwards, 1982).

Subordinate tholeiitic to komatiitic volcanic formations up to 50 km in length by 2 km thick occur as intercalations with calc-alkaline mafic to felsic volcanic units of the Upper Keewatin Assemblage. They consist of aphyric mafic and ultramafic flows with rare intercalations of intermediate tuff and fine epiclastic sedimentary rocks. Pillowed and massive mafic flows are

locally variolitic or vesicular. Ultramafic pillows have very thin selvages and massive ultramafic flows locally have polyhedral-jointing patterns. These mafic to ultramafic units are considered to be coeval with the calc-alkaline volcanism as evidenced by conformable contacts which are either abrupt or marked by fine-grained felsic tuffs and turbiditic wackes (Ayer and Buck 1989; Ayer 1991, Ayer et al. 1991).

2.3.3 Electrum Assemblage

The Electrum Assemblage is restricted to the northern part of the LWGB and is comprised of the Crowduck and the White Partridge groups (Fig. 2.2). Both groups consist of predominantly clastic sedimentary rocks with subordinate volcanic units. Conglomerates bearing granitoid and supracrustal rock clasts occur throughout the Crowduck group, while they only occur in the upper levels of the White Partridge group. The Crowduck group has a locally exposed basal unconformity. Volcanic rocks within the Crowduck group consist of intermediate flows and pyroclastic rocks and felsic flows. The lower part of the White Partridge group consists of turbiditic sandstones and subordinate felsic pyroclastic rocks.

Zircons collected from a debris flow within the Crowduck group (sample DD90-40, location # 5 on Fig. 2.2) have data points which are concordant but which disagree in age (Fig. 2.3e & Table 2.2). The two older zircon ages are close to 2800 Ma and are considered to be detrital. The youngest zircon yields an age of 2699 ± 2 Ma and is also considered to be detrital because its $^{208}\text{Pb}/^{206}\text{Pb}$ ratio indicates an extreme depletion in Th (Th/U of 0.024, as opposed to about 0.5-1.0 in zircon crystallized from felsic magmas). These ratios are normally found in

zircons which crystallized under medium to high grade metamorphism (Krogh and Davis 1972) and suggests that lower crustal rocks were exposed during deposition of the Crowduck group. This zircon places a maximum age limit of about 2700 Ma on the deposition of the Crowduck group.

Zircons collected from a felsic block and ash flow unit within the White Partridge group (sample DD90-39, location #6 on Fig. 2.2) also have concordant data points with differing ages suggesting inheritance. The oldest zircon has an age of 2786 ± 4 Ma, two zircons have an age of about 2717 ± 5 Ma and the three youngest zircons agree within error and define an age of 2709.1 ± 1.6 Ma (62% probability of fit). Such an agreement would normally be taken as good evidence that these zircons date eruption of the rhyolite and therefore White Partridge group sedimentation. However, the zircons all show different U concentrations and $^{208}\text{Pb}/^{206}\text{Pb}$ (Th/U) ratios, suggesting that they originate from different sources and are likely inherited. Therefore the younger age cannot be unambiguously interpreted as dating sedimentation, but it is certainly an older limit on the age of deposition of the sediments.

The above U-Pb zircon isotopic evidence indicates a wide range of ages and diverse rock types in the provenance area. Detrital ages of up to 100 My older than any recognized magmatic event in the LWGB located within the Electrum Assemblage and in the Shore Island formation of the Upper Keewatin Assemblage, strongly suggests a provenance derived in part from the older terrane of the Winnipeg River Subprovince.

2.3.4 Plutonic complexes

The Winnipeg River Subprovince is a granitic domain located immediately to the north of the LWGB (Figs. 2.1 & 2.2). It consists of predominantly plutonic rocks spanning a large range of ages, subdivided into two groups: an older (2830-3170 Ma) gneissic sodic group intruded by post tectonic potassic plutons (2660-2709 Ma) (Beakhouse, 1991).

Blackburn et al. (1991) document a variety of ages of intrusive events in and around the LWGB. The early event, consists of predominantly tonalitic intrusions [e.g., the Sabaskong batholith ($2723-2724 \pm 2$ Ma), or early phases of the High Lake stock (2727 ± 2 Ma), and the Aulneau batholith ($2717 +3, -5$ Ma)], was contemporaneous with Upper Keewatin Assemblage volcanism. A second event consists of tonalite to granodiorite [e.g., the late phase of the High Lake stock (2711 ± 2 Ma), the Canoe Lake stock (2709 ± 1 Ma) and the late phase of the Aulneau batholith ($2710 +4, -2$ Ma)] and post-dates Upper Keewatin Assemblage volcanism.

A third event is represented by alkaline and granite plutons which appear to be coeval with Electrum Assemblage volcanism. Two fractions of titanite, a brown and a clear fraction, were analyzed from a sample of syenite from the Falcon Island pluton (sample DD90-34, location # 7 on Fig. 2.2). Both fractions have similar isotopic characteristics (Table 2.2). The data point from the brown titanite is slightly above concordia, probably as a result of incomplete dissolution. This should have no effect on the $^{207}\text{Pb}/^{206}\text{Pb}$ age. Th/U ratios of 5.0 and 5.3 (Table 2.2), indicate that the titanite is magmatic and therefore should record the cooling age of the rock. Titanite typically has low U/Pb ratios relative to zircon, so the age results are strongly dependent on the isotopic composition assumed for the common Pb component. If a crustal average is assumed (Stacey and Kramers, 1975), both ages agree at 2694 ± 3 Ma.

2.4 Geochemistry

In order to construct tectonomagmatic models for the evolution of the LWGB, high precision lithogeochemical analyses of a number of volcanic and plutonic rock samples are utilized to characterize the geochemical patterns and thus allow comparison with the orogenic environments of modern tectonic regimes. The geochemical characteristics are summarized below. Representative analyses of the major units are presented in Table 2.3.

The volcanic units in the LWGB have been subdivided into four geochemical suites on the basis of distinctive geochemical signatures and stratigraphic position. They are: (1) the Lower Keewatin tholeiitic suite; (2) the Upper Keewatin calc-alkaline suite; (3) the Upper Keewatin tholeiitic to komatiitic suite; and (4) the Electrum shoshonitic to calc-alkalic suite.

Sample selection by a rigorous screening process involving a 2 step evaluation of alkali elements abundances and elimination of samples with high volatile concentrations (see the Analytical Methods Section) has ensured that the trace elements concentrations of the utilized samples were probably not significantly modified by secondary alteration processes. A study of unaltered and altered Archean volcanic rocks from the Abitibi Subprovince by Ludden et al. (1983) demonstrates that secondary modification of the abundance of the alkaline, REE and HFSE elements is evident as erratic patterns in normalized elemental profiles and is independent of their degree of incompatibility. The overall consistency of the LWGB suite profiles (Fig. 2.6) thus indicates that the trace element abundances of the selected samples were predominantly controlled by magmatic rather than alteration processes.

Table 2.3. Major and selected trace element data for Lake of the Woods greenstone belt samples.

Assemblage Group	Lower Keewatin										Upper Keewatin									
	Bigstone Bay					Deception Bay					Cedar Island					Barrier Island				
	upper part	mid pt.	lower part	upper part	lower part	upper part	mid pt.	lower part	upper part	lower part	upper part	mid pt.	lower part	upper part	lower part	upper part	mid pt.	lower part	upper part	lower part
Sample No.	84-810	93-19	93-20	93-21	93-22	84-817	93-23	93-24	93-25	93-26	86-168	90-1	86S-113	86S-336	86-359	86-110	93-25	87-838	87-839	86-634
Rock type	m.f.	per.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.	m.f.
SiO ₂	48.00	45.50	51.20	50.20	45.70	47.20	49.20	49.20	47.10	48.57	47.10	48.57	47.60	47.50	49.90	46.40	46.40	48.60	48.60	79.10
TiO ₂	1.08	0.48	0.92	1.31	1.57	0.86	0.73	0.50	0.87	0.50	0.87	0.50	0.46	0.57	0.60	1.96	0.92	1.58	0.81	0.04
Al ₂ O ₃	13.40	9.40	17.00	12.00	12.80	14.20	14.30	17.80	14.90	14.90	17.80	14.90	10.60	15.60	13.30	15.10	15.20	14.40	15.60	10.80
Fe ₂ O ₃ *	13.70	13.40	12.10	18.00	15.20	13.50	13.50	10.40	13.47	11.36	11.36	11.36	11.04	11.90	11.90	18.00	11.80	16.70	13.80	17.25
MnO	0.20	0.21	0.12	0.23	0.22	0.20	0.21	0.19	0.21	0.19	0.21	0.19	0.17	0.17	0.15	0.24	0.20	0.37	0.22	0.00
MgO	7.08	16.40	4.83	5.42	6.21	8.23	8.05	9.39	7.57	13.80	8.45	8.45	8.54	8.54	8.54	5.88	4.87	3.85	6.66	0.26
CaO	9.37	9.67	7.16	8.25	7.26	9.29	11.30	10.70	10.74	11.90	9.86	9.86	9.12	9.12	9.12	4.36	11.20	9.33	11.00	0.50
Na ₂ O	2.50	0.97	4.00	1.46	2.71	1.58	1.71	1.42	1.88	0.83	2.89	2.97	3.14	1.69	2.72	3.14	1.69	2.72	1.74	0.34
K ₂ O	0.05	0.09	0.78	0.11	2.57	0.00	0.09	0.17	0.62	0.00	0.02	0.03	0.03	0.03	0.03	0.36	0.09	0.44	0.13	6.38
P ₂ O ₅	0.04	0.08	0.08	0.12	0.15	0.02	0.07	0.04	0.08	0.03	0.01	0.04	0.01	0.04	0.01	0.08	0.01	0.12	0.04	0.01
LOI	2.50	3.15	2.10	2.86	6.40	2.90	0.95	1.70	0.40	1.80	2.40	1.90	2.40	1.90	2.40	7.65	1.70	0.30	0.90	0.20
Total	98.92	99.33	100.29	99.95	99.79	97.98	100.11	99.37	99.29	98.57	98.51	98.45	98.45	98.45	98.45	99.82	100.10	99.81	98.90	99.31
Mg #	60.58	70.79	44.15	37.36	40.43	54.70	64.15	64.13	52.67	70.63	60.26	58.70	60.26	58.70	60.26	39.28	44.97	31.35	48.87	38.89
Cr	184	1255	229	7	73	196	169	335	231	994	292	354	292	354	292	95	229	101	103	5
Ni	82	233	123	18	40	112	45	184	124	300	120	75	48	104	104	43	48	59	130	n.d.
Co	43	33	64	33	32	41	16	49	45	56	40	44	43	48	47	43	48	47	48	n.d.
Sc	41	33	43	37	39	39	39	31	45	31	35	45	42	43	33	42	43	33	34	2
V	240					220			261							314		324	264	4
Cu	94	27	423	239	99	118	149	93	120	55	68	91	68	91	68	147	142	119	62	7
Zn	100	46	98	68	114	91	22	73	n.d.	69	74	64	n.d.	n.d.	n.d.	84	n.d.	163	99	46
Rb	n.d.	n.d.	13	7	68	n.d.	n.d.	10	26	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	14	n.d.	13	11	154
Ba	60	53	399	77	323	50	n.d.		60	90						116	68			
Sr	119	27	179	91	94	99	85	92	107	68	118	64	59	114	91	59	114	91	90	53
Ta	0.21	0.10	0.19	0.23	0.23	0.12	0.15	n.d.	0.10	0.15	n.d.	0.07	0.44	0.13	0.31	0.44	0.13	0.31	0.32	3.23
Nb	6.0	n.d.	6.0	n.d.	6.0	6.0	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	6.0	7.0	n.d.	6.0	6.0	25.0
Hf	160	0.84	1.36	1.97	2.65	1.18	1.09	0.87	1.16	0.79	0.92	0.94	3.17	1.35	2.76	2.80	1.35	2.76	2.80	4.36
Zr	68	34	69	78	104	61	46	36	47	36	38	41	117	66	108	108	117	108	108	90
Y	20	13	19	25	34	20	19	17	18	10	16	13	40	20	29	40	20	29	32	48
Th	0.49	0.19	0.29	0.85	0.61	0.21	0.14	0.14	0.14	0.22	0.12	0.48	0.84	0.21	0.57	0.84	0.21	0.57	0.63	20.55
U	0.13	0.06	0.07	0.16	0.19	0.05	0.05	0.08	0.04	0.07	0.05	0.11	0.16	0.06	0.13	0.10	0.06	0.13	0.10	3.45
La	4.83	2.19	2.61	4.33	5.93	2.70	2.16	1.38	2.49	2.08	1.80	2.86	8.74	2.75	6.42	6.28	2.75	6.42	6.28	9.79
Ce	10.77	5.42	5.99	10.18	13.34	5.65	4.64	4.57	5.69	4.21	4.82	6.70	19.91	6.80	13.81	15.47	6.80	13.81	15.47	24.63
Nd	8.40	4.02	4.75	8.31	10.84	4.42	4.93	2.68	4.56	3.45	3.62	3.62	14.50	5.50	10.72	11.30	5.50	10.72	11.30	14.28
Sm	2.34	1.07	1.68	2.60	3.36	1.71	1.54	1.12	1.61	1.05	1.23	1.15	4.25	1.87	3.40	3.26	1.87	3.40	3.26	5.55
Eu	0.91	0.48	0.77	1.05	1.49	0.77	0.67	0.48	0.74	0.58	0.60	0.53	1.40	0.82	1.43	1.42	0.82	1.43	1.42	7.00
Tb	0.56	0.28	0.43	0.64	0.89	0.49	0.40	0.28	0.44	0.28	0.33	0.27	1.05	0.55	0.82	0.75	0.55	0.82	0.75	1.23
Dy	3.42	1.43	2.91	3.96	5.37	2.89	2.53	2.04	2.75	1.63	1.80	2.10	6.02	2.62	4.73	4.68	2.62	4.73	4.68	6.81
Ho	0.71	0.36	0.57	0.87	1.08	0.65	0.57	0.43	0.62	0.27	0.46	0.39	1.14	0.69	1.06	0.91	0.69	1.06	0.91	1.34
Yb	2.18	1.09	1.82	2.68	3.58	2.02	1.84	1.49	1.95	1.15	1.50	1.55	3.85	2.13	2.92	2.80	2.13	2.92	2.80	4.28
Lu	0.36	0.18	0.30	0.43	0.55	0.34	0.28	0.24	0.31	0.17	0.23	0.25	0.64	0.35	0.48	0.41	0.35	0.48	0.41	0.67
(La/Th)cn	1.69	1.44	1.03	1.16	1.19	0.96	0.84	0.67	0.92	1.29	0.86	1.32	1.63	0.93	1.57	1.60	0.93	1.57	1.60	1.64
(Eu/Th)cn	0.96	1.02	1.22	1.04	1.09	1.18	1.09	1.27	1.20	1.36	1.25	1.22	1.07	1.10	1.10	1.10	1.07	1.10	1.10	1.06
Zr/Y	3.40	2.62	3.11	3.12	3.06	2.55	2.42	2.12	2.61	3.60	2.36	3.15	2.93	2.80	3.72	3.38	2.80	3.72	3.38	1.96
Ba/La	12.43	24.22	149.12	17.80	64.47	18.55				24.04	50.08	0.31	0.22	0.10	0.20	0.22	0.10	0.20	0.22	4.80
Th/Yb	0.23	0.18	0.16	0.24	0.17	0.10	0.08	0.09	0.07	0.19	0.08	0.05	0.13	0.05	0.11	0.06	0.11	0.06	0.11	0.75
Ta/Yb	0.09	0.09	0.11	0.08	0.07	0.06	0.08	0.05	0.05	0.13	0.05	0.05	0.13	0.05	0.11	0.06	0.11	0.06	0.11	0.75
A.I.	37.53	60.78	33.45	36.29	43.83	43.09	38.49	44.10	39.36	52.02	39.92	41.48	45.41	27.79	28.25	34.77	28.25	34.77	28.25	85.77
Note: Samples prefixed by 93- have had REE and HFSE traces done by INAA, all others by ICP. See appendix 1 for details.																				

ff felsic flow, ff felsic tuff, ff felsic tuff breccia, ff intermediate flow, ff intermediate tuff, ff mafic flow, ff ultramafic flow, per peridotite, pyx pyroxenite, mzd monzodiorite, sy syenite, may melasyenite, Al=alteration indices

Table 2.3. Major and selected trace element data for Lake of the Woods greenstone belt samples.

Assemblage Group	Upper Keweenaw				Electrum Lake				Falcon Island pluton			
	Long Bay		Sunset Gh		Crowduck Lake		White Partridge Bay					
Formation	Cliff Island											
Sample No	87-203	87-482	93-30	93-29	93-16	90-39	90-40	93-17	93-34	93-35	93-31	93-32
Rock type	m.f.	u.f.	u.f.	l.f.	l.f.	m.f.	l.f.	m.f.	l.f.	l.f.	sy	mzd
SiO ₂	61.30	45.50	43.40	64.40	78.20	55.71	56.89	55.50	69.70	72.20	55.40	52.10
TiO ₂	0.78	0.41	0.35	0.46	0.14	0.64	0.55	0.66	0.24	0.18	0.48	0.97
Al ₂ O ₃	13.90	9.71	7.95	16.80	12.90	14.37	14.85	13.50	14.50	13.50	19.60	13.70
Fe ₂ O ₃	12.40	11.90	10.90	4.59	0.73	8.82	5.59	9.22	1.85	1.24	4.23	10.70
MnO	0.18	0.16	0.19	0.06	0.03	0.16	0.09	0.15	0.05	0.04	0.10	0.19
MgO	7.05	18.60	22.10	2.44	0.59	5.58	2.59	6.04	0.84	0.61	1.23	4.22
CaO	9.48	7.56	7.00	3.83	0.54	7.00	5.90	7.55	2.83	3.90	2.08	7.36
Na ₂ O	2.63	0.64	0.69	4.39	0.61	2.36	3.42	2.46	3.85	4.23	6.73	2.97
K ₂ O	0.12	0.04	0.04	1.80	5.21	3.07	3.36	2.38	2.50	1.58	8.64	5.66
P ₂ O ₅	0.04	0.03	0.04	0.14	0.04	0.22	0.25	0.23	0.09	0.07	0.20	0.78
LOI	1.80	4.70	6.40	1.10	1.50	1.20	5.00	2.20	3.50	3.50	0.75	1.00
Total	99.58	99.85	99.06	100.01	99.89	99.13	98.59	99.98	99.95	100.45	98.84	99.55
Mg #	52.96	75.59	80.06	51.29	61.55	55.58	49.77	56.47	47.35	49.35	36.54	43.65
Cr	241	2120	2099	35	2	209	145	258	7	8	7	61
Ni	65	725	508	25	n.d.	57	35	38	1	1	3	18
Co	45	78	35	11	n.d.	29	18	22	2	2	6	21
Sc	44	31	26	8	3	23	14	25	2	2	3	14
V	262	183	163	9	n.d.	146	107	107	2	2	3	14
Cu	97	22	87	9	n.d.	63	59	52	11	1	13	73
Zn	96	66	38	78	12	128	132	64	22	44	46	102
Rb	n.d.	n.d.	n.d.	27	128	104	132	821	50	37	289	92
Ba	73	387	1240	387	1240	597	631	722	880	503	1690	2820
Sr	94	33	46	509	41	597	631	722	507	379	1520	2710
Ta	0.11	n.d.	n.d.	0.15	1.12	3.0	n.d.	0.42	0.12	0.10	0.84	0.19
Nb	5.0	n.d.	5.0	14.0	14.0	3.0	n.d.	6.0	n.d.	n.d.	11.0	n.d.
Hf	1.23	n.d.	0.44	2.58	4.73	130	149	2.96	2.60	2.20	3.20	4.53
Zr	56	43	27	108	153	130	149	128	109	91	156	180
Y	18	10	11	9	28	18	17	21	8	6	19	29
Th	0.51	0.09	0.09	1.87	14.32	14.00	11.00	7.97	5.09	4.70	7.23	6.06
U	0.14	0.00	0.00	0.44	2.32	2.07	2.07	2.07	1.39	0.67	0.87	0.64
La	3.66	0.70	0.90	13.08	58.97	30.00	29.00	29.98	22.00	22.69	59.30	111.35
Ce	7.83	2.10	2.40	28.43	112.28	66.00	65.00	57.91	37.91	37.95	116.48	228.15
Nd	5.22	2.00	1.46	12.06	43.65	31.00	28.00	27.24	13.24	13.11	47.30	107.08
Sm	1.59	0.59	0.68	1.91	6.93	6.00	4.90	5.03	1.91	1.73	7.87	19.52
Eu	0.84	0.32	0.31	0.71	1.34	1.50	1.30	1.45	0.60	0.59	2.40	5.36
Tb	0.44	0.24	0.20	0.18	0.77	0.62	0.45	0.64	0.12	0.11	0.89	1.63
Dy	2.62	1.80	1.45	1.09	3.49	3.70	2.60	3.18	0.53	0.75	3.45	6.25
Ho	0.56	0.40	0.26	0.18	0.68	0.69	0.49	0.60	0.07	0.09	0.53	1.03
Yb	1.84	1.20	1.03	0.85	2.10	1.70	1.30	1.77	0.24	0.14	0.90	1.49
Lu	0.28	0.18	0.15	0.10	0.32	0.24	0.21	0.26	0.06	0.03	0.13	0.24
(La/Yb) _{cn}	1.39	0.42	0.63	17.18	20.11	12.66	16.00	12.13	64.65	112.87	47.41	53.60
(Eu/Yb) _{cn}	0.97	1.03	1.27	0.80	0.41	0.82	0.62	0.66	0.66	0.66	0.89	0.68
Zr/N	3.11	4.30	2.45	12.00	5.48	7.22	6.76	6.00	13.63	15.17	7.16	6.21
Ba/La	0.28	0.08	0.08	20.69	21.03	8.24	8.46	27.38	40.00	24.81	26.81	26.32
Th/Yb	0.08	0.08	0.08	3.43	6.81	0.24	0.24	0.24	20.85	32.50	8.06	4.07
Ta/Yb	0.08	0.08	0.08	0.28	0.53	0.24	0.24	0.24	0.49	0.69	0.94	0.13
Al	37.10	68.43	74.22	34.03	83.45	48.03	39.38	45.69	33.33	22.63	55.83	48.89

If felsic flow, fl felsic tuff, flb felsic tuff breccia, If intermediate flow, If intermediate tuff, ml mafic flow, ul ultramafic flow, per peridotite, pyx pyroxenite, mzd monzonitoidite, sy syenite, may melasyenite, Al=alteration indices.

2.4.1 Lower Keewatin Assemblage, tholeiitic suite

The volcanic rocks of this suite are mafic and plot within the tholeiitic field on the basis of the Jensen cation plot (Fig. 2.4). Progressive iron enrichment is indicated by more prevalent Fe tholeiites in the upper stratigraphic levels and Mg tholeiites in the lower stratigraphic parts of the Lower Keewatin Assemblage (Ayer et al., 1987; Ayer and Buck, 1989). Mg numbers (mol. wt % $\text{MgO} / \text{FeO} + \text{MgO}$) range from 30 to 65.

Chondrite normalized REE patterns for the samples of this suite are flat to slightly LREE-enriched with $(\text{La}/\text{Yb})_{\text{CN}}$ ratios from 0.9-1.6 at about 6 to 20 times chondrites (Fig. 2.5a). The Fe tholeiites and the plagioclase-megaphyric tholeiites have REE contents at about 10-20 times chondrite. The Mg tholeiites have generally lower REE contents at less than 10 times chondrite. Rare komatiitic basalt flows, with Mg numbers up to about 70, have been identified in the lower part of the Cedar Island group in the Shoal Lake area (Smith, 1987). A sample of these komatiitic basalts demonstrates a similar REE pattern to the Mg tholeiites with $(\text{La}/\text{Yb})_{\text{CN}} = 1.3$ but with lower REE contents at 4-6 times chondrite. A sample of peridotite from the cumulate portion of a differentiated sill coeval with the mafic volcanics of the Bigstone Bay group also has slightly elevated LREE [$(\text{La}/\text{Yb})_{\text{CN}} = 1.4$] (Table 2.3, Fig. 2.5).

On an NMORB-normalized extended trace element plot (Fig. 2.6b), the Lower Keewatin tholeiite samples demonstrate a distinctly different profile from that of normal mid-oceanic ridge basalt (NMORB) (i.e., a straight line at unity). Instead they show the asymptotic profile of progressive enrichment of the LILE with higher degrees of incompatibility. This profile shape is most similar to that of primitive mantle (Fig. 2.6a). Compared with modern oceanic basalt

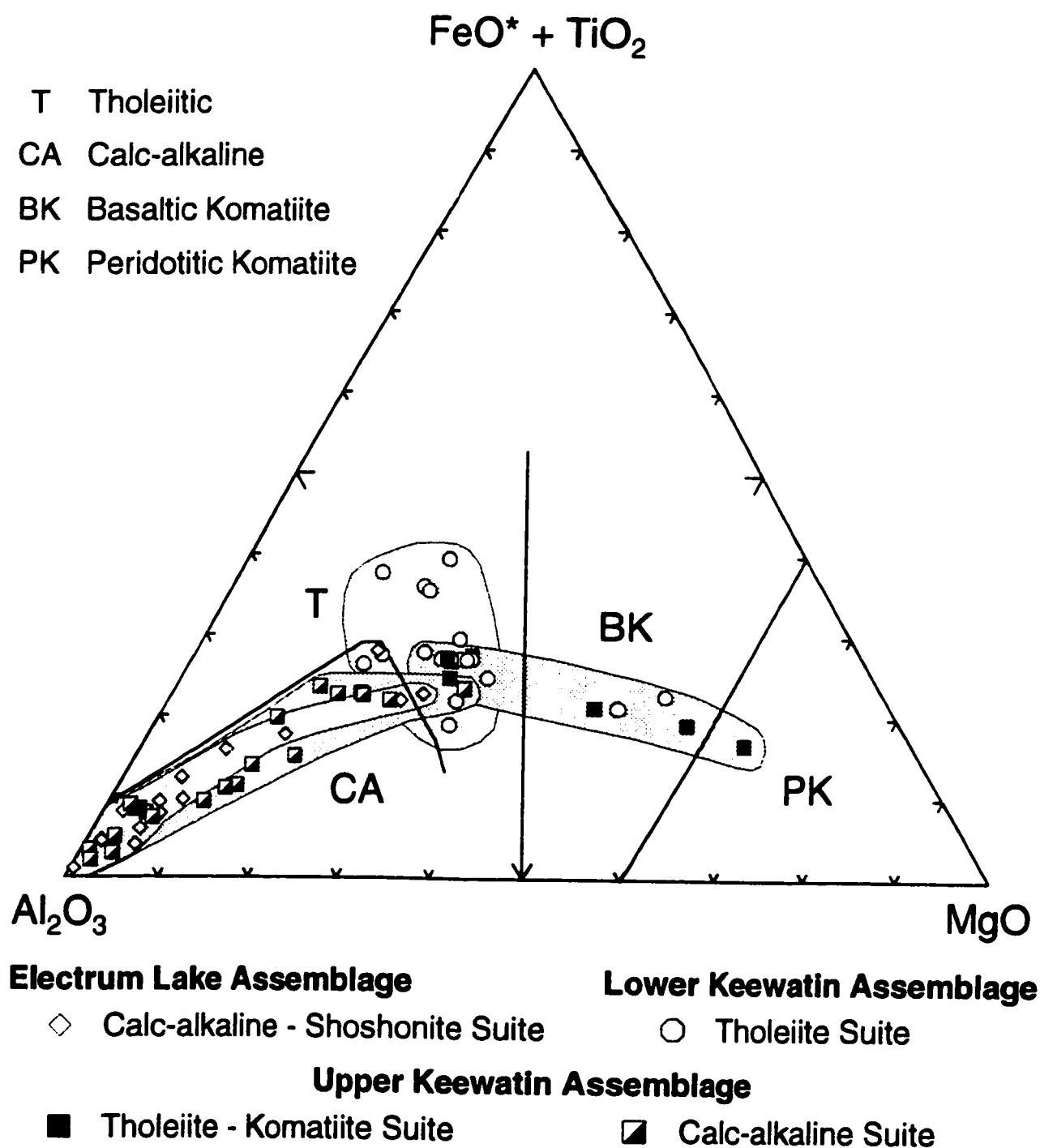


Fig. 2.4. Oxide plot (Jensen, 1976) showing the major element compositional diversity of the volcanic suites.

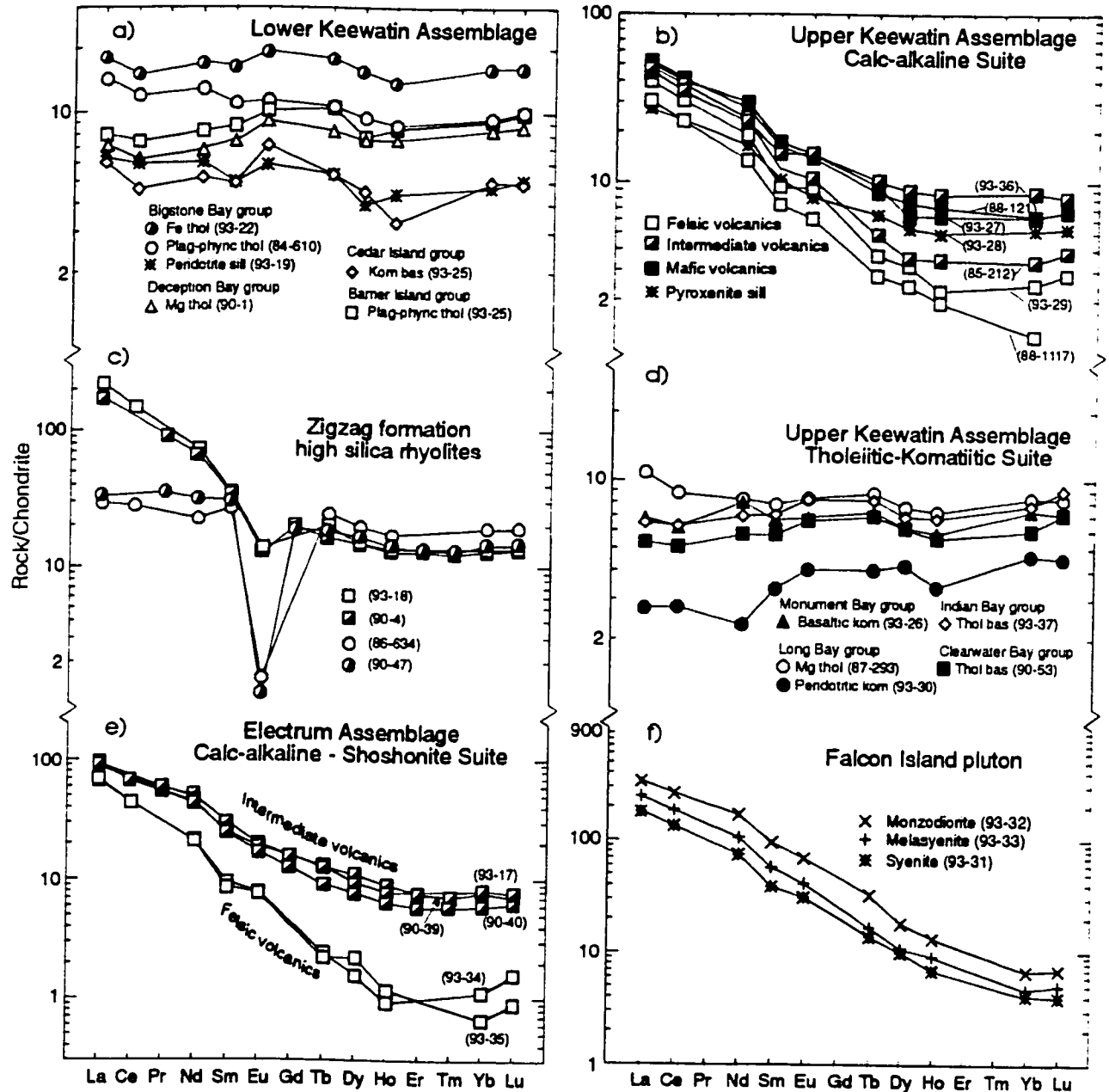


Fig. 2.5. Chondrite normalized (normalizing values from Sun and McDonough, 1989) REE abundances of representative samples from (a) the Lower Keewatin Assemblage tholeiitic suite; (b) the Upper Keewatin Assemblage calc-alkaline suite; (c) the Zigzag formation; (d) Upper Keewatin Assemblage tholeiitic to komatiitic suite; (e) the Electrum Assemblage calc-alkaline to shoshonitic suite; and (f) the Falcon Island pluton.

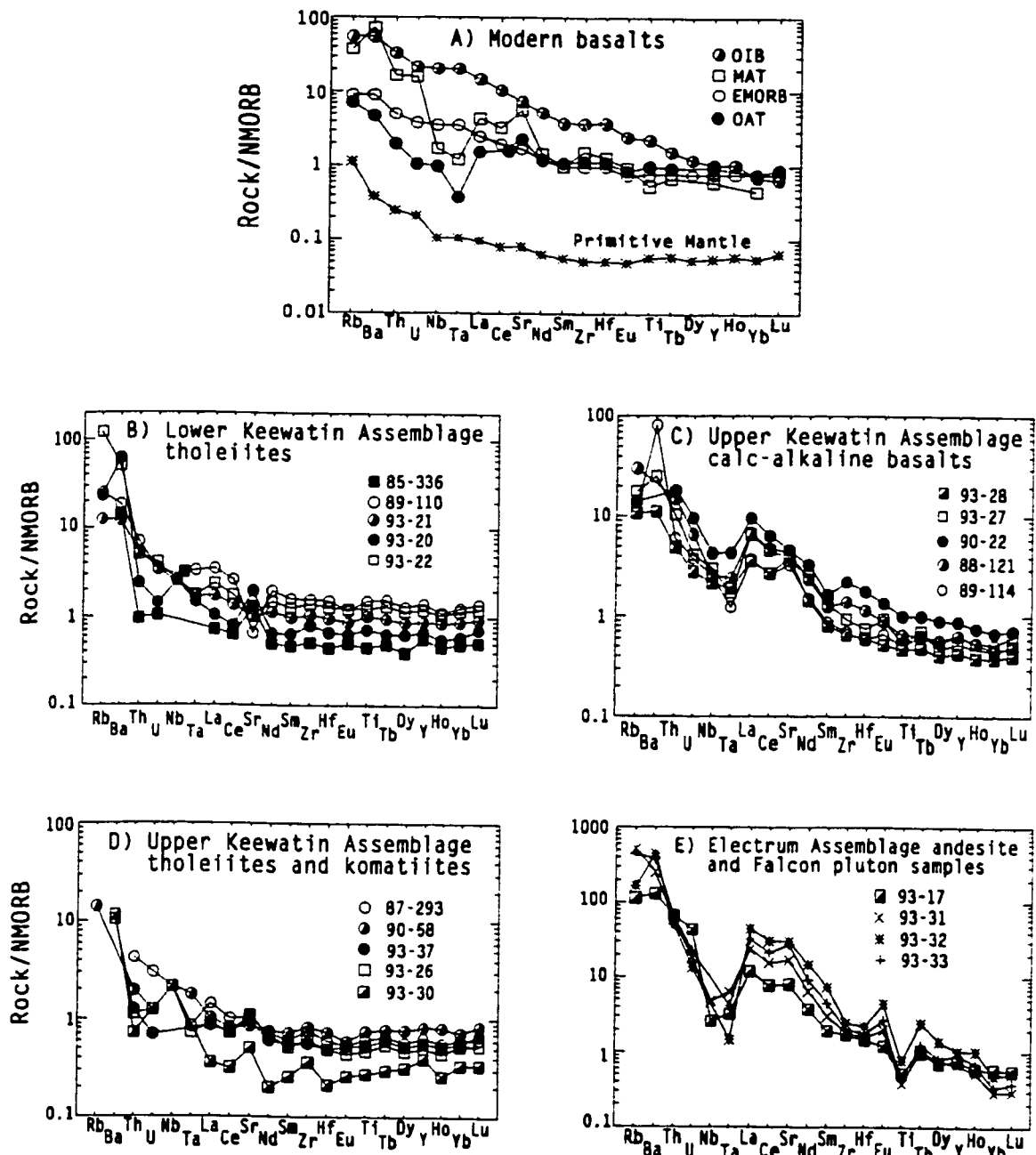


Fig. 2.6. N-MORB-normalized extended trace element plots (normalizing values from Sun and McDonough, 1989) of representative samples from (a) modern oceanic basalt environments (OIB and E-MORB averages of Sun and McDonough, 1989; oceanic arc tholeiite (OAT) representative sample, Ewart and Hawkesworth 1987; mature arc tholeiite (MAT), representative sample, Defant et al. 1991); (b) Lower Keewatin Assemblage tholeiitic suite; (c) Upper Keewatin Assemblage calc-alkaline suite; (d) Upper Keewatin Assemblage tholeiitic to komatiitic suite; and (e) Electrum Assemblage and Falcon Island pluton samples.

averages (also normalized to NMORB) (Fig. 2.6a), the Lower Keewatin tholeiites lie in a generally similar range to those of evolved mid-oceanic ridge basalts (EMORB) and oceanic arc tholeiites (OAT), but have higher abundances of Rb and Ba than EMORB and lack the depletion in Nb and Ta of OAT. Relative enrichment of the LILE is also indicated by Ba/La ratios which range from 12-70 (Table 2.3). These ratios are similar to the range of OAT (i.e., > 10), and are considerably higher than the typical values for modern back-arc basin basalts (BABB), NMORB or OIB (i.e., Ba/La less than 10) (Gill et al., 1984; Hawkins and Melchior, 1985).

On a plot of Th/Yb versus Ta/Yb ratios (Fig. 2.7), the Lower Keewatin tholeiites overlap the fields of arc tholeiites and extensional setting basalts (Pearce, 1983). Hawkesworth et al. (1993) distinguish between oceanic arc and mature island arc tholeiites on the basis of Yb versus Ce abundances. Figure 2.8 illustrates the fields of Yb versus Ce values for selected modern arc environments rocks. When compared with these modern environments, the Lower Keewatin tholeiitic suite samples display the relatively large variation in Yb for small changes in Ce similar to modern OAT and overlap the field of the Tonga-Kermadec OAT samples.

2.4.2 Upper Keewatin Assemblage, calc-alkaline suite

Volcanic rocks ranging in composition from basalt through to rhyolite occur in many of the units of the Upper Keewatin Assemblage. They lie predominantly within the calc-alkaline field on Figure 2.4. The mafic to felsic samples typically exhibit high Zr/Y ratios and chondrite normalized REE patterns with enriched LREE and depleted HREE for $(La/Yb)_{CN}$ ratios of about 4-35 (Table 2.3 and Fig. 2.5b). Mafic and intermediate samples have higher total REE abundances

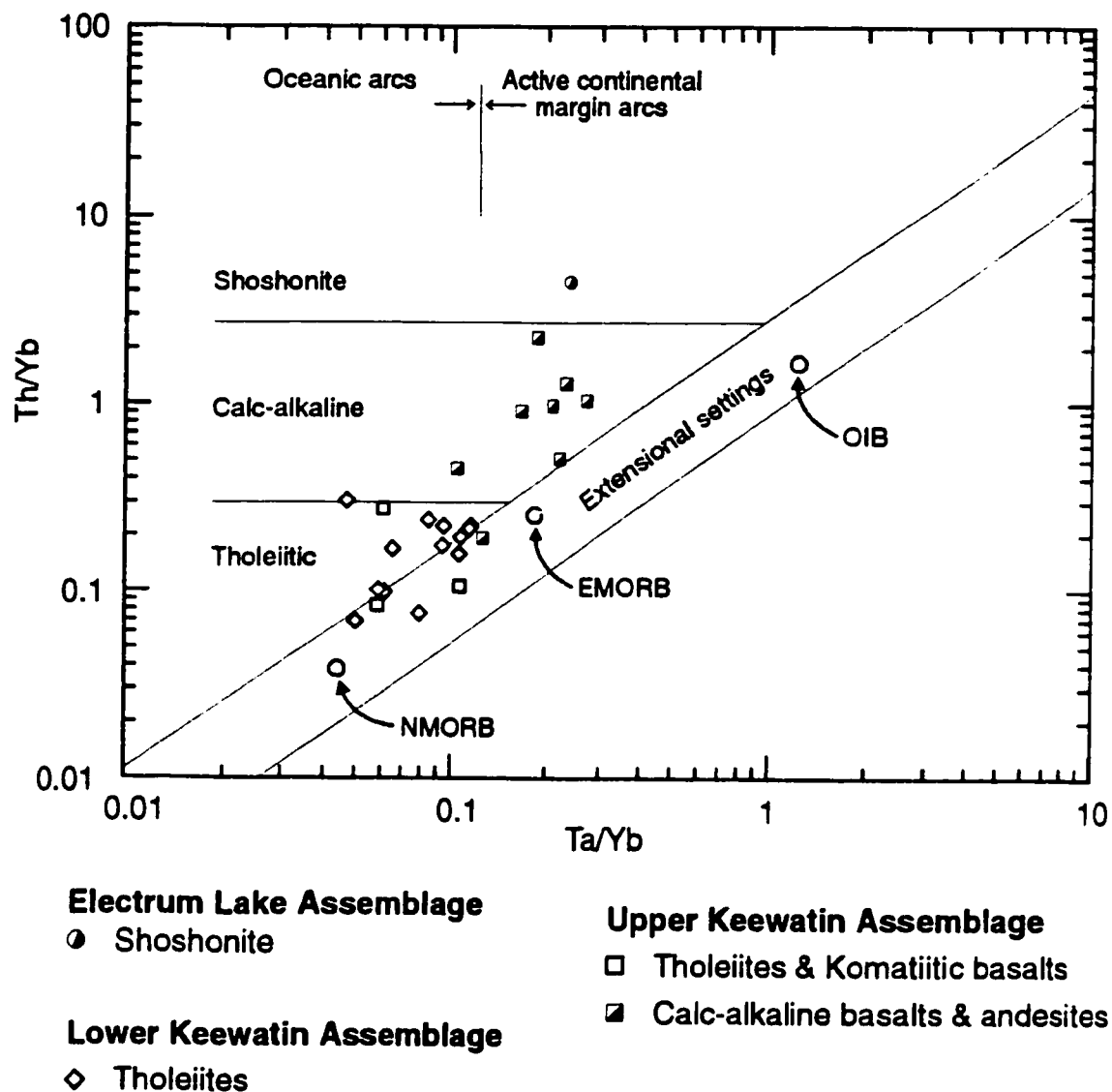


Fig. 2.7. Th/Yb versus Ta/Yb diagram with the fields of modern oceanic arcs, active continental margin arcs, and intraplate oceanic basalts after Pearce (1983).

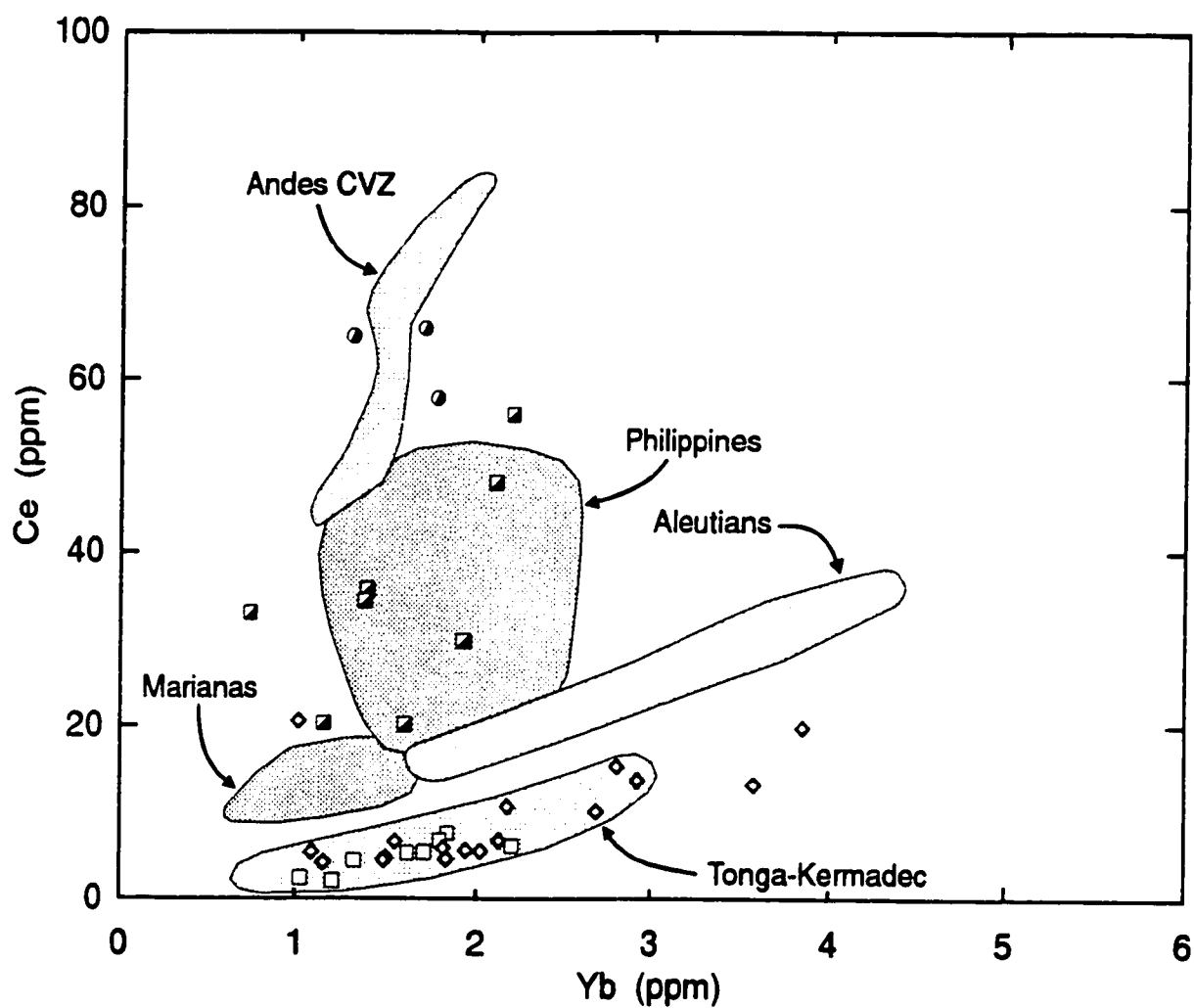


Fig. 2.8. Ce versus Yb diagram with the fields of oceanic arcs (Tonga-Kermadec arcs, Marianas arc and Aleutians arc), island arcs (Philippines), and continental arc (Andes - Central Volcanic Zone) after Hawkesworth et al. (1993). Symbols as indicated on Figure 2.7.

and lower $(La/Yb)_{CN}$ ratios than those of the felsic calc-alkaline samples. A pyroxenite sample from the cumulate portion of a differentiated synvolcanic sill in the calc-alkaline suite also displays fractionated REE patterns [$(La/Yb)_{CN} = 5.6$] indicating that the parental liquids from which the cumulate crystallized were significantly fractionated (Fig. 2.5b, Table 2.3). This is in distinct contrast to the relatively flat chondrite-normalized REE patterns of ultramafic cumulates within synvolcanic sills in the Lower Keewatin Assemblage (Fig. 2.5a).

The Upper Keewatin calc-alkaline suite samples, with their elevated incompatible elements (Rb, Ba and Sr), elevated LREE and depleted HREE and HFSE (Th, U, Nb, Ta) have elemental abundances most similar to those of calc-alkaline volcanic rocks in mature island arcs such as the Philippines (Figs. 2.6a and 2.6c) (Defant et al., 1991). Samples of this suite plot within the calc-alkaline, continental margin arc field on a plot of Th/Yb versus Ta/Yb (Fig 2.7) and the mafic samples show large variations in Ce for relatively small changes in Yb similar to mature island arc suites (Fig. 2.8).

Calc-alkaline dacites and rhyodacites are the typical felsic end-members of the suite. These felsic volcanic rocks are similar to the high-Al trondhjemite-tonalite-dacite suite (TTD) classification of Drummond and Defant (1990) or the FI calc-alkaline rhyolite classification of Leshner et al. (1986). The calc-alkaline felsic volcanics are characterized by high Al_2O_3 (greater than 13%), high Sr (greater than 250 ppm), low Rb (less than 100 ppm), enriched LREE and depleted HREE, without Eu anomalies (Fig. 2.5b). The HREE-depleted and steep negative slopes of the REE patterns indicate retention of garnet in the source and thus derivation from a deep (greater than 75 km) mantle source and is similar to that of the Electrum Assemblage suite (discussed below).

A geochemically distinctive rhyolite unit, identified as the Zigzag formation, occurs within the Clearwater group in the northwestern part of the Upper Keewatin Assemblage. The unit extends over a strike length of about 40 km and is up to 1 km thick. The formation consists of pyroclastic rocks and minor flows of rhyolitic composition with higher SiO_2 (75-80%), lower Al_2O_3 (< 13%), lower Sr (< 100 ppm) and higher Rb (> 100 ppm) abundances than the typical calc alkaline felsic volcanic rocks of the Upper Keewatin Assemblage. The Zigzag formation rhyolites are geochemically divisible into two groups on the basis of their REE patterns: (1) about 75% of the analyzed samples have elevated and steeply sloped LREE, pronounced Eu depletion anomalies and flat but relatively elevated HREE [$(\text{La/Yb})_{\text{CN}} = 14-18$ and $\text{Eu}^* = .3$]; and (2) about 25% of the samples have flat REE with very pronounced Eu depletion anomalies [$(\text{La/Yb})_{\text{CN}} = 1.5-2.5$ and $\text{Eu}^* = .05$](Fig. 2.5c).

The unusual trace element patterns of both groups of the Zigzag rhyolites are considered to be primarily the result of relatively high degrees of partial melting of mafic crust possibly accompanied by feldspar fractionation (Leshner et al. 1987; Barrie et al. 1993). However, the distinctive REE patterns of the latter group of Zigzag formation rhyolites could be either the result of differences in the magmatic processes, or hydrothermal alteration which resulted in depletion of LREE and Eu of some of the samples (Schandl and Gorton, 1991). The alteration hypothesis may be favoured as many of the samples of the latter group have elevated alteration indices ($\text{A.I.} > 80$; Table 2.3).

2.4.3 Upper Keewatin Assemblage, tholeiitic-komatiitic suite

Although similar in appearance to the mafic volcanic rocks of the Lower Keewatin Assemblage, the volcanic rocks of this suite are at least 15 My younger and occur as conformable intercalations with the calc-alkaline volcanic units of the Upper Keewatin Assemblage.

Geochemically the suite is more primitive than the Lower Keewatin tholeiites consisting mostly of magnesium tholeiites and about 5% komatiitic basalts and 5% ultramafic komatiites (Fig. 2.4). Mg numbers range from 45-80 (Table 2.3). Komatiite samples (up to 22% MgO) commonly occur in aphyric pillowed flows demonstrating that the primitive composition is that of the liquid and not that of the cumulate portion of a differentiated flow.

The Upper Keewatin tholeiitic to komatiitic suite rocks have lower concentrations of Fe, Ti, P, V, Y and Zr and higher concentrations of Mg, Cr and Ni than the Lower Keewatin tholeiites. Chondrite normalized REE of the suite are generally less than 10 times chondrites. The tholeiites are flat to slightly LREE depleted with $(La/Yb)_{CN}$ ratios from 0.8-1.0. Komatiite samples have chondrite-normalized REE patterns with slopes similar to the tholeiites but with lower REE values and more depleted LREE [$(La/Yb)_{CN}$ ratios from 0.4-0.6] (Fig. 2.5d). These REE patterns differ from the higher and slightly LREE enriched values of Lower Keewatin tholeiites and rare komatiitic basalts which are flat to slightly LREE enriched (Fig. 2.5a).

The Upper Keewatin tholeiitic and komatiitic suite samples demonstrate patterns of progressive enrichment of the LILE with higher degrees of incompatibility on an NMORB-normalized extended trace element plot (Fig. 2.6d). They are thus similar in profile, but with lower concentrations, to the Lower Keewatin tholeiite samples (Fig. 2.6b). They are also similar

in the profile shape to primitive mantle, but with more elevated trace element concentrations (Fig. 2.6a). Compared with modern basalts, they are most similar to the patterns of EMORB and OAT but lack the depletion in HFSE of the latter (Fig. 2.6a). They overlap the fields of arc tholeiites and extensional basalts on Figure 2.7 and the field of Tonga-Kermadecs OAT on Figure 2.8.

The ultramafic samples of the suite have similar REE values, LREE depletion and chondritic $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratios (ie. about 20) to typical alumina-undepleted Archean komatiites of the Abitibi Subprovince (Barrie et al., 1993; Xie et al., 1993). But they differ from many spinifex textured Archean komatiites by ranging to considerably lower maximum MgO values for chill compositions (22% MgO versus 35% MgO). In these features, the Upper Keewatin Assemblage komatiites are geochemically similar to modern arc picrites from intra-oceanic arcs (Ramsey et al., 1984, Nye and Reid, 1986, Eggins 1993). A significance difference however, is that the arc picrites have considerably more elevated LREE and more depleted Nb, Ta and Ti.

2.4.4 Electrum Assemblage, shoshonitic to calc-alkaline suite

Intermediate to felsic volcanic rocks constitute about 5-10% of the predominantly sedimentary Electrum Assemblage. Although the samples of this suite plot in the calc-alkaline field on Figure 2.4, the intermediate composition samples also plot in the shoshonitic field on a K_2O versus SiO_2 diagram (Fig. 4.3a). Intermediate volcanic rocks of this suite have more elevated Na_2O , Rb, Sr, Ba and LREE than the calc-alkaline intermediate volcanic rocks of the Upper Keewatin Assemblage (Table 2.3). The felsic volcanic rocks of the Electrum Assemblage suite are transitional into the high potassium calc-alkaline field and plot in a similar range to the FI type

felsic volcanic rocks of the Upper Keewatin Assemblage. On chondrite-normalized REE plots, the samples of this suite display highly fractionated patterns with enriched LREE (Fig. 2.5e) and $(La/Yb)_{CN}$ ratios from 12-112 with significantly lower LREE and MREE and markedly depleted HREE values for the felsic volcanic end-members in comparison to the intermediate compositions (Fig 2.5e).

A N-MORB-normalized extended trace element plot of Electrum Assemblage shoshonite shows generally similar pattern to those of the Upper Keewatin calc-alkaline suite, but with more incompatible element enrichment and more depletion in Nb, Ta and Ti. The elevated Th/Yb versus Ta/Yb ratios of this sample plot within the continental arc shoshonitic field on Figure 2.7. Modern shoshonites are near-silica saturated, K-rich volcanic rocks that are transitional between calc-alkaline and alkaline suites (Morrison, 1980). They are interpreted as the late stage products of arc magmatism and are found in island arcs and continental arcs (Morrison, 1980; Hickey et al., 1986).

2.4.5 Falcon Island pluton

The Falcon Island pluton, located in the southern part of the LWGB (Fig. 2.2), consists predominantly of melasyenite and syenite, with minor monzodiorite and clinopyroxenite. All phases have highly elevated K_2O , Na_2O , Rb, Ba, Sr, and LREE (Table 2.3). Samples of the pluton are more alkaline and potassic (higher K_2O , Rb, Sr, Ba), and have slightly more more elevated LREE, and more depleted Nb, Ta and Ti than the Electrum Assemblage shoshonite sample (Figs. 2.5 & 2.6).

The similar alkalic chemistry and potentially similar age of the Falcon Island pluton (2695 Ma) and the Electrum Assemblage (< 2699 Ma), suggests that the pluton is comagmatic with the volcanic rocks of the Electrum Assemblage.

2.5 Discussion

2.5.1 Petrogenesis of the Tholeiitic and Komatiitic Suites

Figure 2.9 illustrates $(\text{Nb/La})_{\text{CN}}$ versus $(\text{Ce/Yb})_{\text{CN}}$ ratios for the LWGB samples and the averages for a number of modern tectonic environments. The ratios are a measure of the degree of HFSE/LREE and LREE/HREE fractionation relative to chondritic values. The Lower Keewatin tholeiitic samples are close to chondritic in Nb/Yb and Ce/Yb ratios (i.e., 1.0) and are mostly constrained to a field between the average values of NMORB, EMORB and OAT suggesting that the Lower Keewatin tholeiites were derived from undepleted primitive mantle not significantly modified by LREE enrichment or HFSE depletion. Rocks of the Upper Keewatin tholeiitic and komatiitic suite have similarly depleted Ce/Yb ratios to average NMORB, but range to more enriched Nb/La ratios than NMORB and EMORB. Derivation of these suites from relatively unmodified primitive mantle is also suggested by Figure 2.6, where the NMORB-normalized samples of the Lower and Upper Keewatin tholeiites are very similar in their profile shapes to that of primitive mantle.

Figure 2.9 also shows that many of the Lower Keewatin tholeiite samples plot in a field similar to the modelled partial melting curve of spinel-bearing primitive mantle for geologically

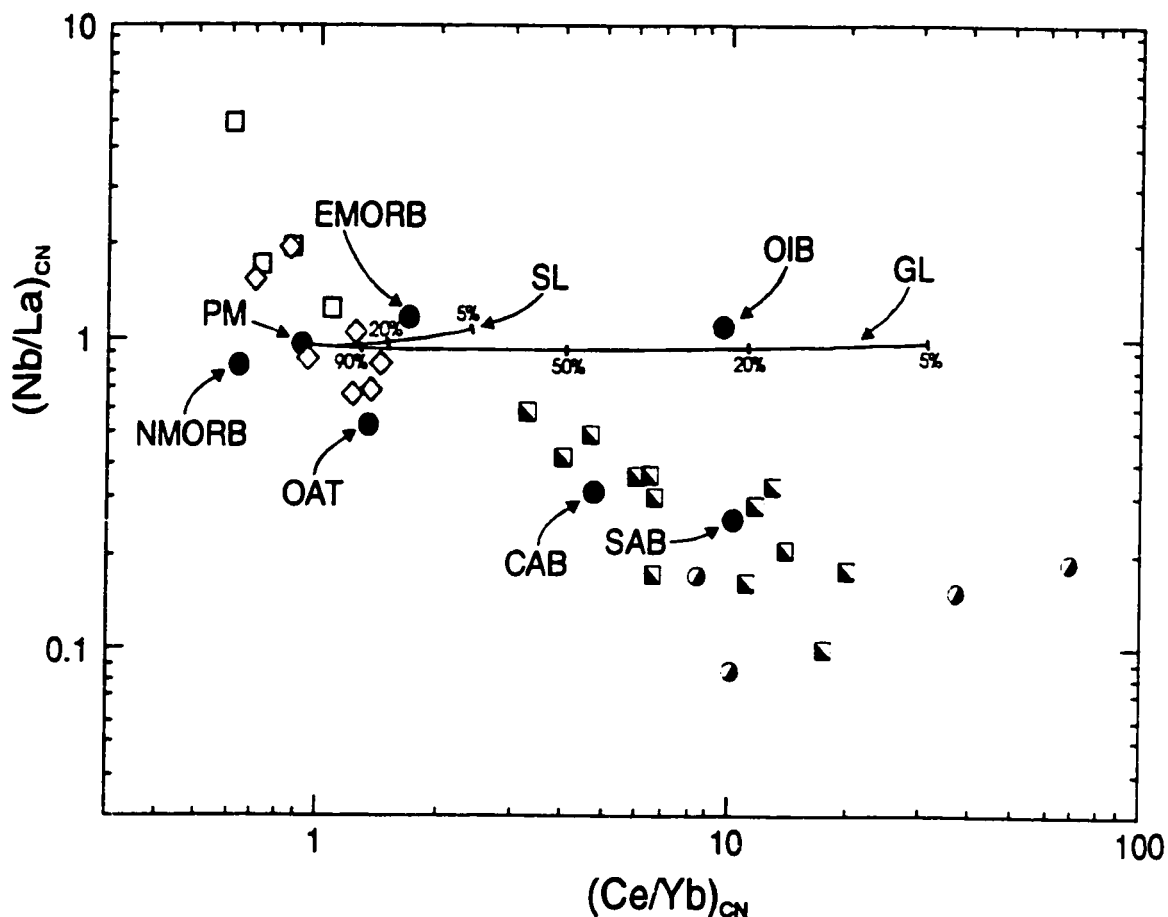


Fig. 2.9. Chondrite normalized Nb/La versus Ce/Yb with representative samples of modern environments as in Figure 2.6 (OAT - oceanic arc tholeiite, CAB - calc-alkaline arc basalt, SAB - shoshonitic arc basalt, Southern Andes - Hickey et al. 1986). Trace element models were calculated using the formula and mineral/melt partition coefficients for basaltic liquids of Rollinson (1993). Equilibrium (or batch) partial melting curves were calculated using primitive mantle as starting material (values of Sun and McDonough, 1989) for GL (garnet lherzolite with 55% olivine, 25% orthopyroxene, 15% clinopyroxene and 5% garnet) and SL (spinel lherzolite with 55% olivine, 25% orthopyroxene, 15% clinopyroxene and 5% spinel). Symbols as indicated on Figure 2.7.

reasonable amounts of melting (i.e., about 20%). However, they are only similar to those of the garnet-bearing primitive mantle model curve values for unreasonably high amounts of partial melting (about 90%). In contrast, the position of average OIB is at an equivalent value to about 20% partial melting of garnet-bearing primitive mantle. This suggests that the mantle source of the Lower Keewatin tholeiites was at relatively shallow depths, above the spinel/garnet phase transition (i.e. above 70 km depth).

Although not perfect analogs, both the Lower Keewatin and the Upper Keewatin tholeiites and komatiites have many similar trace element abundances to modern BABB and OAT suggesting they may have been derived from an Archean mantle equivalent to modern convergent margin geodynamic settings by partial melting of an undepleted, relatively shallow mantle source. The main differences between modern arc tholeiites and BABB are lower HFSE concentrations relative to REE and LILE abundance in the arc basalts (Saunders and Tarney 1984). BABB commonly have a close spatial and temporal association with arc basalts and they are commonly intermingled in their boundary zones. In addition, the similarity of isotopic signatures and some of the trace elements indicate that some intra-oceanic arc basalts may have extracted magma from the associated back arc system through convection in the mantle wedge overlying the subducting slab (Gill, 1987; Price et al., 1990; Hawkesworth et al., 1993, Tatsumi and Eggins, 1995).

Interlayering of tholeiites and komatiites within the Upper Keewatin Assemblage is similar to that of many Archean greenstone belts. Campbell et al. (1989) postulated that this type of rock association is the result of plume-dominated melting in a warmer Archean mantle with the komatiites derived from the hotter axial regions of the plume, and the basalts from the cooler head of the plume. Two hypotheses have been advanced to explain komatiite petrogeneses. In the first,

komatiite liquids are formed by high degrees of partial melting at low pressures (Nisbit and Walker, 1982). In the second, they form at relatively low degrees of partial melting at high pressures (Herzberg, 1992). On the basis of differences in HFSE/REE fractionations, Xie et al. (1993) have proposed that Abitibi Subprovince komatiites were generated by mantle plumes at differing depths ranging from less than 300 km to greater than 700 km. They speculate that the least magnesian group of Abitibi komatiites (MgO from 15-24%) were derived from the shallowest depths. These komatiite are quite similar in MgO, REE and HFSE to the Upper Keewatin suite komatiites suggesting that they also may have been derived from a relatively shallow plume in undepleted mantle.

Allègre (1982) hypothesised an alternative method of producing komatiites by reduction of the solidus temperature in the presence of relatively high quantities of dissolved water. He states that typical mantle is dry with a maximum of up to about 0.1% water. Gallagher and Hawkesworth (1992) have indicated that mantle could accommodate up to 0.4% water in the form of amphibole produced by the hydration of clinopyroxene. They calculate that this amount of hydration would lower the mantle solidus temperature by over 400°C at 3 GPa (about 100 km depth). Thus wet mantle could accommodate a considerably larger degree of partial melting under the same temperature and pressure condition than dry mantle with the same geothermal gradient.

Subduction-related production of komatiites is not a new concept and has already been proposed by Allègre (1982), who noted the widespread association of Archean komatiites and tholeiites with island arc-like calc-alkaline suites. Subduction of an oceanic plate with a crust of sea floor basalts and sediments provides a mechanism for hydration of the overlying mantle wedge. The Archean geothermal gradient was probably steeper than today because of greater heat

production from U, Th and K. The lack of komatiites in modern arcs may indicate that the Phanerozoic geothermal gradients were too low to support the required high degree of partial melting. It may be that the combination of a number of mantle conditions are required in subduction zones to produce komatiites. These could include subduction hydration, a high geothermal gradient (Archean upper mantle conditions) and second stage melting of refractory portions of the mantle wedge which have already produced basaltic melts.

The intercalation of tholeiites and komatiites with coeval calc-alkaline rocks in the Upper Keewatin Assemblage, appears to be unique to the Precambrian and has not been documented in modern arcs. Structural interleaving of the tholeiitic-komatiitic and calc-alkaline suites in the Upper Keewatin Assemblage is dismissed as contacts between the suites are not tectonic. This leads to two possible interpretations: (1) derivation of the two chemically distinctive suites from different geodynamic settings in a common depositional basin or (2) derivation of the two suites from different magmatic sources in a common geodynamic setting.

McKenzie and O'Nions (1991) indicate that magmas associated with island arcs are derived from two sources, a large melt fraction at shallow depths by adiabatic decompression melting and a smaller melt fraction (1%) at greater depths from the advection of water. Mature island arcs, such as the Japan arc, have intimately intercalated tholeiitic basalts and calc-alkaline fractionation series rocks (Yanagi et al., 1994). Thus the intercalation of tholeiitic and calc-alkaline suites in the LWGB could be analogous to modern destructive plate margins. The more primitive composition of the LWGB tholeiites and komatiites could be the result of higher percentages of partial melting at shallow mantle depths under higher Archean geothermal gradients than those of the Phanerozoic. The range to higher Nb/La and lower Ce/Yb ratios of the

Upper Keewatin Assemblage tholeiites and komatiites (Fig. 2.9) may indicate that they formed by continued partial melting of the refractory mantle source from which partial melts of the Lower Keewatin Assemblage suite had already been extracted.

2.5.2 Petrogenesis of the Calc-alkaline and Shoshonitic Suites

Figure 2.9 also demonstrates the distinctive subduction signature of the calc-alkaline samples of the Upper Keewatin suite. They display progressive Nb depletion (i.e., lower Nb/La ratios) for increasing Ce/Yb fractionation, plotting in a similar range to average calc-alkaline island arc basalt (CAB). Small degrees of partial melting of a mantle wedge source at a sufficient depth to have retained garnet in the residue could result in HREE depletion with a resultant increase in Ce/Yb. However, the trend of HFSE depletion in the samples as indicated by decreasing Nb/La for increasing Ce/Yb can not be explained by garnet retention only and requires a process involving mantle wedge metasomatism by subduction-zone fluids (Hawkesworth et al., 1993; Kepper 1996). The higher degree of Ce/Yb fractionation evident in the position of the the Electrum Assemblage shoshonite samples, plotting near the position of shoshonitic continental arc basalts (SAB on Fig. 2.9) indicates greater HREE retention, but the relatively constant Nb/La ratios evident when compared with the Upper Keewatin calc-alkaline basalts suggests comparable amounts of metasomatism.

Documented occurrences of shoshonitic volcanic rocks are increasingly common throughout the Superior Subprovince as the volcanic components of the Neoarchean successor basins are studied in greater detail. Alkalic volcanic rocks (some of shoshonitic affinity) have been

reported in the central part of the western Wabigoon Subprovince (Teal, 1979) and the Lake Shebandowan area of the Wawa Subprovince (Shegelski, 1980). Shoshonites of striking geochemical similarity to those of the Electrum Assemblage also occur in the Oxford Lake area of northeastern Manitoba (Brooks et al., 1981) and the Chibougamau area of northwestern Quebec (Dostal and Mueller, 1992). Volcanic rocks with shoshonitic affinities but ranging to more alkaline compositions occur in the Timiskiming group in northeastern Ontario (Ujike, 1985; Ben Othman et al., 1990).

Shoshonites from modern arc environments are considered to be generated primarily in the mantle wedge overlying the Benioff zone (Gill, 1981). The LREE/HREE ratios are most likely good approximations of the parental melts as these ratios are not significantly modified by low pressure crystal fractionation. Dostal and Mueller (1992) suggest two petrogenetic factors are critical in the derivation of the highly fractionated patterns in the shoshonites at Chibougamau. The first is the percentage of partial melting. Highly fractionated REE patterns are the result of very low degrees of partial melting (about 1%). The second is retention of garnet in the mantle residue, if garnet is a retained mineral phase, small degrees of partial melting will result in enriched LREE but also highly depleted HREE patterns (as seen in the felsic end-members of the Electrum Assemblage and the Upper Keewatin calc-alkaline Assemblage) (Fig. 2.5). Spinel peridotite transforms into garnet peridotite at pressures of greater 22 kbar (at a depth of about 70 km). Thus shoshonites were probably derived from a heterogeneous mantle wedge at depths where this phase transition occurs. This suggests that the Upper Keewatin and Electrum calc-alkaline and shoshonite assemblages were probably derived from an arc with a significantly thickened crust, a situation analogous to the modern mature island arcs and continental arcs.

Morrison (1980) proposed that arc shoshonites are the result of steepening of the subducting slab with eventual "failure" and flipping of the subduction zone accompanied by block faulting and uplift within the overlying island arc. The predominantly coarse clastic sedimentary rocks (alluvial fan-fluvial facies?) with minor associated shoshonitic to calc-alkaline volcanic rocks occurring as localized deposits unconformably overlying the Upper Keewatin Assemblage suggests this type of late stage faulting may have played a significant role in the deposition of the Electrum Assemblage.

2.6 Conclusions:

Hawkins et al. (1984) characterize a complete arc-trench magma sequence as progressing from boninites to arc tholeiites to calc-alkaline rocks, to shoshonites (parts of which may be absent). There is a degree of similarity in the Neoproterozoic LWGB, with the evolution of the Lower Keewatin tholeiites, followed by Upper Keewatin Assemblage calc-alkaline suite and finally the Electrum Assemblage shoshonites.

Although not perfect analogs, the Lower Keewatin Assemblage tholeiites bear some geochemical similarities to modern arc tholeiites and BABB suggesting they may have been derived by destructive margin magmatism involving about 20% partial melting of undepleted primitive mantle at the relatively shallow depth of less than 70 km. They lie at the base of the stratigraphic section in the LWGB and have been dated at 2738 Ma.

The Upper Keewatin calc-alkaline suite rocks are about 15 Ma years younger and disconformably overlie the Lower Keewatin Assemblage. They are geochemically and

petrologically similar to the calc-alkaline volcanic rocks forming in modern mature island arc settings. Intercalated tholeiitic and komatiitic suite rocks are geochemically similar but somewhat more primitive than the older Lower Keewatin tholeiites and may be the result of second stage melting of the refractory mantle source from which the Lower Keewatin tholeiites were derived, perhaps the combined product of a higher Archean geothermal gradient and a subduction hydrated mantle.

The Electum Assemblage occurs as localized successor basins unconformably overlying the Upper Keewatin Assemblage in the northern part of the LWGB. Detrital zircon ages indicate deposition after 2699 Ma. The volcanic rocks are of shoshonitic and calc-alkaline affinity and are most similar to those of modern shoshonitic suites forming in the late stages of development of mature island arcs and continental margins. The syenitic Falcon Island pluton, located in the southern part of the LWGB, has similar trace element geochemical patterns with the shoshonites, has been dated at 2695 Ma and thus appears to be coeval with Electum Assemblage volcanism.

Chapter 3

The Mafic Minerals of the Falcon Island Ultrapotassic Pluton: Progressive Reduction During Fractionation.

3.1 Introduction

Precambrian alkaline magmatism is uncommon (Corriveau & Gorton 1993, Blichert-Toft et al. 1996), and well documented accounts of mineral chemistry of Archean ultrapotassic plutons are mostly restricted to the Abitibi Subprovince of the Superior Province (Ben Othman et al. 1990, Laflèche et al. 1991, Rowins et al. 1993). In the Phanerozoic, ultrapotassic, potassic and shoshonitic suites associated with calc-alkaline magmatism are restricted to arc environments (Foley et al. 1987, Wheller et al. 1987). In the Abitibi Subprovince, late-tectonic, alkaline volcanic suites of the Timiskaming Group and coeval alkaline plutons have important economic implications because of their close spatial and temporal association with gold mineralization (Sinclair 1982, McNeil & Kerrich 1986, Fyon et al. 1989, Rock & Groves 1988, Wyman & Kerrich 1989, Robert 1997). The oxidation state of the Abitibi intrusive suites is variable and is also considered to play an important role in mineralization (Cameron & Carrigan 1987, Cameron & Hattori 1987, Rowins et al. 1991, 1993, Levesque 1994).

The Falcon Island composite pluton is located in the southern part of the Lake of the Woods greenstone belt, within the Wabigoon Subprovince, of the Superior Province in

northwestern Ontario (Fig. 3.1). The pluton was chosen for detailed study because of its ultrapotassic composition, a relatively rare occurrence in the Archean, and its similarity in age and geochemistry to the shoshonitic metavolcanic units in the Electrum Lake Assemblage, an unconformable, "Timiskaming-like" sedimentary basin in the northern part of the belt. Both the pluton and the Electrum Lake Assemblage are considered to be derived from a mantle wedge formed in the late stages of a mature island arc or a continental arc setting of Neoarchean age (Ayer & Davis 1997).

Comparison of the mineral chemistry of the Falcon Island pluton with that of other syenitic complexes, including oxidized complexes in the Kirkland Lake gold camp, provides insight into the geodynamic and mineralization processes associated with gold mineralization in the Neoarchean.

3.2 Regional context

The Falcon Island complex intrudes tholeiites of the Lower Keewatin Assemblage (2738 ± 2 Ma) and calc-alkaline metavolcanic rocks, tholeiitic to komatiitic metavolcanic rocks and turbiditic metasedimentary rocks of the Upper Keewatin Assemblage, (2723 to 2712 Ma) (Chapter 2) (Fig. 3.1). Plutonic rocks of the Lake of the Woods greenstone belt have been subdivided into three groups: (1) pre-tectonic intrusions, including synvolcanic stocks and early tonalite and granodiorite units within the batholiths (2727 to 2717 Ma); (2) syn-tectonic stocks and the late granitic phases within the batholiths (2711 to 2708 Ma); and (3) late tectonic granitic

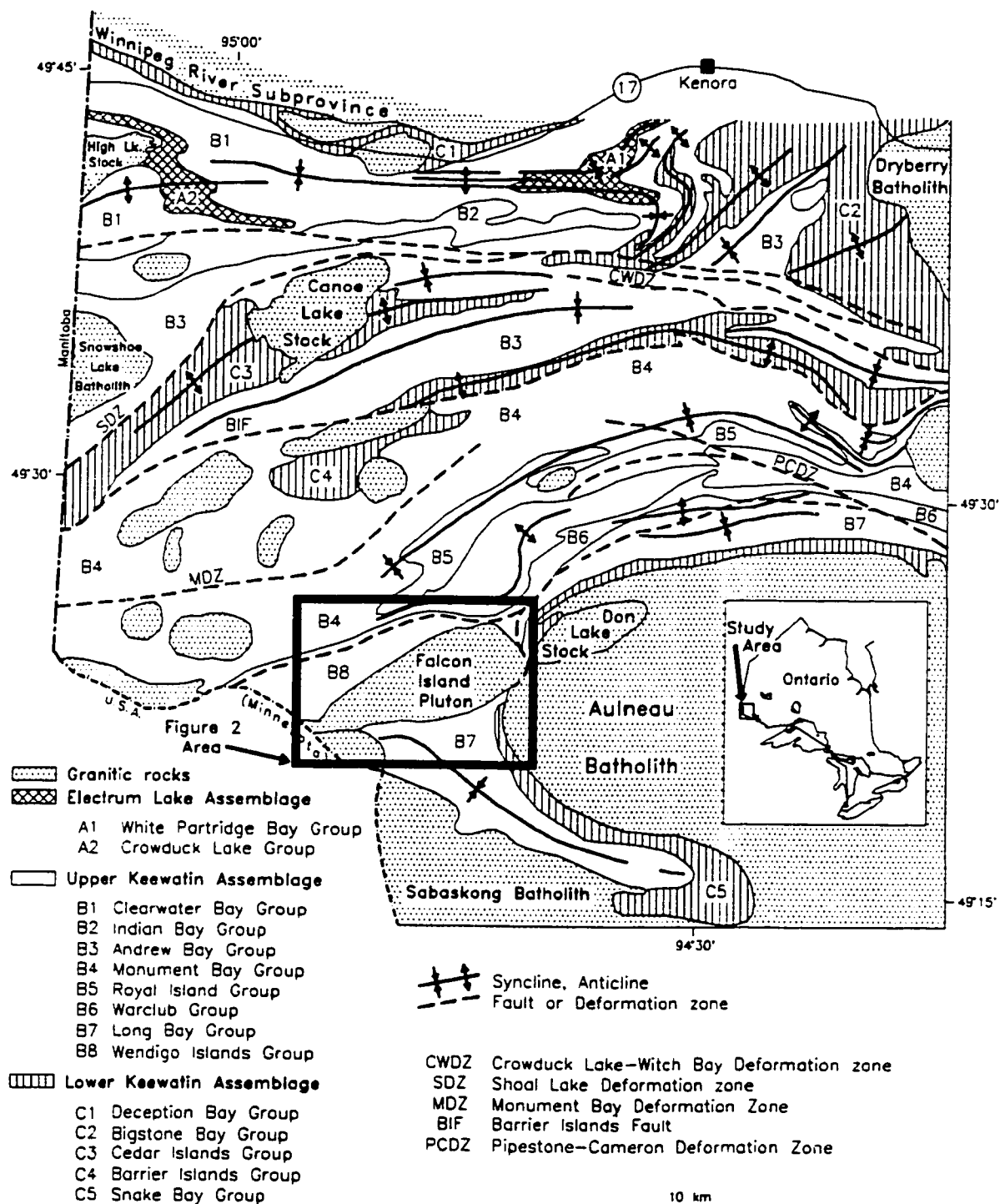


Fig. 3.1. General geology of the Lake of the Woods greenstone belt.

stocks and the alkaline complexes (including the Falcon Island pluton) (2700 to 2695 Ma).

Gold mineralization is widespread in zones of ductile deformation transecting the Lake of the Woods greenstone belt. Gold deposits with sufficient grade and tonnage to be economic appear to be restricted to the western part of the belt, where they have a “porphyry copper” association of coeval timing, metal zonations and alteration assemblages in proximity to syndeformational, calc-alkaline plutons (Davis & Smith 1991). There is no evidence of gold mineralization spatially associated with the Falcon Island pluton, despite its proximity to the southern branch of the Pipestone-Cameron deformation zone, a potentially favorable system of faults that host gold mineralization 20 km northeast of the pluton (Ayer & Buck 1989, Ayer 1991).

3.3 The Falcon Island pluton

The Falcon Island plutonic complex is the largest alkaline intrusion in the Lake of the Woods greenstone belt. It was mapped and described by Ayer (1991). U-Pb isotope analyses of titanite from an alkali feldspar syenite border phase indicates an age of crystallization of 2695 $\pm$ 3 Ma (Chapter 2). This age is similar to the documented ages of other events of alkaline and potassic magmatism in the western Wabigoon Subprovince, e.g., the Jackfish Lake pluton (2696 Ma; Sutcliffe et al. 1990), the Taylor Lake stock (2695 Ma; Davis et al. 1982) and coeval shoshonitic volcanic rocks in the Electrum Lake Assemblage in the northern part of the Lake of the Woods greenstone belt (Ayer & Davis 1997).

The rocks of the complex (Fig. 3.2) were classified in the field according to IUGS nomenclature on the basis of the relative proportions of alkali feldspar and plagioclase as indicated by staining of all hand specimens (Ayer 1991). Subsequent electron microprobe analysis of the minerals indicates the plagioclase is dominantly sodic ($An < 5\%$) and should be included with the alkali feldspar in the IUGS classification (Streckeisen 1976). Thus field-classified monzodiorites and syenites which have ferromagnesian mineral contents $> 15\%$ are herein reclassified as feldspathoid-bearing alkali feldspar melasyenites; and the field-classified monzonites which have ferromagnesian mineral contents $< 15\%$ are reclassified as feldspathoid-bearing alkali feldspar syenites. The prefix “feldspathoid-bearing” is hereafter dropped for the sake of brevity.

Alkali feldspar melasyenite and alkali feldspar syenite are the most widespread units of the intrusion (Fig. 3.2). Contacts between the melasyenite and syenite are gradational, but the presence of melasyenite xenoliths in syenite indicates it predates the syenite (Ayer 1991). Alkali feldspar melasyenite is restricted to two zones: (1) an oval core-zone about 8 km by 6 km toward the northeastern end, and (2) a marginal zone up to 3 km wide along the southwestern part of the pluton. Isolated lenticular bodies (up to 3 km by 0.5 km) of monzodiorite and clinopyroxenite occur within the melasyenitic portions of the intrusions. The mafic rock-units have a gradational contact with the melasyenite, and are interpreted as early cumulates.

The clinopyroxenite is black and consists of medium-grained subhedral clinopyroxene (65%) surrounded by medium-grained anhedral phlogopite (25%) and fine-grained subhedral hornblende (10%). Subhedral magnetite and zircon occurs as fine-grained inclusions in clinopyroxene and biotite. The monzodiorite is dark gray and consists of euhedral to subhedral,

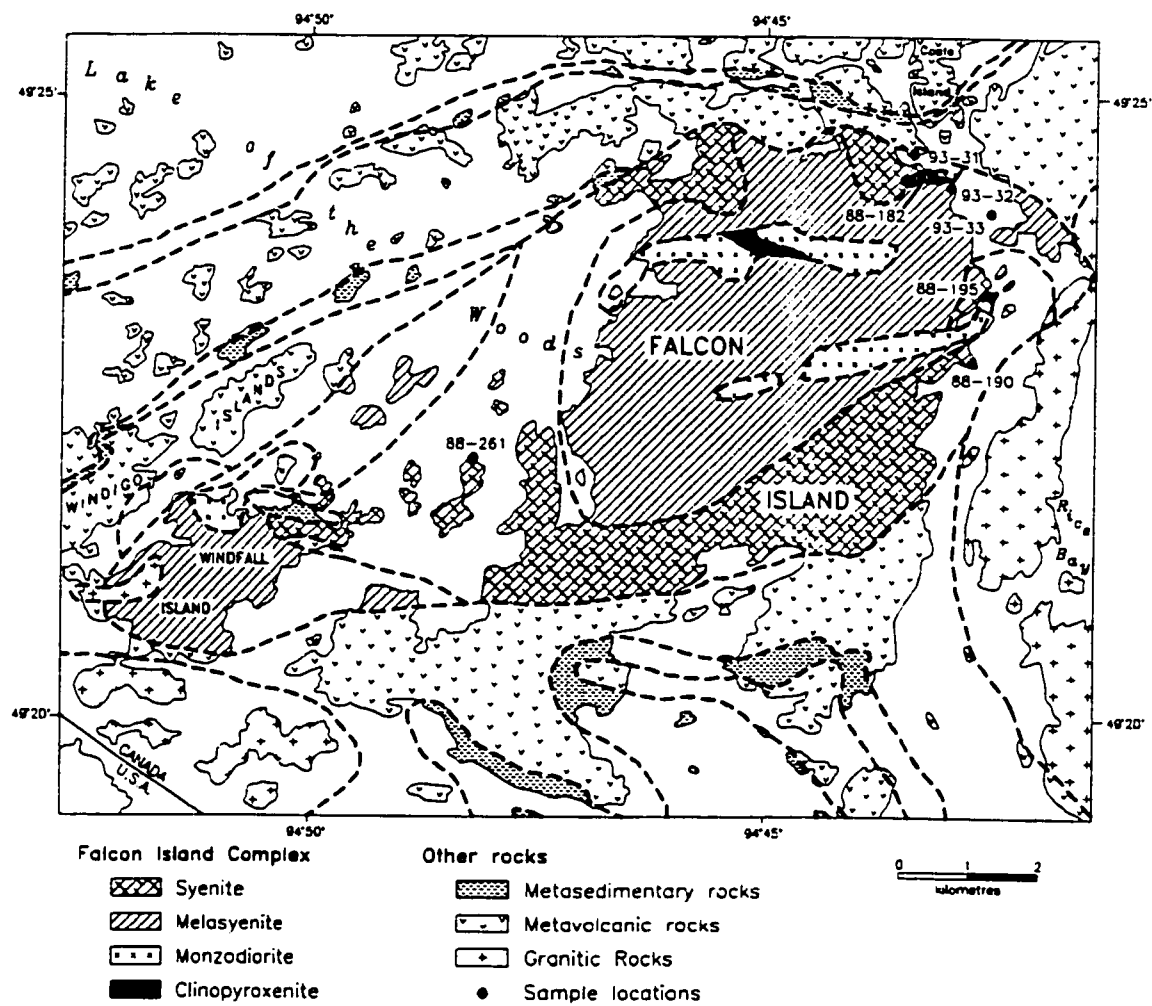


Figure 3.2. General geology of the Falcon Island pluton

medium-grained clinopyroxene (50%) surrounded by fine- to medium-grained, anhedral biotite (10%), plagioclase (30%) and perthitic alkali feldspar (5%). Fine-grained, subhedral to euhedral magnetite, apatite, and zircon occur as inclusions in clinopyroxene, biotite and feldspars. The alkali feldspar melasyenite is pink to red and consists of coarse-grained, tabular, microcline phenocrysts, locally aligned in an igneous foliation. Medium- to fine-grained, subhedral microperthitic microcline (50-70%), anhedral albite (5-10%), and subhedral to euhedral clinopyroxene (10-20%), rimmed by hornblende (5-10%) and ragged biotite (1-5%) are interstitial to the phenocrysts. Fine-grained subhedral magnetite forms inclusions in the ferromagnesian minerals and feldspars. Other accessories include fine-grained, euhedral to anhedral titanite, apatite, and zircon. Fine-grained accessory nepheline (1-5%) occurs as discrete grains with microcline and plagioclase. Subhedral to euhedral epidote and locally andradite are late-stage deuteric overgrowths. The alkali feldspar syenite is pink to gray and is texturally similar to melasyenite. It contains a higher proportion of discrete albite grains (10-20%) relative to microperthitic microcline (60-70%), and a higher proportion of biotite (5-10%) relative to clinopyroxene (1-5%) and amphibole (1-5%). Accessory minerals may include nepheline, magnetite, titanite, apatite and zircon. Replacement minerals may include epidote, andradite and chlorite. Pink to gray peraluminous alkali feldspar syenite contains up to 10% microcline phenocrysts in a fine- to medium-grained groundmass of microcline (50-60%), plagioclase (15-25%), biotite (5-10%) and muscovite (1-5%). Accessories include titanite, apatite and epidote.

Whole rock compositions of representative samples are presented in Table 3.1 and the sample locations are shown on Figure 3.2. The analytical techniques utilised are as indicated in

TABLE 3.1. WHOLE ROCK COMPOSITIONS, FALCON ISLAND PLUTON.

Rock Unit	pyroxenite	monzodiorite	melasyenite	melasyenite	syenite	syenite	syenite
Sample #	88-182	88-195	93-32	93-33	88-190	93-31	88-261
SiO ₂	46.7	46.6	52.1	56.2	55.6	55.4	59.7
TiO ₂	0.66	1.28	0.97	0.56	0.64	0.48	0.36
Al ₂ O ₃	7.13	12.7	13.7	18.0	19.2	19.8	20.0
Fe ₂ O ₃	8.58	13.7	10.7	5.41	4.75	4.23	3.04
MnO	0.14	0.19	0.19	0.11	0.09	0.10	0.06
MgO	16.7	6.91	4.22	1.82	1.56	1.23	0.66
CaO	13.1	10.1	7.36	3.38	3.23	2.08	1.68
Na ₂ O	0.89	3.34	2.97	4.23	3.33	5.73	4.13
K ₂ O	3.04	2.83	5.66	8.44	8.29	8.64	7.73
P ₂ O ₅	0.22	1.27	0.78	0.33	0.45	0.20	0.27
CO ₂	0.45	0.20			0.40		0.34
LOI	1.50	0.80	1.00	1.00	1.40	0.75	1.00
Total	98.7	99.7	100.0	99.7	98.5	98.8	98.6
TRACE ELEMENTS							
Cr	1200	161	61	17	18	7	nd
Ni	440	53	18	4	nd	3	nd
Co	51	45	21	9	11	6	5
Sc	28	22	13.9	5.46	4	3.17	2
V	104	243			75		25
Cu	17	58	73	29	22	13	98
Pb	nd	nd	2	10	nd	14	nd
Zn	79	141	102	60	74	46	69
Rb	109	105	92	245	237	289	231
Ba			2820	2380		1590	
Sr	570	2068	2710	2470	1969	1520	2200
CIPW NORMS							
Q	0.0	0.0	0.0	0.0	0.0	0.0	0.8
C	0.0	0.0	0.0	0.0	0.0	0.0	2.5
Or	0.0	17.1	34.2	50.9	50.6	52.3	46.9
Ab	0.0	7.3	13.8	15.2	18.9	12.1	35.9
An	6.7	11.6	7.5	5.3	13.4	2.8	6.8
Ne	4.2	11.7	6.4	11.5	5.6	20.4	0.0
Lc	26.0	0.0	0.0	0.0	0.0	0.0	0.0
Hy	0.0	0.0	0.0	0.0	0.0	0.0	5.0
Ol	22.0	15.0	9.5	4.9	6.7	3.8	0.0
Mt	2.1	3.4	2.6	1.4	1.2	1.1	0.8
Hm	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Il	1.3	2.5	1.9	1.1	1.3	0.9	0.7
Ap	0.5	2.8	1.7	0.7	1.0	0.5	0.6
INDICES							
ASI	0.25	0.47	0.56	0.81	0.94	0.88	1.10
DI	15.7	36.1	54.5	77.6	75.1	84.8	83.6
AI	0.67	0.67	0.80	0.89	0.75	0.95	0.76

Oxides in wt %; traces in ppm; Aluminum Saturation Index (ASI) = molar Al₂O₃/(CaO+K₂O);
Differentiation Index (DI) = normative Q + Or + Ab + Ne + Ks + Lc;
Apatitic Index (AI) = molar (Na₂O+K₂O)/Al₂O₃

Chapter 2. Duplicates and standards submitted with the samples indicate that precision and accuracy values were within 5% (at 2σ) for values greater than 10 times detection limits. The pluton is silica-undersaturated, as evidenced by normative nepheline and leucite (Table 3.1). On the basis of agpaitic indices [molar $(\text{Na}_2\text{O}+\text{K}_2\text{O})/\text{Al}_2\text{O}_3$] of less than unity (Sørensen 1974, Mitchell 1996), the pluton is miaskitic (A.I. = 0.67 to 0.95). Values of $\text{K}_2\text{O} > 3$ wt.% and $\text{K}_2\text{O} / \text{Na}_2\text{O} > 2$ indicate an ultrapotassic composition and it is similar to the Group-III ultrapotassic suites (Foley et al. 1987) on the basis of elevated Mg (in the mafic units) and Al contents.

3.4 Mineral assemblages

3.4.1 Analytical and computational techniques

Mineral analyses were carried out by wavelength-dispersion X-ray spectrometry on a Cameca electron microprobe at the Geoscience Laboratories of the Ontario Geoservices Centre, Ministry of Northern Development and Mines, Sudbury, Ontario. Operation conditions were set to 15kV and 20 nA using a beam defocused to approximately 10 μm . Count times varied from 15 to 30 seconds per element for the ferromagnesian minerals, and 10 seconds per element for feldspars and nepheline. Recalculation of pyroxene compositions is based on the scheme of Yoder & Tilley (1962). Iron recalculation to Fe^{2+} and Fe^{3+} is done by charge balance considerations. Pyroxene nomenclature and site assignments are according to current IMA guidelines (Morimoto 1989). The amphibole results are recalculated on an anhydrous basis to a total of 13 cations,

excluding Ca, Na, and K, per 23 atoms of oxygen, using the charge balance method to assign ferrous and ferric iron, and the cations assigned to each site according to IMA guidelines (Robinson et al. 1982, Leake et al. 1997). Biotite compositions are recalculated to 24 atoms of oxygen and 2 OH groups (Deer et al. 1963a). The biotite from three samples was separated using heavy-liquid techniques, and selected by hand-picking using a binocular microscope. These separates were analyzed by wet chemistry to determine ferrous iron content, and the results were utilised to adjust FeO and Fe₂O₃. The biotite-annite-siderophyllite-eastonite rectangle of Deer et al. (1963a) is utilised for nomenclature. Feldspars and nepheline are recalculated on the basis of 32 atoms of oxygen. Oxide compositions were not recalculated.

3.4.2 Clinopyroxene

Green pleochroic clinopyroxene occurs in all units except peraluminous alkali feldspar syenite. The clinopyroxene grains are subhedral, with equant to prismatic sections up to 6 mm in length. Zoning is visually evident as variation from a pale green pleochroic core to a slightly darker green rim in all units, with the exception of monzodiorite and clinopyroxenite, which exhibit uniform pale green to pale pink pleochroism. The clinopyroxene from alkali feldspar syenite and melasyenite is typically mantled by magmatic amphibole. Biotite, magnetite, apatite, and titanite may occur as inclusions in clinopyroxene.

Clinopyroxene compositions range from magnesian diopside to aegirine-augite in the various units of the complex. Compositional averages from representative samples are presented

in Table 3.2 as most of the clinopyroxenes exhibit systematic zoning from core to rim. The degree of variation is indicated by the standard deviations (Table 3.2) and the positions of the individual compositions on Figures 3.3, 3.4 and 3.5. Intragrain variation from core to rim is magmatic as indicated by its similarity to the overall differentiation trend of the complex, from early near end-member diopside in clinopyroxenite to progressively more Fe-rich diopside in monzodiorite and melasyenite and finally to Fe- and Na-rich and Ca-poor aegirine-augite in syenite (Figs. 3.3, 3.4 & 3.5).

The compositions demonstrate increasing $\text{Fe}/(\text{Fe} + \text{Mg})$ (from 0.15 to 0.43: Table 3.2) with fractionation. Inferred $\text{Fe}^{3+}/(\text{Fe}^{2+} + \text{Fe}^{3+})$ values also increase with fractionation (from 0.3 to 0.6), which might normally suggest oxidation. However, McGuire et al. (1989) have shown that calculated Fe^{2+} and Fe^{3+} values assuming electrostatic neutrality and the stoichiometry of the low-pressure end-members of Cawthorn and Collerson (1974) are suspect for pyroxene crystallized under higher pressures, the likely pressure conditions of the pluton.

Clinopyroxenes from differentiated alkaline complexes evolve with a variety of compositional trends. Typically, they evolve toward a more iron-rich composition (LeBas 1962). In many cases, such as the Shiant Isles (Gibb 1973) and the Klokken (Parsons 1979) intrusions, a “hedenbergite-enrichment” trend roughly parallels the diopside-hedenbergite join, indicating extensive substitution of Fe^{2+} for Mg without any significant substitution in the Ca site. In some alkaline intrusions, decreasing or constant $\text{Fe}/(\text{Fe} + \text{Mg})$ values are evident (e.g. Kirkland Lake and Baie-des-Moutons trends on Fig. 3.3). This type of trend, caused by incorporation of iron dominantly as Fe^{3+} in oxide phases, leading to a deficiency in Fe^{2+} and substitution of Mg for

TABLE 3.2. CHEMICAL COMPOSITION OF CLINOPYROXENE, FALCON ISLAND PLUTON

Rock Unit Sample # # of anal.	clinopyroxene		monzodiorite		melasyenite		melasyenite		syenite		syenite	
	8	s.d.	11	s.d.	9	s.d.	10	s.d.	8	s.d.	9	s.d.
SiO ₂ (wt.%)	53.35	1.14	51.16	0.16	5.25	0.39	5.51	0.80	51.00	0.65	53.70	0.25
TiO ₂	0.23	0.28	0.47	0.40	0.64	0.35	0.83	0.10	0.73	0.90	0.56	0.31
Al ₂ O ₃	1.49	0.86	2.98	0.39	3.54	0.80	3.33	0.00	3.90	0.32	3.40	0.45
FeO	4.68	1.40	9.16	0.12	11.50	2.40	1.76	1.51	9.43	0.34	12.60	0.93
Cr ₂ O ₃	0.24	0.25	0.20	0.20	0.40	0.20	0.10	0.20	0.00	0.00	0.00	0.00
MnO	0.14	0.70	0.35	0.50	0.39	0.70	0.40	0.20	0.32	0.00	0.40	0.90
MgO	15.54	0.40	12.38	0.21	1.49	1.23	1.82	0.83	11.59	0.26	9.61	0.77
CaO	23.66	0.56	22.25	0.24	21.26	1.20	21.33	0.18	21.78	0.11	19.64	1.97
Na ₂ O	0.58	0.26	0.91	0.20	1.81	0.48	1.87	0.30	1.82	0.10	2.85	1.13
K ₂ O					0.30	0.50	0.10	0.20	0.10	0.10	0.20	0.20
Total	99.90	0.17	99.66	0.22	99.49	0.15	99.87	0.22	99.77	0.17	99.44	0.37

Structural formula based on 4 cations and 6 atoms of oxygen.

TSi (a.p.f.u.)	1.96	1.91	1.89	1.89	1.89	1.89	1.89	1.89	1.89	1.89	1.89	1.89
TAI	0.04	0.09	0.11	0.11	0.11	0.11	0.11	0.11	0.10	0.10	0.11	0.11
M1Al	0.02	0.04	0.04	0.04	0.03	0.03	0.03	0.03	0.03	0.03	0.04	0.04
M1Ti	0.01	0.01	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02
M1Fe ³⁺	0.05	0.09	0.16	0.16	0.17	0.17	0.17	0.17	0.16	0.16	0.25	0.25
M1Fe ²⁺	0.07	0.17	0.18	0.18	0.17	0.17	0.17	0.17	0.13	0.13	0.15	0.15
M1Cr	0.01											
M1Mg	0.65	0.69	0.59	0.59	0.6	0.6	0.6	0.6	0.64	0.64	0.54	0.54
M2Fe ²⁺	0.03	0.03										
M2Mn		0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
M2Ca	0.93	0.89	0.86	0.86	0.85	0.85	0.85	0.85	0.87	0.87	0.79	0.79
M2Na	0.04	0.07	0.13	0.13	0.14	0.14	0.14	0.14	0.13	0.13	0.21	0.21
Fe/(Fe+Mg)	0.15	0.30	0.37	0.37	0.36	0.36	0.36	0.36	0.31	0.31	0.43	0.43
Na/(Na+Ca)	0.04	0.07	0.13	0.13	0.14	0.14	0.14	0.14	0.13	0.13	0.21	0.21
Fe ³⁺ /(Fe ²⁺ +Fe ³⁺)	0.33	0.31	0.47	0.47	0.50	0.50	0.50	0.50	0.55	0.55	0.63	0.63

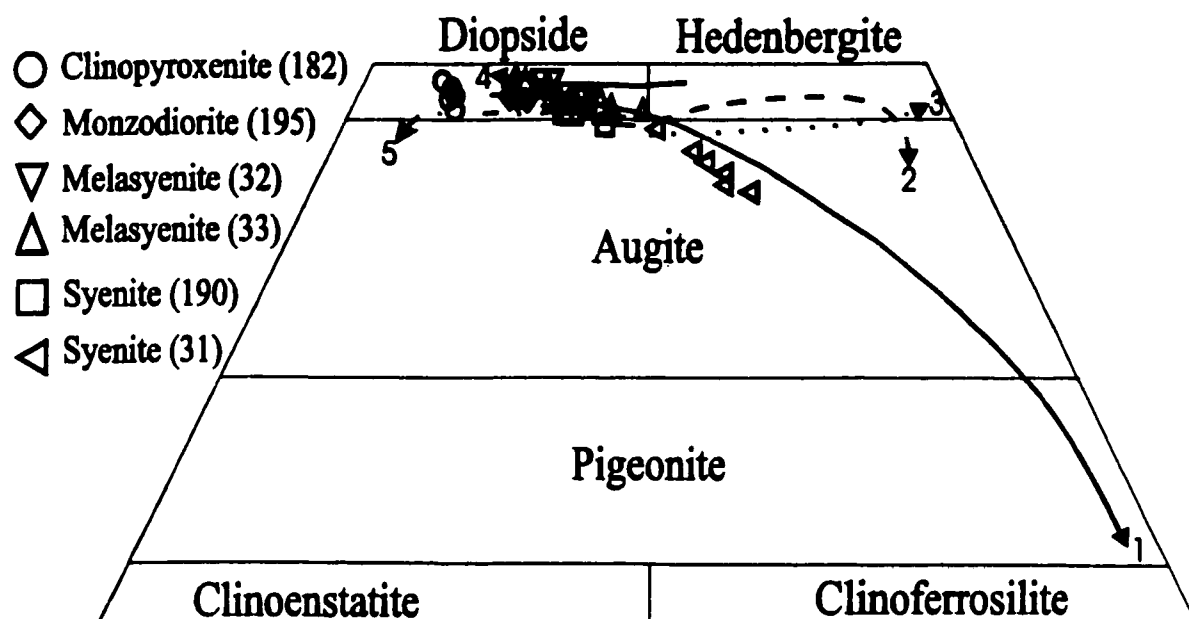


Fig. 3.3. Compositions of the clinopyroxene from the Falcon Island pluton in terms of the pyroxene quadrilateral (Morimoto 1989). For comparative purposes, trends of crystallization from a number of other differentiated alkaline intrusions are shown; (1) Shonkin Sag laccolith (Nash & Wilkinson 1970), (2) Klokken gabbro-syenite complex (Parsons 1979), (3) Kûngnât syenites (Upton 1960), (4) Baie des Moutons syenitic complex, early-group syenites (Lalonde & Martin 1983) and (5) Kirkland Lake suites (Levesque 1994).

ferrous iron, is attributed to progressively higher fugacity of oxygen with differentiation (Czamanske & Wones 1973, Lalonde & Martin 1983, Rowins et al. 1991, Levesque 1994). Clinopyroxene compositions from still other alkaline suites exhibit progressive Ca depletion from diopside toward the clinoferrosilite corner of the pyroxene quadrilateral (e.g., Falcon Island and Shonkin Sag trends on Fig. 3.3). This “agerine enrichment” trend reflects substitution of NaFe^{3+} for CaFe^{2+} and is evident on Figure 3.4, as an evolution away from the Ca-Mg-Fe plane of the quadrilateral into the field of aegirine augite, and on Figure 3.5 as progressive enrichment in Na and Fe. The inflection from iron enrichment to more sodic compositions can occur anywhere along the Fe^{2+} -enrichment trend and is a function of a complex interplay between oxygen fugacity, temperature, and the composition of the melt, most notably that of increasing Na and decreasing Ca contents (Platt 1996). The degree of silica saturation and the agpaitic index of the magma also seem to play a role, as progressively later stages of Na enrichment are typical of agpaitic undersaturated, miaskitic undersaturated, peralkaline saturated, and metaluminous silica saturated suites respectively (Fig 3.5) (Bonin & Giret 1995).

Crystallization of sodic pyroxenes is not exclusively indicative of high or increasing oxygen fugacities (Aoki 1964, Yagi 1966). Experiments have shown that aegirine is stable to conditions more reducing than those defined by the quartz-magnetite-fayalite (QFM) buffer (Bailey 1969). A sodium-enrichment trend is attributed to decreasing oxygen fugacity with fractionation at the Shonkin Sag laccolith (Nash & Wilkinson 1970), the Ilímaussaq intrusion (Larsen 1976), the Oslo rift (Neumann 1976), and the Coldwell complex (Mitchell & Platt 1978). The Falcon Island trend is similar to that of alkaline clinopyroxene from Shonkin Sag laccolith

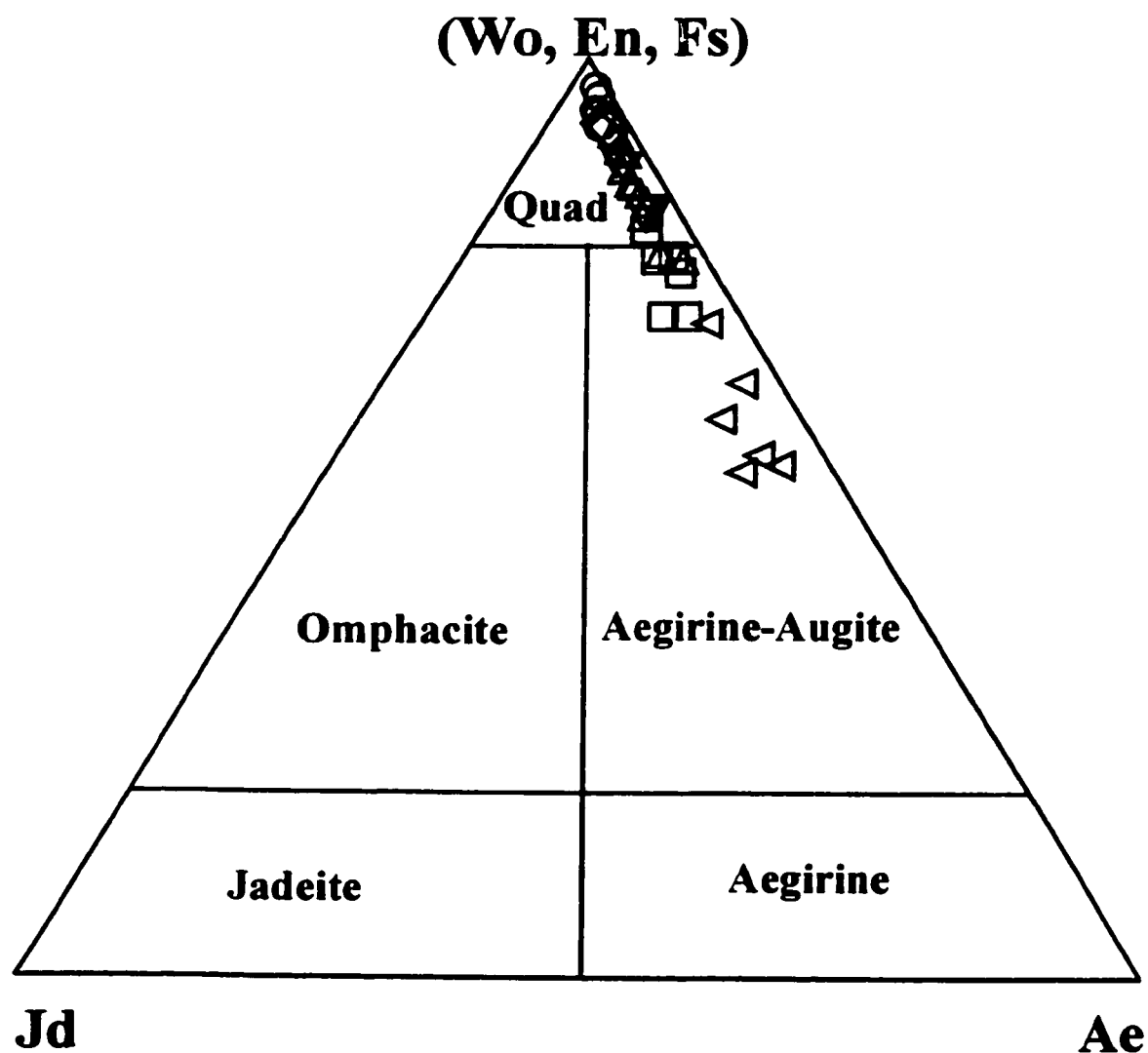


Fig. 3.4. Compositions of the clinopyroxene in the Falcon Island pluton in the quad-jadeite-aegirine plot (Morimoto 1989). Symbols as in Figure 3.3.

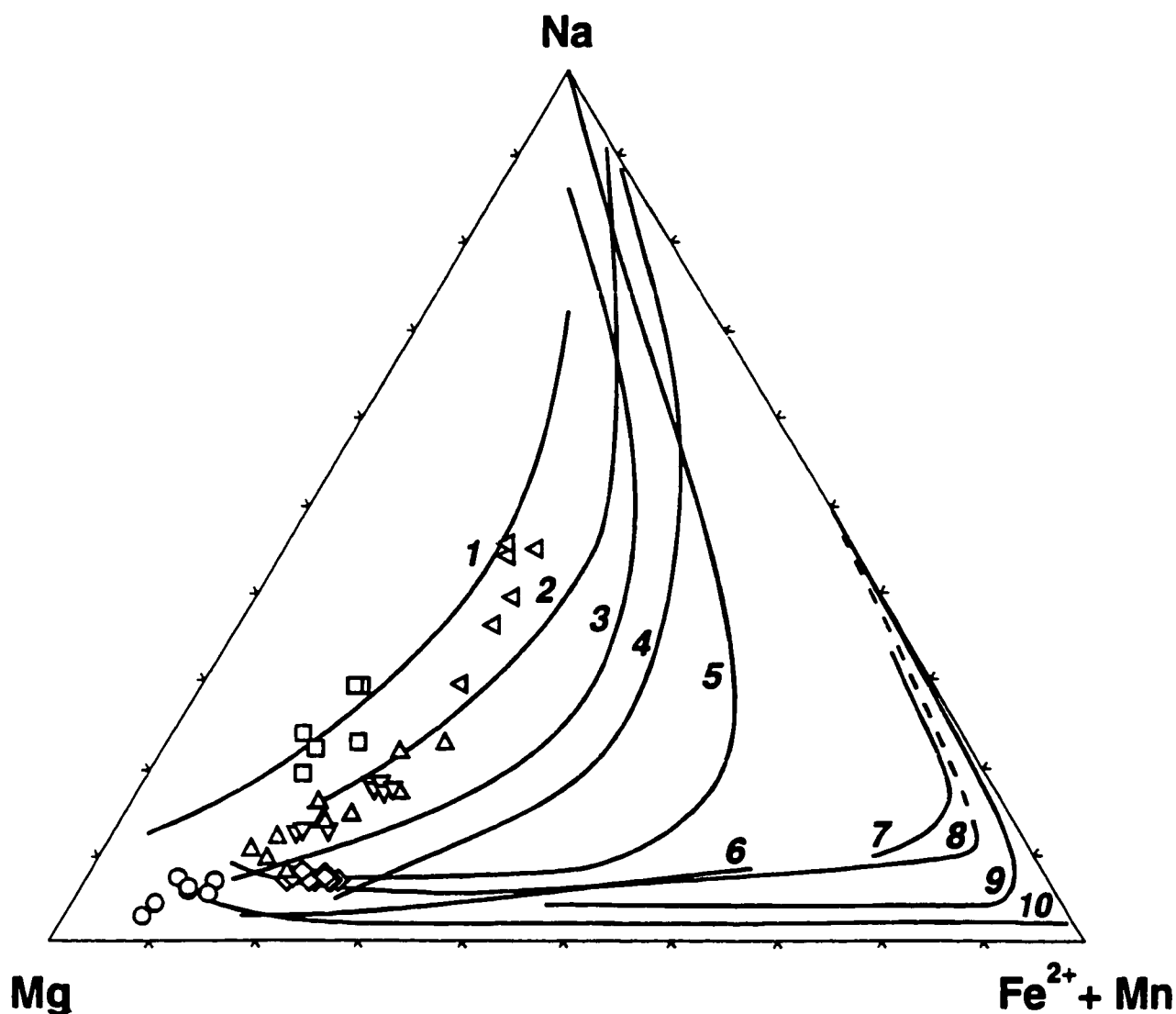


Fig. 3.5. Mg-Na-(Fe²⁺ + Mn) plot of the compositions of clinopyroxene of the Falcon Island pluton (in terms of atomic %). Symbols as in Figure 3.3. Numbered lines are trends from: undersaturated suites, (1) Auvergne, France (Varet 1969), (2) Itapirapuã, Brazil (Gomes et al. 1970), (3) Uganda (Tyler & King, 1967), (4) Morotu, Japan (Yagi 1966), (5) Qôroq, Greenland (Stephenson 1972), (9) Ilímaussaq, Greenland (Larsen 1976); saturated to oversaturated suites, (6) Japanese alkali basalts (Aoki 1964), (7) pantellerite trend (Nicholls & Carmichael 1969), (8) Nandewar volcano, Australia (Abbott 1969); and (10) subalkaline pyroxenes from the silica oversaturated Skaergaard intrusion, Greenland (Brown 1957, Brown & Vincent 1963).

(Fig. 3.3), Uganda, Auvergne, Itapirapuã (Fig. 3.5), and the Oslo rift (Neumann 1976), but does not reach the aegirine end-member compositions of these other alkaline suites. The magma at Falcon Island did not achieve the peralkalinity required (i.e., $A.I. > 1$) to form end-member aegirine (Neumann 1976, Bonin & Giret 1985). In this respect, it is similar to the trend of the Cléricy pluton, a Neoproterozoic ultrapotassic complex from the Abitibi Subprovince (Lafleche et al. 1991).

The Al content is sufficiently elevated in the clinopyroxenes and the amphiboles (see below) of all units of the pluton to fill the tetrahedral sites and to appear in the octahedral sites (Tables 3.2, 3.3). This feature is most probably a function of the composition of the magma, as the pluton is compositional similar to Group-III ultrapotassic rocks, which are rich in Al and Ca (Foley et al. 1987). Other factors which may control the Al content of mafic minerals are crystallization temperature, total pressure, oxygen fugacity and silica activity (Leake 1965, Helz 1973, Neumann 1976). Woolley et al. (1996) have documented high Al and Na contents in the clinopyroxenes of some alkaline complexes attributed to metamorphic re-equilibration. Metamorphosed clinopyroxenes from these complexes have total Al > 0.15 atoms per formula unit (a.p.f.u.), while those from igneous clinopyroxenes unaffected by metamorphism have Al < 0.15 a.p.f.u., values similar to the clinopyroxene at Falcon Island (Table 3.2).

3.4.3 Amphibole

Amphibole occurs as subhedral to euhedral crystals mantling or interstitial to

clinopyroxene. In melasyenite and syenite, the amphibole has a yellowish brown to dark green pleochroism and occurs as a mantle on clinopyroxene. In clinopyroxenite, the amphibole has a light yellow to medium green pleochroism and occurs as intergranular crystals and only rarely mantles clinopyroxene.

All units, with the exception of peraluminous alkali feldspar syenite, contain calcic amphibole. Average compositions (with standard deviations) are presented in Table 3.3, and the individual compositions are plotted on Figure 3.6 (based on the nomenclature of Leake et al. 1997). The amphibole in the clinopyroxenite ranges from edenite to magnesiohastingsite displaying decreasing Si (and correspondingly higher Al in the tetrahedral site) at relatively constant Fe and Mg contents, whereas amphibole from melasyenite and syenite range from magnesiohastingsite to hastingsite and show decreasing Mg at a relatively constant Si contents of 6 a.p.f.u. Average Fe/(Fe + Mg) values increase with fractionation (from about 0.2 in clinopyroxenite to about 0.7 in alkali feldspar syenite) (Table 3.3). Coincident increases in concentration of Mn, Na, K, and F, and decreases in Si and Ca are evident, but the amphibole suite does not achieve the degree of sodium enrichment evident in the clinopyroxene suite (Figs. 3.5 & 3.7).

3.4.4 Micas

In alkaline volcanic rocks, leucite is commonly regarded as the indicator mineral of potassic character. However, in plutonic rocks, phlogopite, biotite and alkali feldspar are the

TABLE 3.3. CHEMICAL COMPOSITION OF AMPHIBOLE, FALCON ISLAND PLUTON.

Sample #	182		32		33		31	
# of anal. / s.d.	6	s.d.	4	s.d.	3	s.d.	3	s.d.
SiO ₂ (wt.%)	49.22	2.11	39.54	0.15	38.59	1.67	38.94	1.70
TiO ₂	0.47	0.24	1.21	0.23	1.48	1.21	0.59	0.23
Al ₂ O ₃	6.68	1.68	11.91	0.17	12.46	1.82	12.21	1.63
FeO	8.34	0.67	22.45	0.02	22.96	2.21	24.90	1.61
Cr ₂ O ₃	0.16	0.06	0.01	0.01			0.02	0.02
MnO	0.18	0.05	0.56	0.05	0.59	0.02	0.65	0.02
MgO	17.67	0.73	7.60	0.19	6.48	0.92	5.76	1.22
CaO	12.2	0.41	10.55	0.03	9.33	0.40	8.75	0.12
Na ₂ O	2.04	0.42	2.69	0.01	3.33	0.30	3.72	0.20
K ₂ O	0.39	0.10	1.97	0.04	2.38	0.40	2.34	0.24
F	0.20	0.03	0.23	0.01	0.25	0.07	0.55	0.23
Cl	0.01		0.01					
Total	97.38	0.33	98.69	0.08	97.83	0.12	98.39	0.01

Structural formula based on 23 atoms of oxygen.

TSi (a.p.f.u.)	7.03	6.05	6.00	6.05
TAI	0.97	1.96	2.01	1.96
CAI	0.15	0.19	0.27	0.28
CCr	0.02			
CFe ³⁺	0.33	0.85	0.82	1.05
CTi	0.05	0.14	0.17	0.07
CMg	3.76	1.73	1.50	1.34
CFe ²⁺	0.67	2.02	2.16	2.19
CMn	0.02	0.08	0.08	0.09
BCa	1.87	1.73	1.55	1.46
BNa	0.13	0.28	0.45	0.55
ANa	0.43	0.53	0.56	0.58
AK	0.07	0.39	0.47	0.47
Sum A	0.5	0.91	1.03	1.04
Sum Cations	15.5	15.91	16.03	16.04
CF	0.09	0.11	0.13	0.27
Fe/(Mg+Fe)	0.21	0.62	0.67	0.71
Fe ³⁺ /(Fe ²⁺ +Fe ³⁺)	0.33	0.30	0.28	0.32

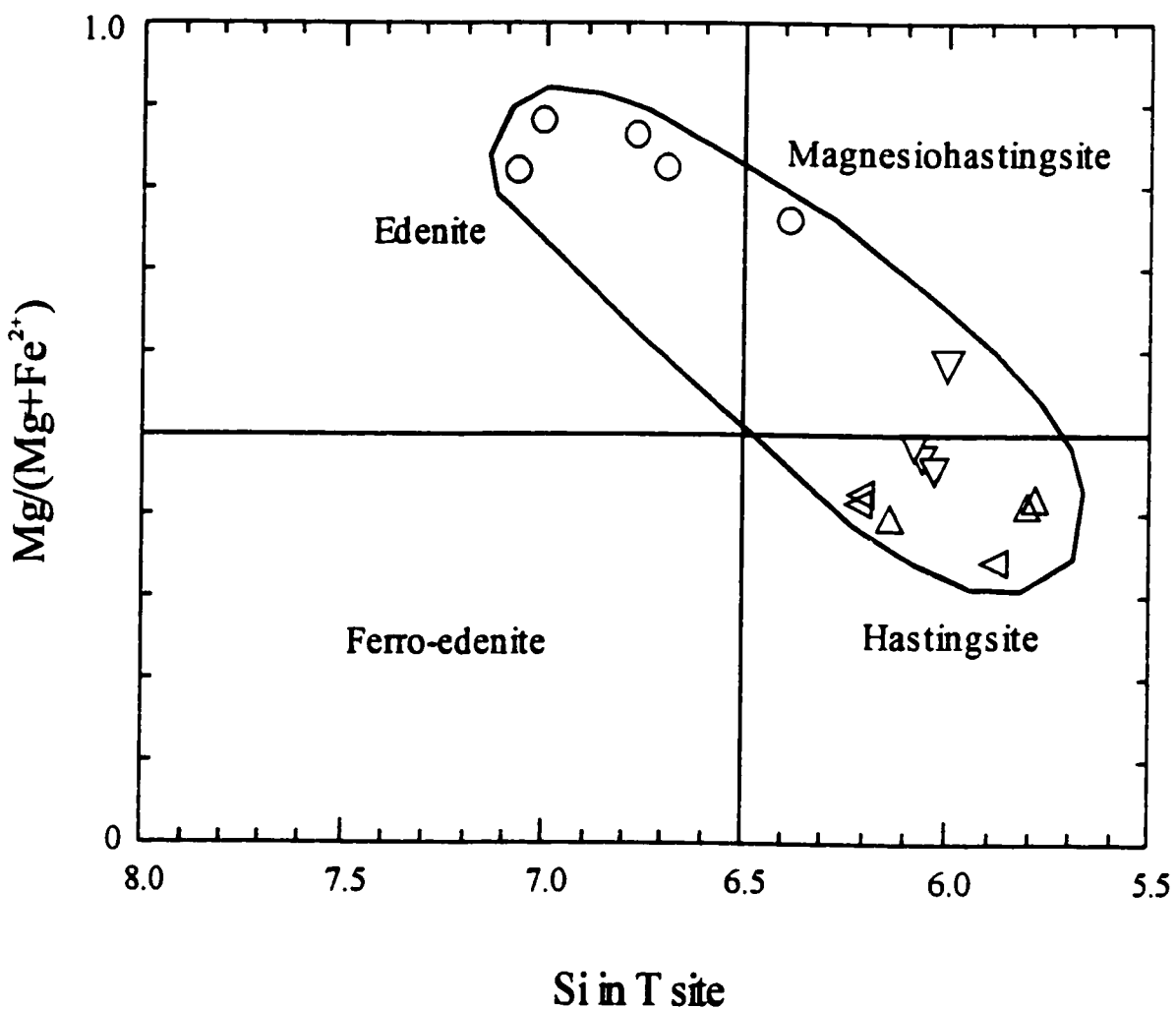


Fig. 3.6. Compositions of amphibole in the Falcon Island suite plotted on calcic amphibole quadrilateral with $\text{Ca}_B \geq 1.50$; $(\text{Na} + \text{K})_A \geq 0.50$; $\text{Ti} < 0.50$; $\text{Fe}^{3+} > {}^{\text{VI}}\text{Al}$ (Leake et al.1997).

Symbols as on Figure 3.3.

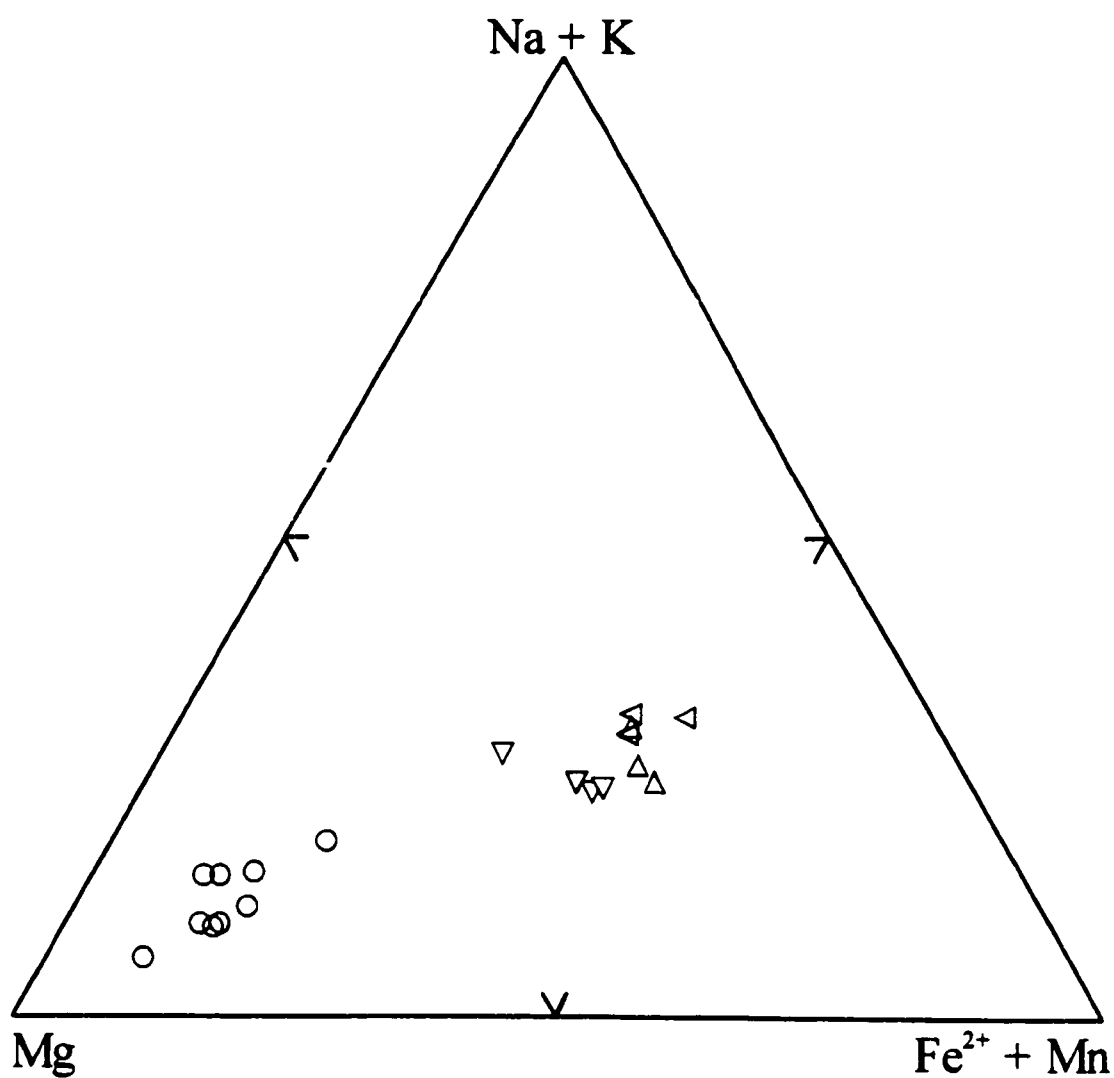


Fig. 3.7. $\text{Mg}-(\text{Na} + \text{K})-(\text{Fe}^{2+} + \text{Mn})$ plot of the compositions of amphibole in the Falcon Island suite (in terms of atomic %). Symbols as per Figure 3.3.

diagnostic minerals (Luth 1967, Barton & Hamilton 1982, Corriveau & Gorton 1993). Mica is present in all units of the complex. In clinopyroxenite, it is a light yellow-green pleochroic phlogopite, which occurs as abundant medium-grained, anhedral poikilitic grains enclosing clinopyroxene grains. Dark green pleochroic biotite with a similar habit occurs in the monzodiorite, melasyenite, and syenite. Biotite is locally replaced by dark green chlorite in all units of the complex. Muscovite and biotite, without any accompanying pyroxene, amphibole or magnetite, occur in the peraluminous syenite. In this most fractionated unit, the biotite has a dark green pleochroism and occurs as fine-grained, subhedral plates. Muscovite is non-pleochroic and occurs as fine-grained, anhedral grains intergrown with biotite and feldspar.

Results of analyses of the micas are presented in Table 3.4. Structural formulas have been calculated on the basis 24 atoms of oxygen. Wet chemical determinations of FeO in biotite separated from the clinopyroxenite, monzodiorite and peraluminous alkali feldspar syenite was used in conjunction with electron microprobe determined FeO(total) to calculate Fe_2O_3 , FeO, Fe^{2+} and Fe^{3+} . In the remaining samples, all iron was assumed to be Fe^{2+} in recalculation of the structural formulas.

No systematic intra-grain zonation was observed in any mica. Biotite and phlogopite have octahedral-site occupancies ranging from 5.75 to 6.07 a.p.f.u. (Table 3.4). Muscovite has a octahedral-site occupancy of 4.20 a.p.f.u. The Falcon Island suite displays a positive correlation of enrichment in $\text{Fe}/(\text{Fe} + \text{Mg})$ and $^{\text{IV}}\text{Al}$ with fractionation (Fig. 3.8). The only other alkaline suite displayed on Figure 3.8 that has increasing $\text{Fe}/(\text{Fe} + \text{Mg})$ with fractionation is the Shonkin Sag laccolith, which Nash & Wilkinson (1970) attribute to reduction. Other alkaline complexes, such

TABLE 3.4. CHEMICAL COMPOSITION OF MICA, FALCON ISLAND PLUTON

sample #	182	195	33	32	31	190	261									
mineral	phlogopite	biotite	biotite	biotite	biotite	biotite	biotite	muscovite								
# of anal / s.d.	5	5	5	6	5	5	7	5								
	s.	s.	s.	s.	s.	s.	s.	s.d.								
SiO ₂ (wt%)	39.65	37.05	0.14	36.11	0.15	36.36	0.93	33.97	0.31	34.23	0.21	46.00	0.29			
TiO ₂	0.75	0.10	2.38	0.15	1.20	0.41	1.05	0.13	1.22	0.17	1.31	0.31	2.12	0.24	0.48	0.16
Al ₂ O ₃	14.35	0.49	14.63	0.21	16.88	0.67	15.07	0.23	14.62	0.82	19.33	0.48	17.75	0.47	30.66	0.35
Cr ₂ O ₃	0.16	0.03	0.02	0.02	0.02	0.01	0.01	0.01	0.01	0.01	0.00	0.00	0.01	0.01	0.01	0.01
FeO	5.93		14.21		22.01	1.33	20.76	0.54	23.32	1.52	22.72	0.32	21.09		5.68	0.29
Fe ₂ O ₃	2.50		4.54										3.56			
MnO	0.11	0.02	0.27	0.03	0.54	0.07	0.47	0.06	0.57	0.04	0.60	0.04	0.47	0.03	0.04	0.01
MgO	21.62	0.42	13.29	0.19	10.20	0.70	12.40	0.44	9.87	1.19	8.44	0.19	7.50	0.31	1.39	0.17
CaO	0.00	0.00	0.00	0.01	0.01	0.02	0.02	0.01	0.06	0.06	0.01	0.01	0.01	0.01	0.01	0.01
Na ₂ O	0.21	0.03	0.10	0.02	0.08	0.03	0.05	0.01	0.07	0.01	0.08	0.01	0.08	0.03	0.18	0.03
K ₂ O	10.09	0.04	9.78	0.13	9.95	0.17	9.92	0.09	9.51	0.69	10.04	0.08	9.77	0.10	10.94	0.11
F	0.41	0.06	0.38	0.04	0.21	0.04	0.47	0.05	0.48	0.20	0.07	0.02	0.27	0.10	0.16	0.07
Cl	0.00	0.00	0.02	0.01	0.00	0.01	0.01	0.01	0.01	0.00	0.01	0.01	0.00	0.00	0.00	0.00
H ₂ O	1.80	0.03	1.70	0.02	1.76	0.02	1.64	0.03	1.61	0.09	1.83	0.01	1.73	0.05	4.31	0.03
Total	97.58		98.37		98.13		97.98		97.71		98.41		98.59		99.86	
O=F, Cl	0.21	0.03	0.16	0.02	0.09	0.02	0.20	0.02	0.20	0.09	0.01	0.01	0.11	0.04	0.07	0.03

Structural formula on the basis of 24 atoms of oxygen.

TSI (a.p.f.u.)	5.74	5.85	5.46	5.59	5.70	5.25	5.31	6.29
TAI	2.25	2.42	2.54	2.41	2.30	2.75	2.69	1.71
CAI	0.19	0.17	0.54	0.34	0.40	0.77	0.50	3.22
Ti	0.08	0.27	0.14	0.12	0.14	0.15	0.25	0.05
Fe ³⁺	0.27	0.51					0.42	
Fe ²⁺	0.72	1.79	2.85	2.69	3.06	2.94	2.73	0.65
Cr	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mn	0.01	0.03	0.07	0.06	0.08	0.08	0.06	0.00
Mg	4.67	2.98	2.35	2.86	2.31	1.94	1.73	0.28
total octahedral	5.96	5.75	5.95	6.07	5.99	5.88	5.69	4.20
Ca	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00
Na	0.06	0.03	0.02	0.02	0.02	0.02	0.02	0.05
K	1.86	1.88	1.96	1.96	1.90	1.98	1.93	1.91
Cations	15.89	15.68	15.94	16.05	15.92	15.89	15.70	14.16
F	0.36	0.36	0.21	0.46	0.48	0.07	0.26	0.14
Cl	0.00	0.01	0.00	0.01	0.00	0.00	0.01	0.00
OH	1.74	1.74	1.62	1.70	1.68	1.89	1.79	3.93
O	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00
Fe/(Fe+Mg)	0.17	0.44	0.55	0.48	0.57	0.60	0.65	0.70
Fe ³⁺ /(Fe ³⁺ +Fe ²⁺)	0.27	0.22					0.13	

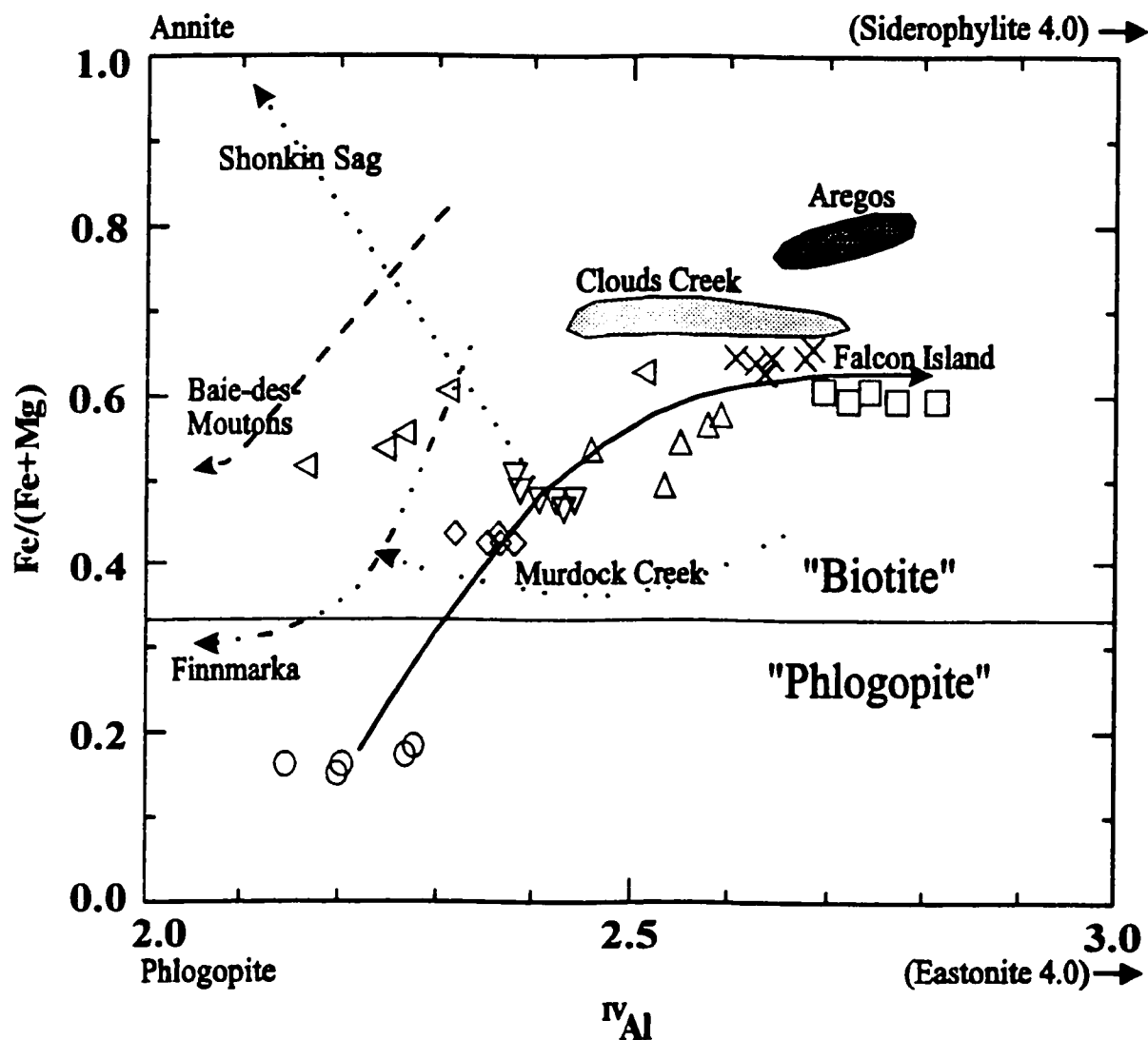


Fig. 3.8. Biotite compositions from the Falcon Island pluton plotted on part of the phlogopite-annite-eastonite-siderophyllite quadrilateral. The crystallization paths (arrows) of biotite from other differentiated alkaline complexes including Finnmarka (Czamanske & Wones 1973), Baie des Moutons (Lalonde & Martin 1983), Shonkin Sag (Nash & Wilkinson 1970) and Murdock Creek (Rowins et al. 1991) are also shown for comparison. Fields of S-type granite biotites from Aregos (de Albuquerque 1973) and Clouds Creek (Speer 1984) are also shown. Symbols as on Figure 3.3. The "x" symbols represents biotite from the peraluminous alkali feldspar syenite, sample 88-261.

as the Finnmarka (Czamanske & Wones 1973), Ben Nevis (Haslam 1968), Baie-des-Moutons (Lalonde & Martin 1983), and Murdock Creek (Rowins et al. 1991), have decreasing or flat $\text{Fe}/(\text{Fe} + \text{Mg})$ trends attributed to increasing or constant $f(\text{O}_2)$ with fractionation. A progressive increase in the $^{\text{IV}}\text{Al}$ content of biotite with fractionation is only evident in the Falcon Island suite. In fact, the trend to a biotite-muscovite, peraluminous, quartz saturated, syenite end-member [sample #261; Aluminium Saturation Index (ASI) = 1.1, normative qz = 0.8%; Table 3.1] is somewhat unusual in alkaline igneous systems. Biotite from all but one of the syenite samples (sample #31) have elevated $^{\text{IV}}\text{Al}$ contents and $\text{Fe}/(\text{Fe} + \text{Mg})$ values similar to those of peraluminous granites (Fig. 3.8) (De Albuquerque 1973, Speer 1984).

Variation in $^{\text{IV}}\text{Al}$ content in biotite may be controlled by variables such as silica activity, temperature, and total pressure (Czamanske & Wones 1973, Thompson 1974). As both the Shonkin Sag laccolith and the Falcon Island complex are interpreted to have evolved from reducing magmatic systems, differences in the $^{\text{IV}}\text{Al}$ trends of the biotite in the two intrusions may be the result of different total pressure. The Shonkin Sag laccolith was emplaced at a very shallow depth, with an estimated total pressure of 0.3 kbar (Nash & Wilkinson 1970). The Falcon Island complex is constrained between a minimum of about 3.5 kbar for igneous muscovite (Speer 1984) and a maximum of about 4 kbar defined by regional metamorphic assemblages in metasedimentary enclaves and country-rock units that contain andalusite, staurolite and almandine (Ayer 1991).

Another factor controlling Al content may be the precipitation of hornblende and clinopyroxene, both of which have relatively low ASI values and can drive a subaluminous magma system to produce mildly peraluminous late fractionates (Zen 1986). Contamination by sedimentary

rocks, which occur in the surrounding country rocks and as a large enclave in the southwestern end of the pluton (Fig. 3.2), could also be responsible for Al enrichment. Silica-saturated end-members have been attributed to assimilation of sedimentary rocks in a number of alkaline complexes (Fowler 1988, Lafléche et al. 1991, Landoll and Foland 1996).

The Ti content of the biotite at Falcon Island is relatively low and increases from 0.09 to 0.26 a.p.f.u. with fractionation. This trend may also be caused by progressive reduction with fractionation. Oxidized alkaline complexes such as Murdock Creek typically display decreasing Ti content with fractionation (Rowins et al. 1991). The Mn content in biotite also increases with fractionation from 0.02 to 0.09 a.p.f.u. A trend of increasing Mn is observed in both oxidized complexes such as Finnmarka (Czamanske & Wones 1973), Klokken (Parsons 1979) and Murdock Creek (Rowins et al. 1991) and in reduced plutons such as the Sierra Nevada batholith (Ague & Brimhall 1988).

Reduction is indicated by a decline in the $\text{Fe}^{3+}/(\text{Fe}^{2+} + \text{Fe}^{3+})$ values of biotite determined by wet-chemical analyses of biotite separates. The values decline from 0.27 in clinopyroxenite, to 0.22 in monzodiorite, to 0.13 in the peraluminous syenite (Table 3.4). Wones & Eugster (1965) have provided estimates of the compositions of biotite coexisting with alkali feldspar and magnetite under various oxygen-buffered conditions. The atomic proportions of $\text{Fe}^{2+}:\text{Fe}^{3+}:\text{Mg}^{2+}$ for three biotite separates are shown in Figure 3.9. They plot as a broad trend of increasing Fe^{2+} and decreasing Mg with fractionation within the field bounded by the nickel-nickel oxide (NNO) and hematite-magnetite (HM) buffers. Compared to other plutonic suites, the Falcon Island trend is parallel to the reducing trend of the calc-alkaline Sierra Nevada batholith (Dodge et al. 1969),

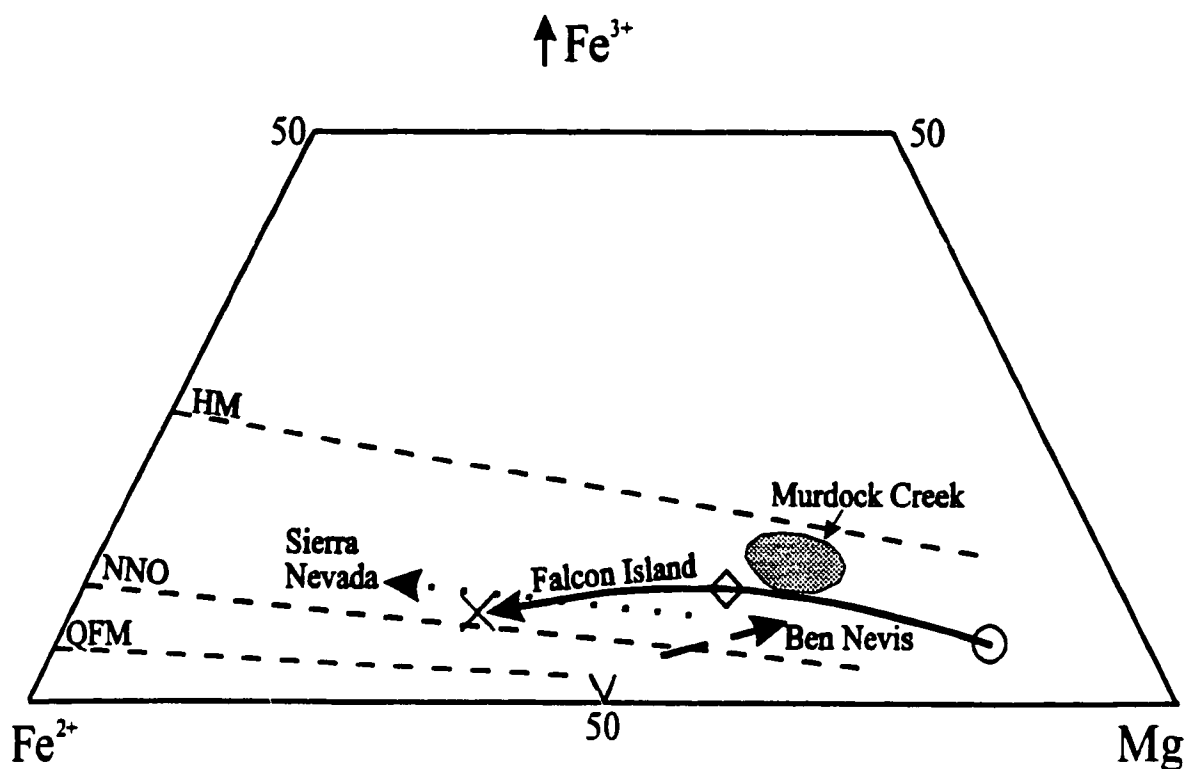


Fig. 3.9. Fe^{3+} - Fe^{2+} -Mg contents of biotite from the Falcon Island suite. Dashed lines represent compositions of "buffered" biotite, Hematite-Magnetite (HM), Nickel-Nickel Oxide (NNO) and Quartz-Fayalite-Magnetite (QFM) in the system annite-phlogopite- $\text{KFe}^{2+}_3\text{AlSi}_3\text{O}_{12}(\text{H}_1)$ (Wones & Eugster 1965). Paths of crystallization for biotite from other felsic plutons including the Sierra Nevada batholith (Dodge et al. 1969), Ben Nevis (Haslam 1968) and Murdock Creek (Rowins et al. 1991) are also shown for comparison. Symbols as on Figure 3.9.

and is opposite to that of oxidized alkaline plutonic suites such as the Ben Nevis (Haslam 1968) and Murdoc Creek (Rowins et al 1991) complexes.

Wones & Eugster (1965) relate the atomic ratio $[(\text{Fe}/(\text{Fe} + \text{Mg})) \times 100]$ during biotite crystallization to temperature and $f(\text{O}_2)$ (Fig. 3.10). The precise location of the fractionation path for the Falcon Island suite is uncertain, but the relative position is constrained by several factors: (1) $\text{Fe}/(\text{Fe} + \text{Mg}) \times 100$ values in biotite ranging from 17 to 65 (Table 3.4, Fig. 3.8); (2) the field constrained by the HM and the NNO buffers (Fig. 3.9); and (3) a trend that is progressively closer to the NNO buffer in the more fractionated rocks (Fig. 3.9), i.e., declining relative $f(\text{O}_2)$. Collectively, these features indicate decreasing $f(\text{O}_2)$ from about 10^{-10} to 10^{-15} in conjunction with a decline in temperatures from about 1000°C to 750°C and a declining relative $f(\text{O}_2)$ with fractionation. Declining relative $f(\text{O}_2)$ is a more precise indicator of reducing magmatic conditions (Frost 1991, Frost & Lindsley 1991). This extensive range from oxidized cumulates to highly reduced fractionates is clearly different from the oxidation trend of the Murdock Creek suite (Rowins et al. 1991), and is similar in trend to the Sierra Nevada batholith (Dodge et al. 1969) (Fig. 3.10). Increasing the total pressure from about 2 kbars (the pressure utilised in the experiments for construction of Fig. 3.10) to a total pressure of 4 kbars (the estimated emplacement pressure of the pluton) would shift biotite isopleths to slightly higher temperatures, but would have little effect on $f(\text{O}_2)$ (Wones & Eugster 1965).

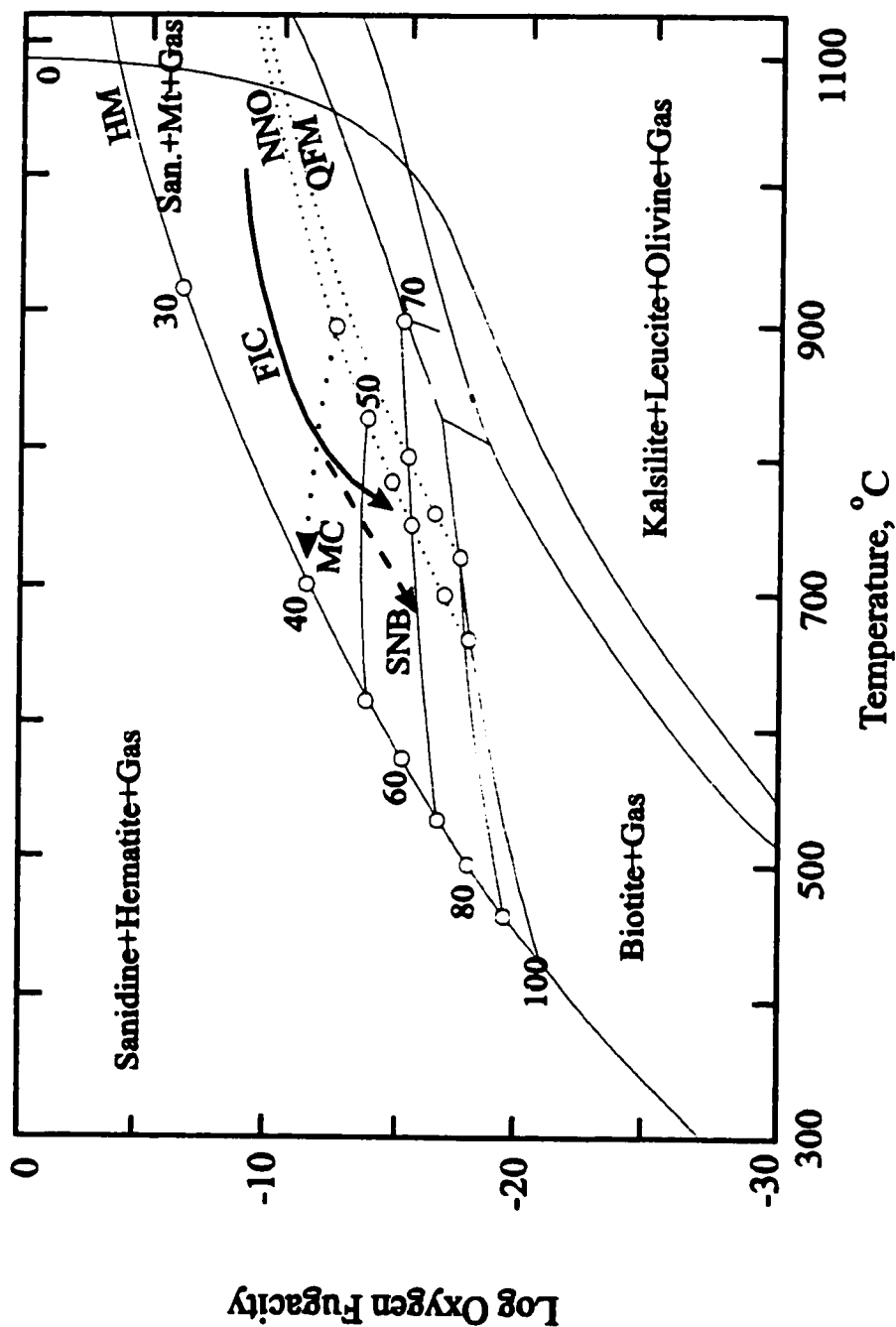


Fig. 3.10. Stability of biotite with differing $\text{Fe}/(\text{Fe}+\text{Mg})$ ratios as a function of oxygen fugacity and temperature at 2070 bars total pressure (Wones & Eugster 1965). Numbers 30-100 represent the $[\text{Fe}/(\text{Fe}+\text{Mg}) \times 100]$ values for biotite in the assemblage biotite + alkali feldspar + magnetite on the same buffer curves as in Figure 3.10. Crystallization paths for the Falcon Island complex (FIC), Murdoch Creek pluton (MC) (Rowins et al. 1991) and the Sierra Nevada batholith (SNB) (Dodge et al. 1969) are shown for comparison.

3.4.5 Feldspars and Nepheline

Coarse-grained subhedral alkali feldspar occurs as phenocrysts in monzodiorite, alkali feldspar melasyenite, and alkali feldspar syenite. These grains typically have a perthitic core with finely exsolved albite, and non-perthitic rims. Plagioclase and microcline form as small distinct anhedral crystals intergranular to the phenocrysts. Early perthitic alkali feldspar followed by coexisting alkali feldspar and plagioclase suggests a change from hypersolvus to subsolvus crystallization conditions and thus increasing $P(H_2O)$ with fractionation (Tuttle & Bowen 1958, Martin & Bonin 1976). Increasing $P(H_2O)$ is also independently supported by an evolution from early crystallized clinopyroxene to later hydrous phases such as amphibole and biotite.

The plagioclase in syenites is albite ($An_{0.75}$ to $An_{1.6}$), whereas in monzodiorite it is normally zoned and ranges from An_{32} to An_{20} (Table 3.5, Fig. 3.11). Alkali feldspar from melasyenite and monzodiorite show small decreases in K content (from Or_{92} to Or_{89}), decrease in Sr, from 0.60% to 0.33% SrO, and a substantial increase in Ba from 0.42% to 2.79% BaO. Alkali feldspar from the monzodiorite lacks the distinctive tartan twinning evident in the microcline of more fractionated units and may be a higher-temperature, more disordered variety. Barium-rich feldspar and localised inclusions of barite in the cores of large crystals of diopside indicates that the early mafic cumulates were derived from a barium- and SO_4 -enriched magma.

The nepheline occurs as fine-grained anhedral crystals interstitial to microcline and albite in syenite and melasyenite. The nepheline is a relatively Ca-poor and K-rich (Table 3.5) variety similar to that documented from other nepheline syenites (Deer et al. 1963b).

TABLE 3.5. CHEMICAL COMPOSITION OF FELDSPAR AND NEPHELINE.

mineral sample #	microcline			albite		plagioclase		nepheline
	32	33	195	32	33	195		33
						core avg.	rim avg.	
# of anal.	2	2	4	2	2	3	3	2
SiO ₂	63.30	63.89	62.56	67.62	68.54	60.77	62.06	43.55
TiO ₂	0.06	0.05	0.02		0.02	0.05	0.00	0.09
Al ₂ O ₃	18.59	18.73	18.92	19.66	19.53	24.20	23.42	33.57
FeO	0.07	0.01	0.05	0.15	0.04	0.06	0.30	0.08
MnO	0.03	0.01	0.02		0.01	0.02	0.00	
BaO	0.50	0.42	2.79	0.03	0.03	0.11	0.14	
CaO				0.34	0.16	5.72	4.93	0.37
Na ₂ O	0.89	0.94	1.11	11.62	11.63	8.10	8.57	15.99
K ₂ O	15.68	15.32	14.34	0.11	0.08	0.17	0.19	6.27
SrO	0.60	0.48	0.33	0.11	0.08	0.73	0.62	0.10
Total	99.73	99.84	100.12	99.59	100.12	99.93	100.25	100.12

Structural formula based on 32 atoms of oxygen.

Si	11.84	11.88	11.76	11.90	11.97	10.86	11.04	8.37
Al	4.10	4.10	4.19	4.07	4.02	5.09	4.90	7.59
Ti	0.01	0.01				0.01		
Fe ²⁺			0.01	0.02	0.01	0.01	0.05	0.01
Ba	0.04	0.03	0.21			0.01	0.01	0.00
Ca				0.06	0.03	1.10	0.94	0.08
Na	0.32	0.34	0.40	3.97	3.94	2.81	2.96	5.95
K	3.74	3.64	3.44	0.02	0.02	0.04	0.04	1.54
AB	7.90	8.50	10.53	97.80	98.80	71.17	75.13	
AN	0.00	0.00	0.03	1.60	0.75	27.80	23.80	
OR	92.10	91.50	89.45	0.55	0.45	1.03	1.10	

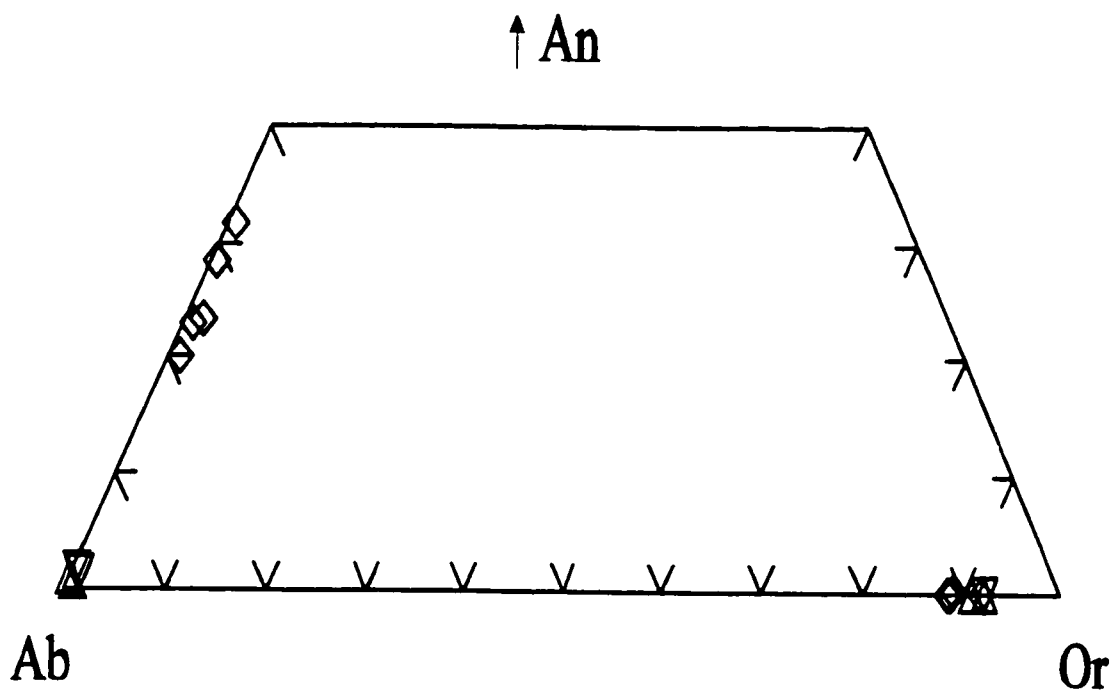


Fig. 3.11. Compositions of the feldspar in the Falcon Island pluton in terms of end member compositions for albite (Ab), orthoclase (Or) and anorthite (An) (Deer et al. 1963b). Symbols as on Figure 3.3.

3.4.6 Fe-Ti oxides

Magnetite is common in all but the most fractionated rocks of the complex. The early mafic units contain abundant magnetite and a minor amount of subhedral ilmenite crystals commonly intergrown with magnetite. Magnetite may display variable degrees of alteration to hematite, preferentially developed along the margin of grains, oriented partings and irregular fractures. In many cases the grains of magnetite also have a rim of “leucoxene” or titanite which is interpreted to be the result of subsolidus oxidation (Carmichael & Nicholls 1967).

Only the clinopyroxenite has elevated Cr in whole rock analyses (1200 ppm Cr; Table 3.1). The magnetite of clinopyroxenite contains an average of 2.35% Cr₂O₃. The other units of the pluton, are relatively Cr-poor (161 to 7 ppm Cr), as are their magnetites, e.g., the magnetite in monzodiorite has an average of 0.27 % Cr₂O₃ (Table 3.6). The mafic silicate minerals of all units are also relatively depleted in Cr, including those of the clinopyroxenite (Tables 3.2, 3.3, 3.4). The above features indicate that most of the Cr content of the magma was concentrated in magnetite and it was depleted early in the fractionation sequence. Some of the the larger diopside crystals in monzodiorite exhibit concentric bands containing abundant tiny acicular inclusions of oxide oriented at right angles. Although too small for precise electron microprobe analysis, the elevated Fe and Cr contents of these inclusions suggest they are a Cr-rich iron oxide. Similar-appearing inclusions in clinopyroxene in the Murdock Creek complex were identified as Cr-rich magnetite (Rowins et al. 1991).

Although the ferrous/ferric ratio of minerals has long been used as a monitor of oxygen

TABLE 3.6. CHEMICAL COMPOSITION OF FE AND TI OXIDES

mineral sample # # of anal /s.d.	magnetite				ilmenite	
	182		195		195	
	4	s.d.	8	s.d.	2	s.d.
SiO ₂	0.03	0.03	0.03	0.01	0.03	0.00
TiO ₂	0.02	0.01	0.25	0.41	51.06	0.11
Al ₂ O ₃	0.02	0.02	0.14	0.11	0.01	0.01
Cr ₂ O ₃	2.35	0.66	0.27	0.13	0.05	0.04
FeO	31.09	0.15	31.29	0.32	38.17	0.90
Fe ₂ O ₃	66.75	0.89	67.98	0.83	3.57	0.21
V ₂ O ₃	0.16	0.02	0.34	0.08	0.09	0.05
MnO	0.05	0.06	0.1	0.06	6.88	1.15
MgO	0.04	0.01	0.03	0.02	0.22	0.01
NiO	0.09	0.05	0.02	0.03	0.05	0.03
ZnO	0.01	0.02	0.03	0.03	0.24	0.16
CaO	0.04	0.02	0.05	0.08	0.13	0.01
Total	100.65	1.94	100.53	2.11	100.5	2.68

fugacity, a better measure of oxidation or reduction is the relative $f(\text{O}_2)$, a measure of $f(\text{O}_2)$ minus the QFM buffer values (Ghiorso & Sack, 1991). This value is commonly determined from the compositions of the Fe-Ti oxide minerals in a rock. At Falcon Island, magnetite and minor ilmenite were the only oxides to crystallize from the magma. The nearly pure end-member compositions of magnetite and ilmenite, the very low Ti content of the magnetite, and the occurrence of a rims of “leucoxene” and titanite around the magnetite crystals suggests subsolidus re-equilibration from an originally more titaniferous composition. Low-temperature re-equilibration of the oxides precludes their utilization as petrogenetic indicators of the magma’s evolution (Frost & Lindsley 1991). The current end-member compositions of < 5 mol % ulvöspinel content in magnetite and > 98% mol. % ilmenite (Table 3.6) suggests that the re-equilibration temperatures were less than 500° C (Ghiorso & Sack 1991, p. 261) and probably were related to post-magmatic hydrothermal fluids in a late metamorphic or alteration event.

3.5 Discussion and conclusions

The Falcon Island complex is an Archean, undersaturated, ultrapotassic pluton. Ayer & Davis (1997) speculated that the complex and coeval shoshonitic volcanic rocks of the Electrum Lake Assemblage in the northern part of the Lake of the Woods greenstone belt may have been derived from a mantle-wedge source beneath a Neoarchean mature island arc or a continental arc source. This hypothesis was based on chemical characteristics such as highly fractionated rare-earth element (REE) patterns, enrichment in large-ion lithophile elements, and depletion in high

field-strength elements (HFSE) similar to those of modern arc suites. Low-degree partial melts of mantle material beneath island arcs tend to have high contents of alkalis and REE, low HFSE and high oxygen fugacities (Foley & Wheller 1990).

The Falcon Island magma evolved with increasing $P(H_2O)$, as indicated by the transition from early crystallization of clinopyroxene to amphibole and biotite, and from an early perthitic hypersolvus feldspar to a later subsolvus assemblage of microcline and albite. An "agerine-enrichment trend" is evident in the clinopyroxene as progressive depletion in Mg and Ca, and enrichment in Fe and Na (both as zoning in individual grains and in the successively fractionated units) (Figs. 3.3, 3.4 and 3.5). Mg-rich diopside in clinopyroxenite was followed by edenite and phlogopite. In monzodiorite, diopside was followed by biotite and normally zoned plagioclase (An_{32} to An_{20}). In alkali feldspar melasyenite and alkali feldspar syenite, Fe-rich diopside cores evolved to aegirine augite and hastingsite rims and was followed by albite ($An_{1.6}$ to $An_{0.7}$).

Elevated Al is evident in whole rock compositions and is reflected in the Al content of clinopyroxene, amphibole and biotite, characteristics which are typical of the Group-III ultrapotassic suites from modern collisional orogenic areas (Foley et al. 1987). Based on its mineral assemblage, and those of metasedimentary enclaves, the Falcon Island pluton was emplaced at mid-crust. The relatively elevated ^{VI}Al -content of the clinopyroxene is similar to that of the alkaline Murdock Creek pluton in the Kirkland Lake camp of the Abitibi Subprovince, emplaced at a similar crustal depth (Rowins et al. 1991). However, differences in Al evolution with fractionation is evident in the biotite trends of the two intrusions. Biotite at Murdock Creek shows a depletion in ^{IV}Al at constant $Fe/(Fe + Mg)$, whereas at Falcon Island the biotite displays

increasing ^{IV}Al and $Fe/(Fe + Mg)$ with fractionation. The differing trends are probably the result of a variety of factors including: differing $f(O_2)$ paths, differing relative ASI values, the types and relative abundance of precipitating minerals, and differing crustal contamination sources.

The lack of olivine (or olivine pseudomorphs) in all rocks and the presence of titanite, magnetite and diopside in the early crystallized mafic units, indicates crystallization under more oxidizing conditions than those defined by the QFM-buffer (Carmichael & Nicholls 1967, Neumann 1976). During ascent and cooling, magmas tend to follow temperature- $f(O_2)$ paths parallel to major buffers. Loss of oxidized or reduced, solid or gaseous species produces deviations from this path. Formation of minerals with a higher proportion of Fe^{3+} than are present in the melt, such as magnetite, causes reduction.

The redox conditions of alkaline complexes may have economic significance. Late Archean alkali-rich intrusive rocks have been implicated in models for gold mineralization on the basis of their close spatial and temporal association (Rock & Groves 1988, Wyman & Kerrich 1989, Sutcliffe et al. 1990, Robert 1997). Fyon et al. (1989) suggested that gold mineralization and alkaline magmatism may be a manifestation of the cratonization process involving mantle-derived melts and CO_2 . Cameron & Hattori (1987) have proposed that gold mineralization in the Kirkland Lake camp originates from oxidized hydrothermal fluids derived from magmatic sources. Many of the syenites associated with mineralization in the Kirkland Lake gold camp represent oxidized magmas with possible links to the gold-bearing fluids (Cameron and Carrigan 1987; Rowins et al. 1991, Levesque 1994),

Gold mineralization environments in the Lake of the Woods greenstone belt have both

porphyry and shear-zone associations (Davis & Smith 1991). There is, however, no evidence of anomalous gold mineralization in the vicinity of the Falcon Island pluton, despite its alkaline nature, and the proximity of potentially favorable shear-zones (Ayer 1991). This fact suggests a negative correlation between gold mineralization and reduced alkaline plutons. The Falcon Island complex is comparable in its array of rock types and depth of emplacement to syenitic complexes associated with gold mineralization in the Kirkland Lake camp, but there are significant differences in the redox state of the evolving magmas. It would therefore appear that alkaline complexes with progressively reduced magmatic fluids, such as the Falcon Island complex, have less potential to produce the oxidized hydrothermal fluids conducive to gold mineralization. Thus analyzing the composition of ferromagnesian minerals in a differentiated complex, and in particular the biotite, may provide an additional tool for evaluating their potential to produce gold deposits.

Chapter 4

Petrogenesis of the Late-Tectonic Potassic Suite in the Lake of the Woods Greenstone Belt: Implications for Neoproterozoic Subduction-Related Magmatism in the Superior Province.

4.1 Introduction

A late-tectonic potassic igneous suite within the Neoproterozoic Lake of the Woods greenstone belt (LWGB) consists of three chemically distinctive rock series exemplified by: (1) silica undersaturated, ultrapotassic rocks at the Falcon Island pluton, (2) silica saturated monzodiorite at the Don Lake stock, and (3) shoshonitic volcanic rocks in the metasedimentary basins of the Electrum Assemblage. Potassic igneous rocks (*senso stricto*) are defined by molar K_2O/Na_2O ratios of about, or slightly higher than, unity (Peccerillo 1992). In modern within-plate tectonic settings, potassic igneous rocks tend to occur as isolated geological bodies, whereas those from subduction-related settings normally occur as the end-members of a continuous igneous rock spectrum that might range from boninites and tholeiites to high K calc-alkaline rocks and shoshonites during arc evolution (Morrison 1980, Müller and Groves 1997). The similarity of the LWGB potassic suite to Phanerozoic arc analogues has led to its interpretation as a manifestation of a Neoproterozoic, late-tectonic, subduction-related magmatic event (Ayer and Davis, 1997).

Stern et al. (1989) have defined the monzodioritic rocks in the Superior Province as a mantle-derived, sanukitoid-like suite based on chemical characteristics such as 55-60 wt % SiO_2 , $\text{Mg\#} > 0.6$, Ni & Cr both > 100 ppm, $\text{Na}_2\text{O} + \text{K}_2\text{O} > 6$ wt %, Sr & Ba both 600-1800 ppm and elevated light rare earth elements (LREE) without Eu anomalies. These characteristics are similar to the K-rich alkaline, shoshonitic and calc-alkaline magmatic association of the Sunda arc in Indonesia (Wheller et al., 1987). Ultrapotassic and shoshonitic magmatism similar to that of the Superior Province, attributed to late stage subduction events, are documented from the Neoproterozoic Ottawa orogeny (ca. 1080 Ma) of the Grenville Province (Corriveau et al., 1993) and the Paleozoic Caledonian orogeny (ca. 410 Ma) of NW Scotland (Thompson and Fowler, 1986). Thus the formation of late-tectonic, subduction-related, ultrapotassic to shoshonitic rock association was a remarkably consistent plate tectonic process spanning the last 2.7 billion years of Earth's history. This chapter utilises a geochemical and isotopic approach to determining relevant magmatic processes that have influenced the distinctive chemical trends of the LWGB potassic suite.

4.2 Regional geological setting

The regional geological setting of the LWGB has been described in Chapter 2 and the setting and characteristics of the Falcon Island pluton in Chapter 3.

The 9 km by 3.5 km Don Lake stock is located immediately northeast of the Falcon Island pluton and intrudes the northwest margin of the Aulneau batholith (Fig. 4.1). It was originally mapped and included as a unit of the batholith by Ziehlke (1974). Ayer (1991)

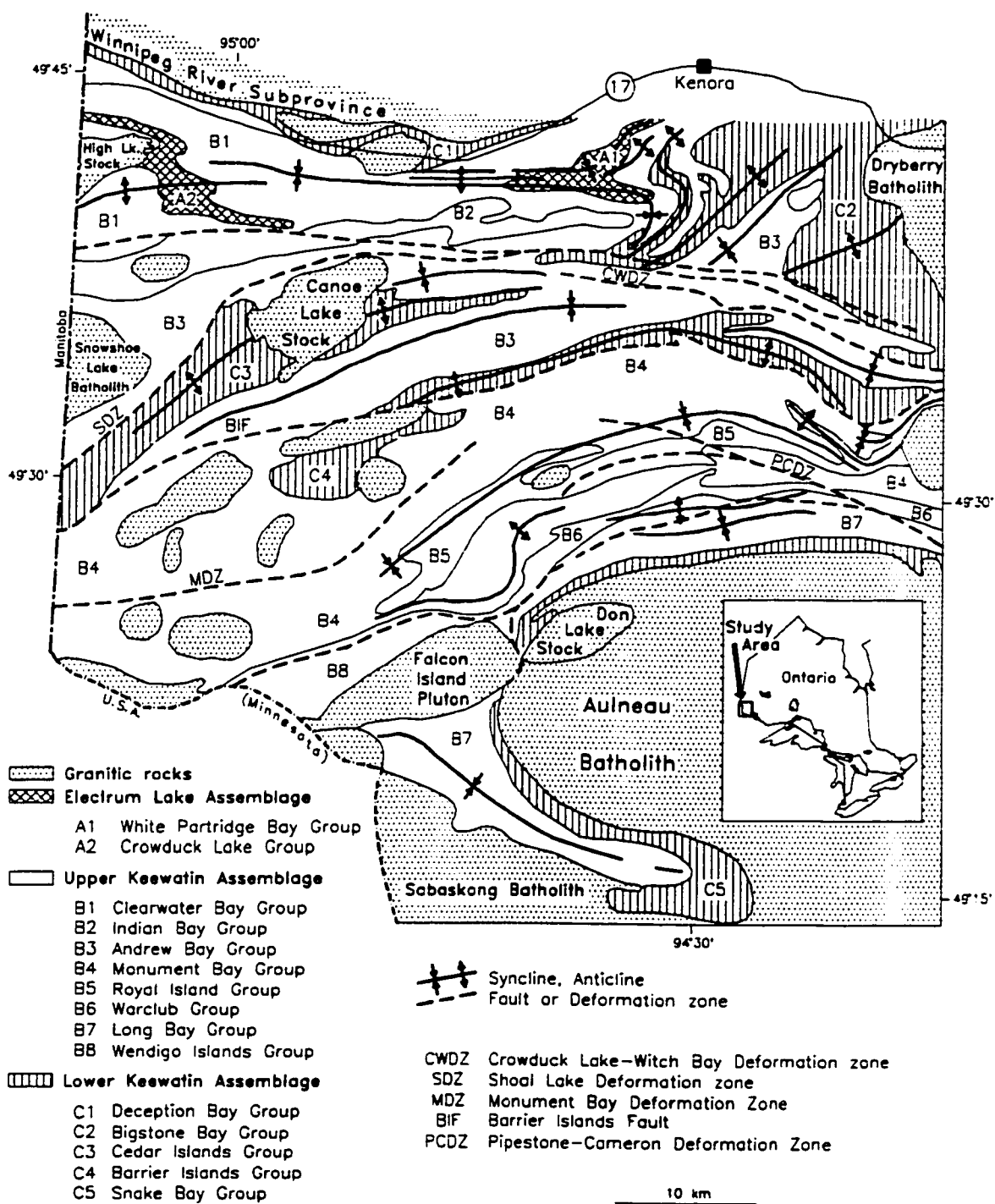


Fig. 4.1. General geology of the Lake of the Woods greenstone belt.

re-interpreted the age of the stock as consanguineous with the Falcon Island pluton (2695 Ma), based on similar mineralogy and lithogeochemistry (discussed below), a positive aeromagnetic signature, similar to the melasyenitic parts of the Falcon Island pluton, and a lack of tectonic fabric. This contrasts the adjacent tectonically foliated, tonalite and granodiorite of the Aulneau batholith which ranges in age from 2710-2718 Ma (Davis and Edwards; 1986). The stock is more homogeneous than the Falcon Island pluton. It consists of pinkish-gray monzodiorite with medium-grained, subhedral-granular plagioclase, clinopyroxene, hornblende and biotite and fine-grained, intergranular, anhedral microcline (10%), accessory magnetite (1%), epidote (tr.) and apatite (tr.). Plagioclase is fresh and normally zoned from andesine cores to oligoclase rims. Where present, clinopyroxene occurs as early crystallized cores rimmed by hornblende, similar in habit to that of the Falcon Island units (Chapter 3). Biotite is also similar in habit to the Falcon Island units, occurring as ragged subhedral crystals interstitial to earlier-formed clinopyroxene and hornblende.

Shoshonites occur in a unit about 2 km in length and 0.75 km thick in the northwestern part of the Crowduck Group of the Electrum Assemblage (Fig. 4.1). The rocks consist of fine-grained, dark to medium gray, massive flows and pyroclastic units of intermediate composition. Both units are porphyritic, containing fine- to medium-grained euhedral clinopyroxene and plagioclase phenocrysts and have undergone greenschist facies metamorphism resulting in a weak to moderate tectonic foliation, replacement of clinopyroxene by amphibole and plagioclase by albite and epidote. Groundmass minerals consist of amphibole, chlorite, biotite, plagioclase, sericite, epidote and carbonate. The chemical characteristics of calc-alkaline felsic flows and

pyroclastic rocks within other parts of the Crowduck and the White Partridge groups of the Electrum Assemblage are documented in Chapter 2.

4.3 Geochemistry

The whole rock compositions of representative samples of the potassic suite are presented in Table 4.1. The analytical techniques utilised are documented in Chapter 2. Duplicates and standards submitted with the samples indicate that error is within 5% (at 2 σ) for values greater than 10 times detection limits (Appendix I). The samples were screened for high volatile content and alkali alteration by rejection of samples outside of the igneous spectrum of Hughes (1972) or an alteration index of less than 25 or greater than 65 for non-ultramafic samples (Date et al., 1983).

Geochemical results include seven samples from the Falcon Island plutonic complex, three from the Don Lake stock and three from the Electrum Assemblage (Table 4.1). The relatively elevated volatile contents in the volcanic rocks of the Electrum Assemblage (LOI > 1.2%) reflect a higher proportion of hydrous minerals and carbonatization in the supracrustal rocks than is evident in the intrusions. Units of the Falcon Island pluton are devoid of modal quartz and contain up to 5% modal nepheline. Most samples contain normative nepheline and olivine and the clinopyroxenite also contains normative leucite. While present in calculated norms, olivine, leucite and hypersthene are not present modally because of the total pressure and oxygen fugacity conditions of the crystallizing magma (Chapter 3). The most evolved unit of the pluton is a silica saturated syenite containing normative quartz and hypersthene (Table 4.1).

Table 4.1. Whole Rock Compositions, Alkaline Suites, Lake of the Woods Greenstone Belt

Sample Rock	Falcon Island Pluton							Don Lake Stock Electrum Assemblage					
	182 C	195 M	32 MS	33 MS	190 S	31 S	261 PS	54 M	2117 M	2144 M	17 F	39 T	40 F
SiO ₂ (wt%)	46.7	46.6	52.1	56.2	55.6	55.4	59.7	56.2	53.2	57.2	55.5	55.7	56.9
TiO ₂	0.66	1.28	0.97	0.56	0.64	0.48	0.36	0.98	1.11	0.86	0.66	0.64	0.55
Al ₂ O ₃	7.13	12.7	13.7	18.0	19.2	19.8	20.0	16.2	15.6	16.0	13.6	14.4	14.8
Fe ₂ O ₃	8.58	13.7	10.7	5.41	4.75	4.23	3.04	8.19	9.66	7.36	9.22	8.82	5.59
MnO	0.14	0.19	0.19	0.11	0.09	0.10	0.06	0.11	0.15	0.11	0.15	0.16	0.09
MgO	16.7	6.91	4.22	1.82	1.56	1.23	0.66	3.23	4.42	3.66	6.04	5.58	2.69
CaO	13.1	10.1	7.36	3.38	3.23	2.08	1.68	5.37	6.53	5.82	7.55	7.00	5.90
Na ₂ O	0.89	3.34	2.97	4.23	3.33	5.73	4.13	4.85	4.28	4.35	2.46	2.36	3.42
K ₂ O	3.04	2.83	5.66	8.44	8.29	8.64	7.73	3.11	3.22	2.69	2.38	3.07	3.36
P ₂ O ₅	0.22	1.27	0.78	0.33	0.45	0.2	0.27	0.44	0.48	0.29	0.23	0.22	0.25
CO ₂	0.45	0.20			0.40		0.34	0.20	0.32	0.16			
LOI	1.50	0.80	1.00	1.00	1.40	0.75	1.00	0.50	0.70	0.80	2.20	1.20	5.00
Total	98.7	99.7	100.0	99.7	98.5	98.8	98.6	99.2	99.3	99.1	99.9	99.1	98.6
CIPW NORMS													
Qz	0.0	0.0	0.0	0.0	0.0	0.0	0.8	3.3	0.1	6.9	12.4	11.6	12.3
Ne	4.2	11.7	6.4	11.5	5.6	20.4	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Ol	22.0	15.0	9.5	4.9	6.7	3.8	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Hy	0.0	0.0	0.0	0.0	0.0	0.0	5.0	4.2	5.4	5.3	10.2	10.1	2.6
TRACE ELEMENTS													
Ba (ppm)			2820	2380		1590					821		
Rb	109	105	92	245	237	289	231	114	77	100	64	104	132
Sr	570	2068	2710	2470	1969	1520	2200	997	1104	997	722	597	631
Th			6.1	7.8		7.2					8.0	14.0	11.0
U			0.64	1.05		0.87					2.07		
Pb			2	10		14							
Y	20	43	29	25	15	19	10	18	21	15	21	18	17
Zr	125	145	180	149	126	136	181	156	112	171	126	130	149
Nb	14	nd	0	0	nd	11	8	8	nd	nd	6	3	
Ta			0.19	0.55		0.84					0.42		
Hf			4.53	3.62		3.20					2.96		
Ni	440	53	18	4	nd	3	nd	10	56	62	38	57	35
Cr	1200	161	61	17	18	7	nd	55	92	105	258	209	148
Sc	28	22	14	5	4	3	2	10	14	12	25	23	14
V	104	243			75		25	105	137	118	107	146	107
Co	51	45	21	9	11	6	5	21	28	22	22	29	18
La	34	105	111	82	76	59	94	49	52	44	30	30	29
Ce	105	240	228	160	170	116	180	110	115	94	58	66	65
Pr	15	34			21		18	13	14	12			
Nd	70	140	107	67	77	47	59	50	61	46	27	31	28
Sm	13.0	25.0	19.5	11.7	13.0	7.9	8.1	9.1	11.0	7.9	5.0	6.0	4.9
Eu	3.3	6.2	5.4	3.2	3.3	2.4	2.6	2.3	3.0	2.2	1.5	1.5	1.3
Gd	8.7	17			8.7		4.5	6.5	7.4	5.2			
Tb	0.9	2.0	1.63	0.83	0.98	0.69	0.45	0.73	0.86	0.6	0.64	0.62	0.45
Dy	4.3	9.0	6.25	3.73	4.2	3.45	2.3	3.8	4.2	2.9	3.18	3.7	2.6
Ho	0.66	1.4	1.03	0.7	0.65	0.53	0.38	0.61	0.7	0.51	0.6	0.69	0.49
Er	1.5	3.3			1.4		0.95	1.6	1.8	1.3			
Tm	0.16	0.37			0.16		0.11	0.18	0.22	0.17			
Yb	0.98	2.2	1.49	1.03	0.89	0.90	0.80	1.40	1.40	1.10	1.77	1.70	1.30
Lu	0.15	0.32	0.24	0.17	0.14	0.13	0.10	0.19	0.2	0.16	0.26	0.24	0.21
ε _{Nd} (2.7Ga)	2.08	1.78	1.93	2.20	2.01	2.24	3.07				0.41		

C-clinopyroxenite, M-monzodiorite, MS-melasyenite, S-syenite, PS-peraluminous syenite, F-flow, T-tuff.
 Note: samples 17, 31, 32, & 33 had the REE and HFSE analysed by INAA. See appendix 1 for details.

Modal quartz was not observed in the Don Lake monzodiorite nor the Electrum Assemblage shoshonite, but normative quartz and hypersthene are evident in analyses of these samples.

There appears to be a general trend from silica-undersaturated to -oversaturated compositions in the LWGB potassic suite. The Falcon Island pluton is predominantly silica-undersaturated and ultrapotassic (Chapter 3) but its evolution trend yields progressively less undersaturated phases, culminating in silica-saturated syenite (0.8 wt % normative Qz) (Table 4.1). Monzodiorite of the Don Lake stock is silica-saturated (0.1-6.9 wt % normative Qz), while shoshonite of the Electrum Assemblage is silica-oversaturated (i.e., > 10% wt % normative Qz). On a total alkali vs. SiO₂ diagram (Fig. 4.3c), the Falcon Island and Don Lake samples plot within the alkaline field, while the Electrum Assemblage samples straddle the boundary into the subalkaline field. Silica-undersaturated units of the Falcon Island pluton plot in the ultrapotassic field on a K₂O vs. SiO₂ diagram (Fig. 4.3a), but the silica-saturated syenite of the pluton, the Don Lake stock and the Electrum Assemblage samples all plot in the shoshonitic field. Shoshonites are potassic igneous rocks which commonly occur in subduction zones (Morrison 1980). The shoshonite association is geochemically defined by high total alkalis (Na₂O + K₂O > 5 wt %), high K₂O/ Na₂O ratios (> 0.6), low TiO₂ (< 1.3 wt %), high but variable Al₂O₃ (14-19 wt %), and strong enrichment in LILE and LREE (Morrison 1980).

Harker diagrams (Figs. 4.2, 4.3 & 4.4) indicate the degree of compositional continuity of the LWGB potassic suite by variation of the major elements versus SiO₂. In general, Fe₂O₃, MgO, CaO, TiO₂, P₂O₅, MnO, Ni, Cr, Co, Sc and V are negatively correlated with increasing silica content while Al₂O₃ increases with increasing SiO₂. Roughly linear trends are evident for the Falcon Island samples on many of the Harker diagrams. However, the Falcon Island

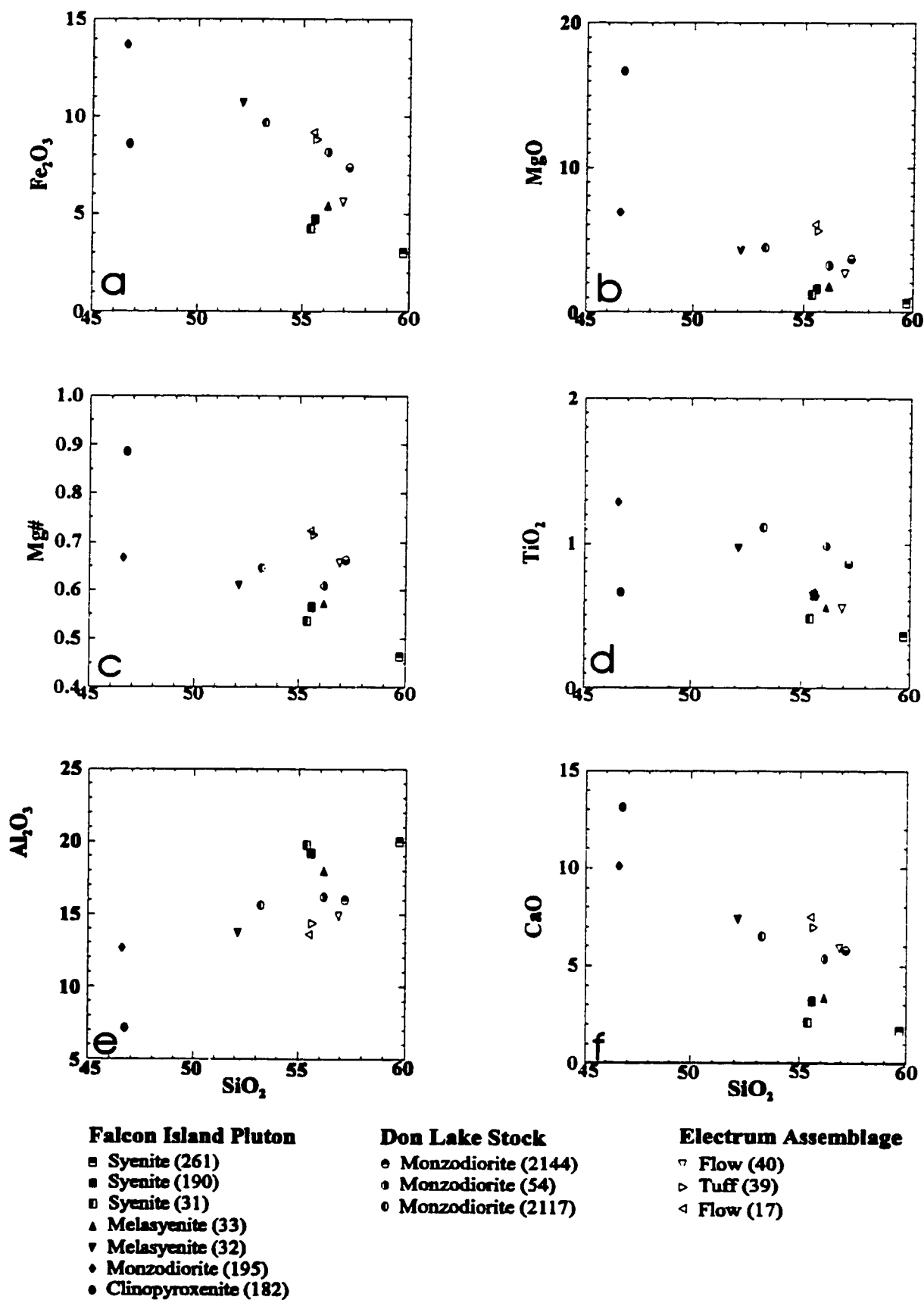


Fig. 4.2. Harker variation diagrams for the Lake of the Woods potassic suite.

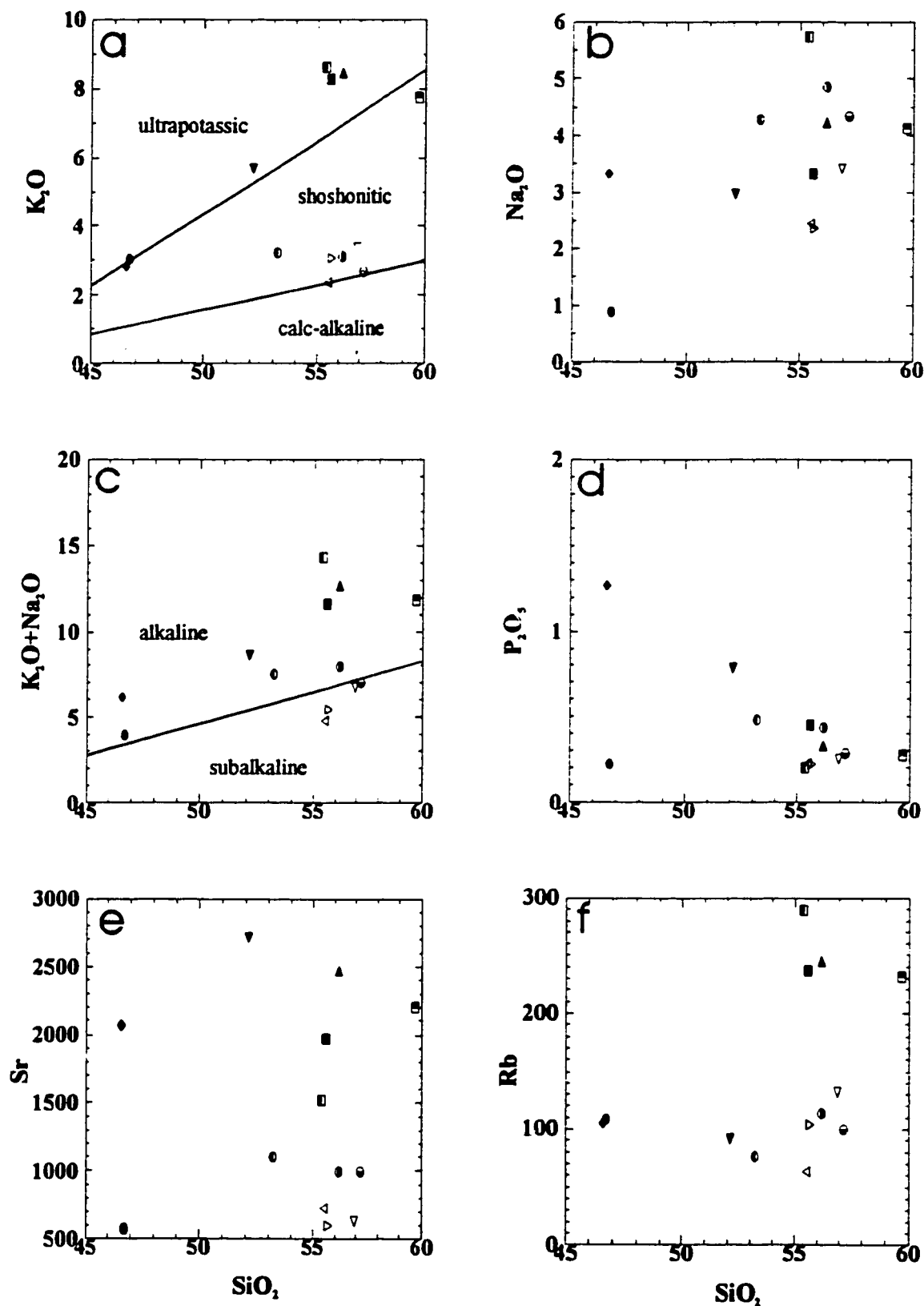


Fig 4.3. Harker variation diagrams for the Lake of the Woods potassic suite (symbols as on Fig. 4.2). Boundary between ultrapotassic and shoshonitic (Fig. 4.3a) after Wheller et al. (1987) and between shoshonitic and calc-alkaline from Peccerillo and Taylor (1976). Figure 4.3c boundary from Irvine and Barrager (1971).

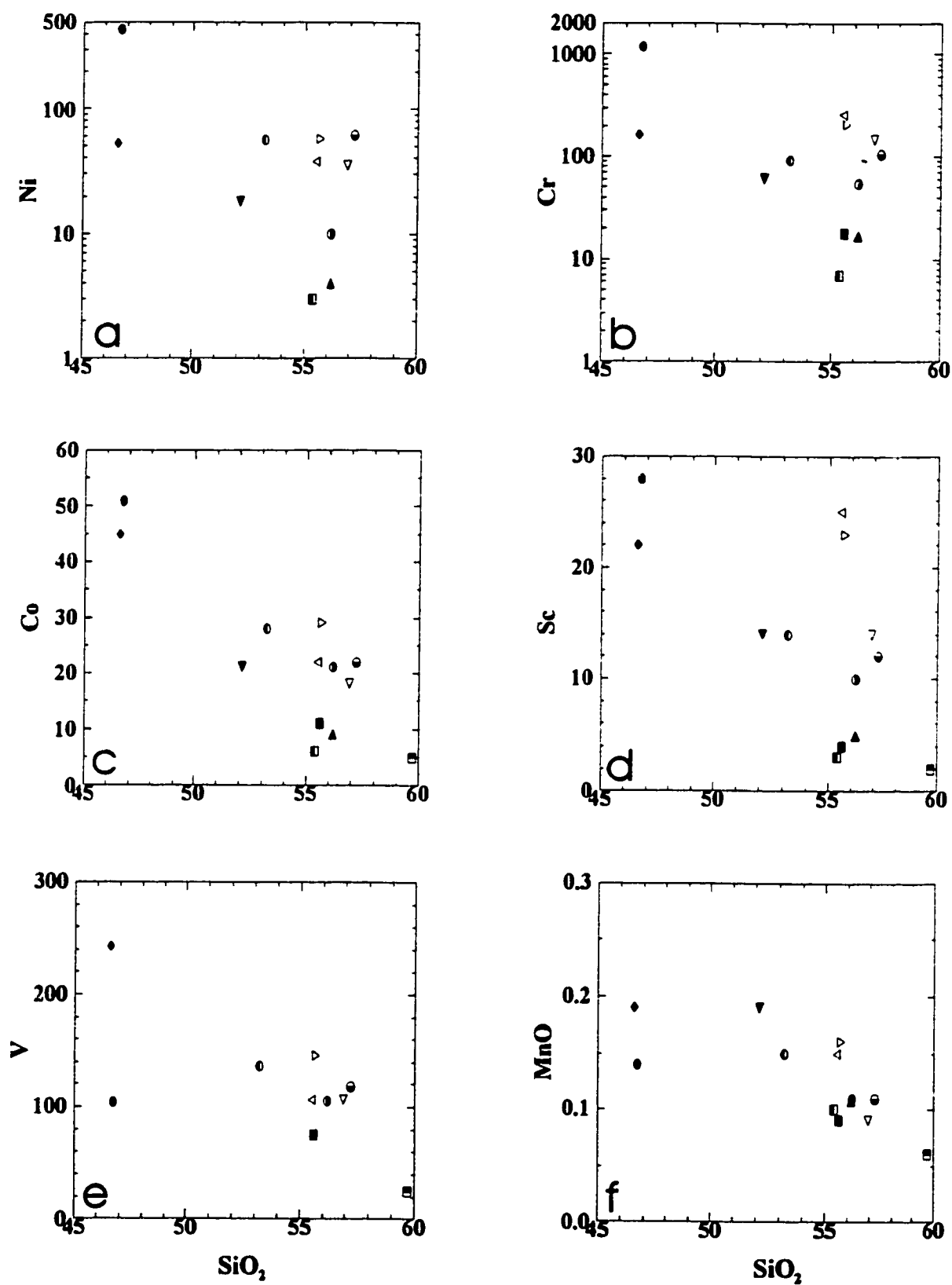


Fig. 4.4 Harker variation diagrams for the LWGB potassic suite (symbols as on Fig. 4.2).

clinopyroxenite, the Don Lake and the Electrum Assemblage samples deviate from this colinearity in a number of the plots. In particular, these units have higher MgO, CaO, Ni, Cr, Co and Sc, and lower Al_2O_3 , Sr and Rb than the other Falcon Island members for similar SiO_2 values. Rb content remains low and constant for the mafic end-members of the Falcon Island pluton, the Don Lake monzodiorite and the Electrum Assemblage shoshonite, but change to relatively high values (200-300 ppm) for the syenitic units of the pluton (Fig. 4.3f). P_2O_5 is negatively correlated with SiO_2 , except in the clinopyroxenite, which has very low values despite its low silica content (Fig. 4.3d). This indicates that apatite was not on the liquidus during crystallization of the clinopyroxenite, as is also evident by its absence in the rock.

4.3.1 Rare earth elements

Rare earth element (REE) patterns for the different potassic units of the LWGB are illustrated in Figure 4.5a & 4.5b. All suites exhibit similar patterns of elevated light REE (LREE) and depleted heavy REE (HREE) without any Eu depletions and an overall decrease in REE abundances for increasing silica content in individual samples of a particular suite. However, the suites have small distinctions from one another in patterns. At Falcon Island, the overall REE abundances decrease progressively from monzodiorite to melasyenite to syenite. Enrichment of the REE, coincident with increases in TiO_2 and P_2O_5 , in the mafic units of the pluton, suggests that clinopyroxene, titanite and apatite were being accumulated. Clinopyroxenite is an exception to this trend, in that while it is compositionally the most primitive unit, it has similar abundances to the syenitic units in middle REE (MREE) and HREE, but distinctively lower LREE. The

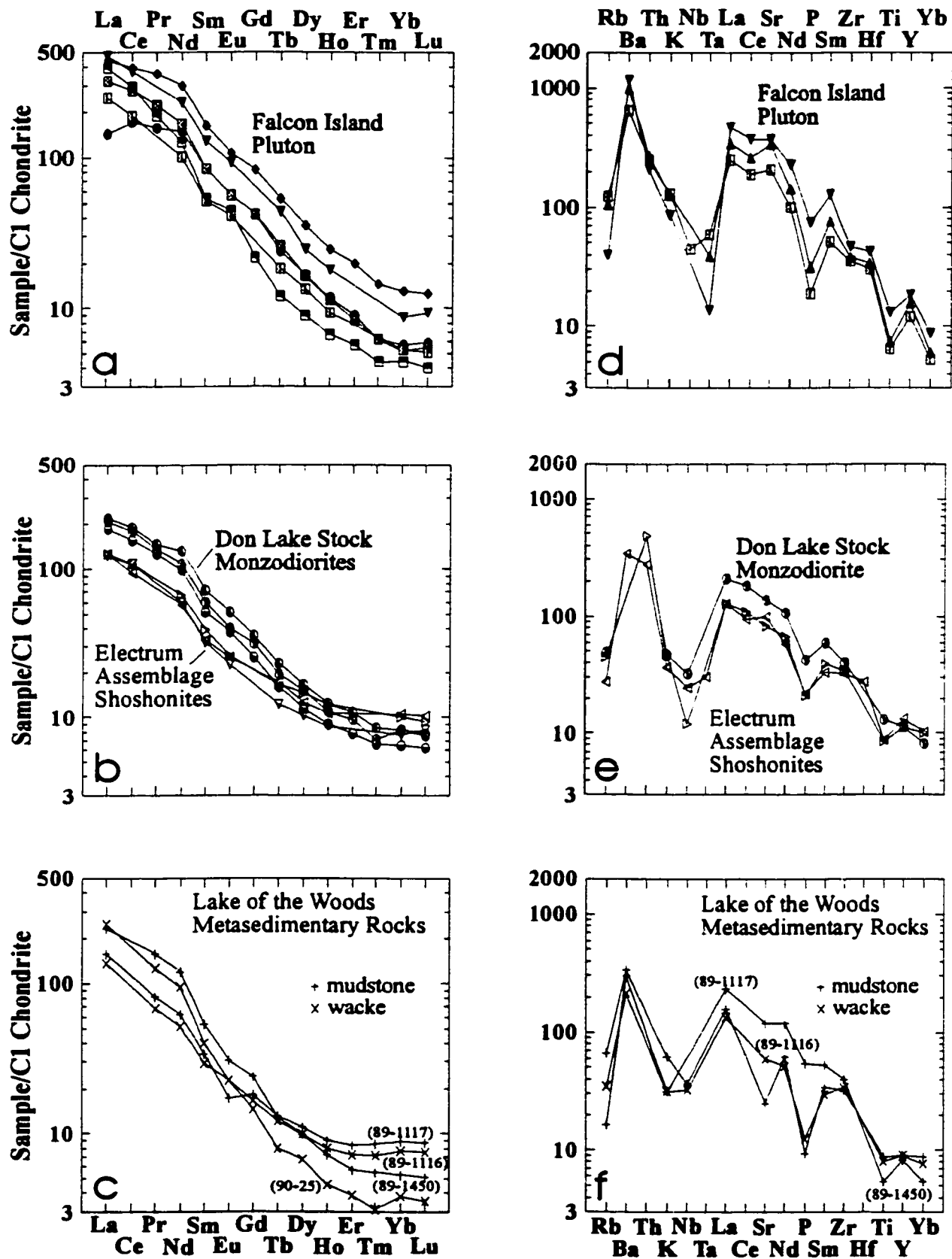


Fig. 4.5. Chondrite-normalized rare earth element and extended trace element diagrams for the LWGB potassic suite (symbols as on Fig. 4.2) and Upper Keewatin Assemblage metasedimentary rocks (symbols and sample numbers as illustrated).

concave downward pattern from La to Pr is suggestive of pyroxene accumulation. The silica saturated syenite is another exception, in that while it has the expected lowest MREE and HREE for the most evolved unit of the pluton, it has relatively enriched LREE. Crossing of the REE patterns indicates that both the least and the most evolved units can not be related to the other units of the Falcon Island pluton exclusively by Raleigh fractionation or differing degrees of equilibrium partial melting (Rollinson, 1993).

The Don Lake monzodiorite is very similar to the Falcon Island syenite in MREE and HREE abundances, but have slightly lower LREE (Fig. 4.5a & 4.5b). Compared with Falcon Island and Don lake, the Electrum Assemblage shoshonite has lower LREE and higher, concave upwards HREE patterns, suggesting more retention of garnet or amphibole in the residuum and variable HREE depletion in the source(s) (Stern and Hanson, 1991).

4.4 Petrogenesis

The chemical patterns of the LWGB potassic suite suggests petrogenesis is the result of the interplay of a variety of magmatic processes which could include: (1) crystal fractionation; (2) different degrees of partial melting of the same or different sources and (3) mixing of magmas, or mixing by contamination of the magma by wall rock assimilation. These processes are evaluated below to determine to what extent they may have influenced the overall petrogenesis of the suite. However, because there is considerable freedom in using these models in terms of variability of input parameters (e.g. composition of assimilants, partition coefficients, degree of mixing, etc.) it is possible to model with high precision, a geologically improbable petrogenesis. Consequently

geochemical models are used as a guide or adjunct with other geological and geochemical data in order to decipher the evolution of the suite.

4.4.1 Crystal fractionation

Pearce element ratio (PER) diagrams (Pearce, 1968) are X-Y plots of ratios of elements designed to avoid the closure problem that can distort trends on weight percent oxide-oxide diagrams. Such diagrams allow recognition of cogenetic rocks and permit the identification and quantification of mineral gains and losses from the system thus allowing discrimination among processes of magmatic differentiation such as crystallization, assimilation and magma mixing (Russell and Nicholls, 1988). The denominators of the ratios are common to both axes and are usually single elements that must be conserved (i.e. remain incompatible during the differentiation process or any subsequent alteration). The numerators may be single elements or complex algebraic combinations of elements (Ernst et al. 1988). Appropriately chosen combinations of these elements yield linear equations that isolate the variation in the abundance of specific minerals and allow testing of fractionation hypotheses. The best-fit lines of such diagrams have slopes equal to the stoichiometric ratio of the numerator elements of the ordinate/abscissa in the minerals responsible for the accumulation or fractionation. Note that all minerals are expressed as molar abundances. La was used as the conserved element (i.e. the denominator) in Figures 4.6a and 4.6b. La is a suitable element because it is incompatible for all of the observed rock forming minerals in melts of mafic to ultramafic composition (i.e., clinopyroxene, hornblende, biotite, alkali feldspar and plagioclase all have mineral-liquid partition coefficients (K_d) less than unity;

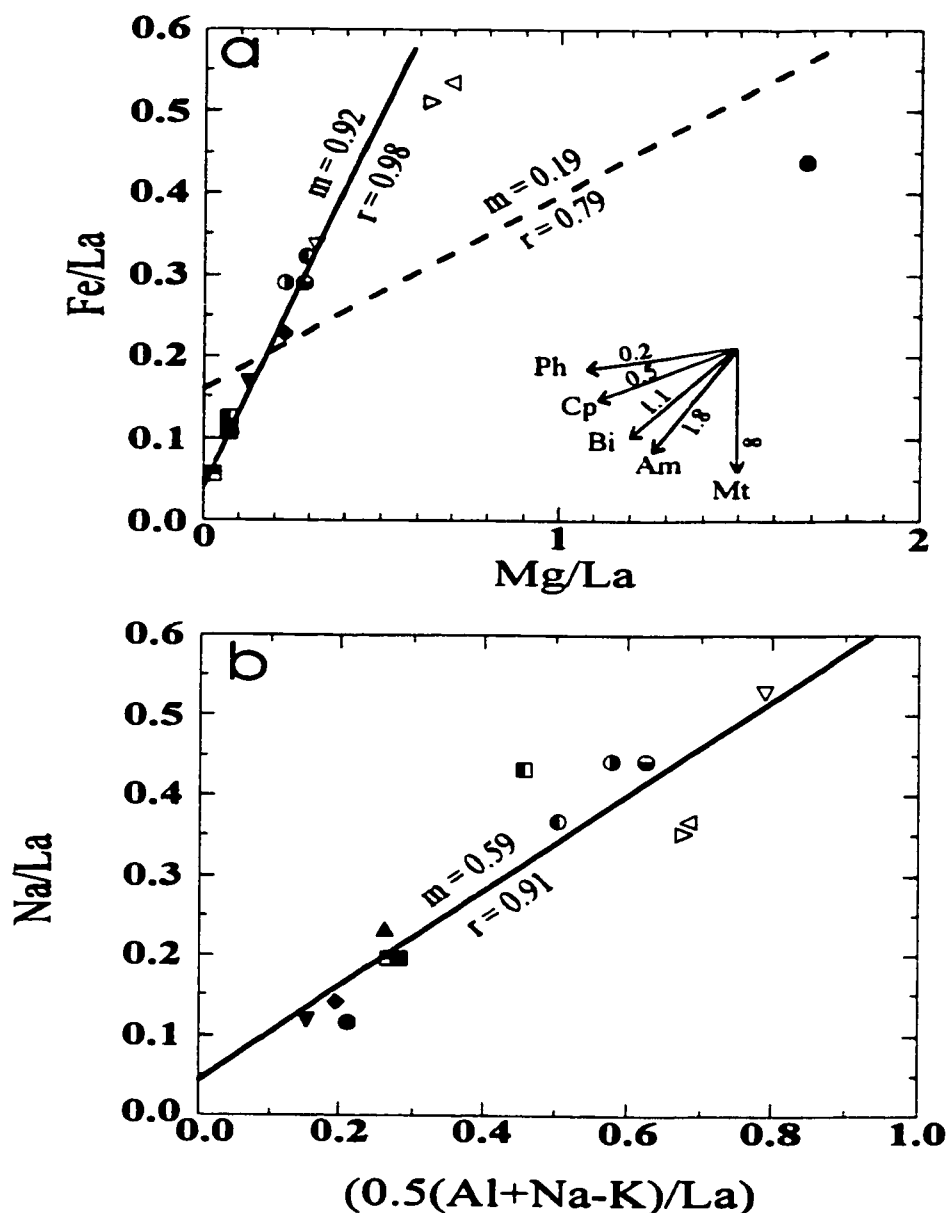


Fig. 4.6. LWGB potassic suite data plotted on Pearce element ratio (PER) diagrams (symbols as on Fig. 4.2). Fig. 4.6a tests for the effects of ferromagnesian mineral-controlled fractionation. The dashed line is a best fit line that includes all samples, while the solid line is a best fit line for the Falcon Island samples only (excluding clinopyroxenite), see text for explanation. M values represent slopes and r values the correlation coefficients, also shown are m values of fractionation vectors for average Falcon Island pluton phlogopite (Ph), clinopyroxene (Cp), biotite (Bi), amphibole (Am) and magnetite (Mt). Fig 4.6b tests the effects of feldspar fractionation. The solid line is a best fit line for all samples. The numerators, albite (Na) and anorthite plus albite $[0.5(\text{Al} + \text{Na} - \text{K})]$, have been constructed based on mineral stoichiometry which sum the contributions from each mineral.

Rollinson, 1993). La is also relatively immobile and occurs at high concentrations relative to its detection limits thus minimising the effects of analytical error.

Fractionation of the ferromagnesian minerals can be tested with a simple PER diagram of Fe/La versus Mg/La (Fig. 4.6a) having a slope equivalent to the stoichiometric ratio of Fe:Mg in the fractionated or accumulated minerals. The Falcon Island samples (excluding clinopyroxenite) and the Don Lake samples demonstrate a well correlated (correlation coefficient $r = 0.98$) best fit line suggesting a high probability of mafic mineral fractionation (or accumulation) and a common magma source for the two intrusions. Zoning and variable Fe/Mg ratios of the ferromagnesian phases in different units of the Falcon Island pluton (Chapter 3) complicates the interpretation of the straight-line relationship. An approximation, calculated on the basis of the average Fe/Mg ratios of the mafic minerals of the pluton is illustrated in the inset in Figure 4.6a. The slope of the best fit line (0.92) is significantly greater than that average Fe/Mg ratios of the Falcon Island clinopyroxenes (0.5), indicating that mafic minerals with higher average ratios, such as biotite (1.1), hornblende (1.8) or magnetite (∞) were also involved in the fractionation. The Falcon Island clinopyroxenite and Electrum Assemblage shoshonites were not included in the above correlation calculation as they appear to represent differing trends. Inclusion of the Falcon Island clinopyroxenite shifts the best fit line to a less well correlated best fit ($r = 0.79$) with a considerably lower slope value of 0.19 (dashed line on Fig. 4.6a) suggesting that the higher magnesian content of the clinopyroxenite's mafic minerals (i.e., clinopyroxene, $m = 0.18$; amphibole, $m = 0.26$, phlogopite, $m = 0.22$; see Chapter 3) were accumulated in an event distinctive from that which formed the remainder of the intrusion.

Thus the depletion evident in Fe, Mn, Ti, and V in the Falcon Island clinopyroxenite can not be explained by a simple Rayleigh fractionation model and may be consistent with early fractionation of oxides or possibly by variation in partition coefficients of the precipitated minerals which can vary extensively according to the temperature, pressure, oxygen activity and composition of the melt (Rollinson 1993). Considerable changes in temperature and oxygen fugacity are evident during the crystallization history of the Falcon Island complex. Estimates based on biotite microprobe data indicate crystallization temperatures declined from a maximum of about 1000° C at $f(\text{O}_2)$ of 10^{-8} for clinopyroxenite, to a minimum of 750° C at $f(\text{O}_2)$ of 10^{-15} for syenite (Chapter 3).

Figure 4.6b is a PER plot of Na/La versus $[0.5(\text{Al}+\text{Na}-\text{K})/\text{La}]$ which monitors the effect of plagioclase (anorthite plus albite) fractionation. The best fit line, suggesting a cogenetic relationship between the Falcon Island and Don Lake intrusions, is well correlated ($R = 0.96$) and has a slope of 0.78. The slope is a rough measure of the average anorthite content $[\text{An}/(\text{Ab}+\text{An})]$ of the plagioclase during crystallization (i.e., An_{22}). This value is an underestimate in units where plagioclase dominates over alkali feldspar because the slope-calculated plagioclase composition has albite components from both plagioclase and the alkali feldspar. Conversely, it will be an overestimate in rocks where alkali feldspar is the dominant feldspar. However, the estimate is in approximate agreement with the microprobe data from plagioclase in the Falcon Island monzodiorite which is normally-zoned from andesine cores to oligoclase rims (An_{32} to An_{20}) and syenite where the plagioclase is albite ($\text{An}_{0.7}$ to $\text{An}_{1.6}$) (Chapter 3).

At the Cléricy pluton in the Abitibi Subprovince, Laflèche et al. (1991) demonstrated the clinopyroxene accumulative nature of the ultrapotassic mafic syenites and clinopyroxene

accumulates by the constant FeO/MgO ratios accompanied by variable CaO/Al₂O₃, whereas the more evolved silica oversaturated nordmarkite samples had a markedly different trend of constant CaO/Al₂O₃ ratios accompanied by variation in FeO/MgO. A plot of CaO/Al₂O₃ vs. FeO/MgO (Fig 4.7) for the LWGB potassic suite indicates there are differences in trends from those observed at Cléricy. A steeply inclined slope for the mafic end-members of the Falcon Island pluton indicate highly variable CaO/Al₂O₃ ratios for small changes in FeO/MgO, whereas the evolved end-members demonstrate a subhorizontal trend. On the other hand, the Don Lake stock and Electrum Assemblage samples have a collinear trend with an intermediate slope indicating variation in both ratios. Similar to at Cléricy, the different trends can be considered as vectors that point to a parental composition, which in the LWGB is, most similar to the Falcon Island melasyenite and the Don Lake monzodiotite.

4.4.2 Partial melting

A plot of Rb vs. Cr (Fig. 4.8) demonstrates distinctive trends consistent with more than one magmatic processes in the genesis of the LWGB potassic suite. In the melasyenitic to syenitic portion of the Falcon Island pluton the trend is one of decreasing Cr with increases in Rb content, while in the mafic portion of the pluton and in the Don Lake stock and Electrum Assemblage, Rb is relatively constant for changes in Cr content of 10-15 times magnitude. Increasing Rb contents in the evolved members of the Falcon Island trend, strongly suggests that alkali feldspar and/or biotite was involved in their fractionation. Feldspar is considered more likely however, because the evolved units of the pluton contain a higher proportion of alkali feldspar than biotite (Chap.3).

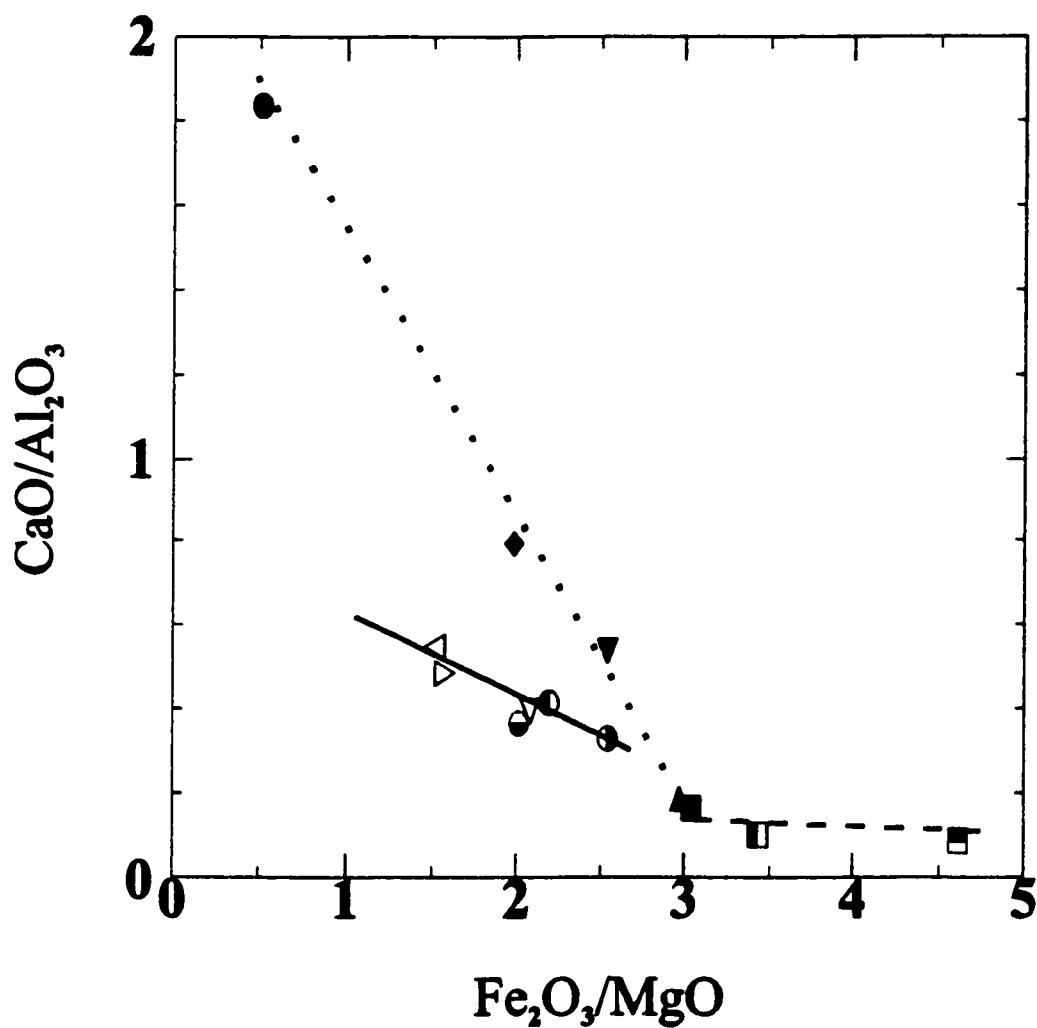


Fig. 4.7. $\text{CaO}/\text{Al}_2\text{O}_3$ vs. $\text{Fe}_2\text{O}_3/\text{MgO}$ plot with vectors indicating magmatic fractionation trends in the potassic suite of the Lake of the Woods. The dotted and dashed lines represent trends for mafic and felsic end-members of the Falcon Island pluton respectively and the solid line for the Don Lake and Electrum Assemblage samples (symbols as on Fig. 4.2).

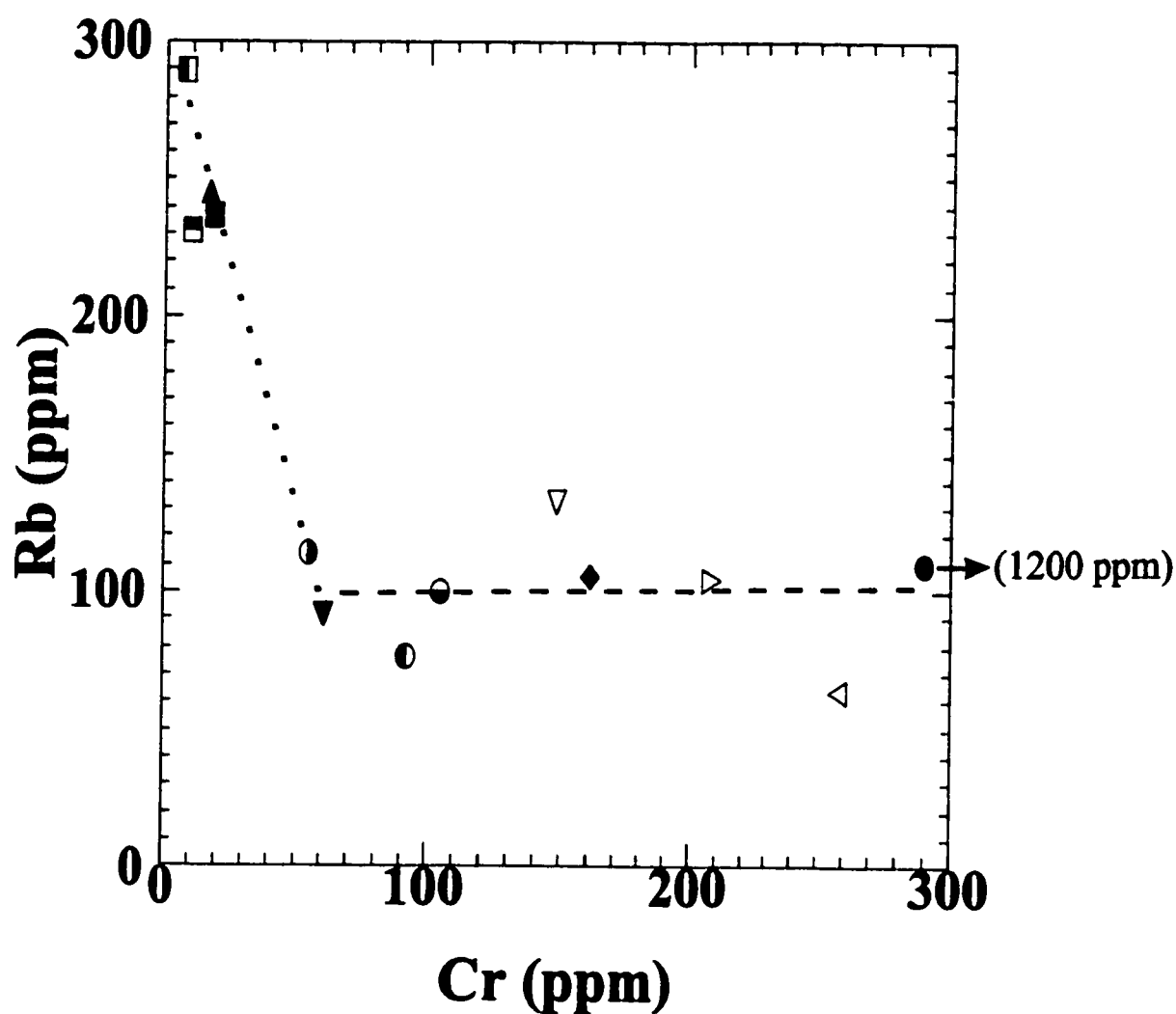


Fig. 4.8. Cr vs. Rb for the Lake of the Woods potassic suite (symbols as on Fig. 4.2). The dotted line illustrates a distinctive vector for the evolved parts of the Falcon Island pluton, while the dashed line a vector for the cumulative parts of the pluton, the Don Lake and Electrum Assemblage samples.

Figure 4.8 indicates that variation in the degrees of partial melting of the source did not influence the chemical trends in the mafic portions of the suite because for sources which leave pyroxene or garnet in the residue (as is evident in the REE patterns of the LWGB potassic suite), Cr is strongly compatible ($D \gg 1$) and Rb is incompatible ($D \ll 1$), and thus Cr will exhibit much less variation than Rb for geologically reasonable amounts of partial melting (i.e. $< 50\%$) (Stern and Hanson, 1991).

4.4.3 Magma mixing

Langmuir et al (1978) developed a general mixing equation that can be used to test if mixing is a viable geological process, and if so, place constraints on end-member compositions and allow distinction between mixing of magmas or sources. Simple mixing produces straight lines on element-element plots and straight or hyperbolic lines on ratio or isotopic plots but can be complicated by igneous processes such as partial melting of mixed sources, differentiation, contamination, or alteration processes. When mixing is involved, the difference in behaviour between incompatible ($D \ll 1$) and slightly compatible elements ($D \cong 1$) should allow a determination of whether mixing occurred in the source before separation of magmas, or between magmas after separation from source. This distinction is possible because partial melting and differentiation offset the ratios of slightly compatible elements to those of incompatible elements, but does not affect the ratios of incompatible elements. Therefore ratio-ratio plots involving exclusively incompatible elements will show straight line or hyperbolic mixing relations for either magma and source mixing. Compatible/incompatible ratio-ratio plots only show mixing trends if

the mixing is between magmas. The ratio-ratio plots of the LWGB potassic suite suggest mixing of magmas is more likely than mixing of sources. This is indicated by plots of incompatible ratios such as Rb/Ba vs. K/Ba and Rb/K vs. Ba/K (Figs. 4.9a & 4.9b), and the compatible/incompatible ratio plots such as La/Th vs. Nd/Th, and La/Rb vs. Sr/Rb (Figs. 4.9c & 4.9d), all of which show linear relationship with high correlation coefficients.

A number of model-derived trends for fractional crystallization (FC), assimilation fractional crystallization (AFC) and partial melting (PM) are illustrated on Figure 4.9. In Figure 4.9a, an AFC trend for relatively low amounts of assimilation ($r = 0.2$) is modelled as a function of the variation of the incompatible ratios in the fraction of remaining liquid in 20 steps (i.e. from 5% to 95% fractionated or AFC = 0.05 to 0.95 on Fig 4.9a). The AFC trend was calculated utilizing the Falcon Island melasyenite (#32) as a hypothetical parental magma and average continental crust as the assimilant (specific parameters are as indicated in the figure caption). In general, the trend is quite similar to the observed trend in the LWGB potassic suite but extends to considerably more evolved ratios for over 50% fractionation. Modelling of Rayleigh fractionation using the same parental melasyenite and proportions of fractionating minerals yields a generally similar trend, but extends to ratios which are an order of magnitude higher (not shown on Fig. 4.9a) than for AFC. A Rayleigh fractional crystallisation trend modelled using the same parameters is displayed on Figure 4.9b. The trend is hyperbolic (as is the undisplayed trend of AFC) and is thus somewhat different to the observed linear trend of the LWGB potassic suite.

Fractional crystallization and AFC model-derived trends are also shown on the compatible/incompatible trace element ratio-ratio plots (Figs. 4.9c & 4.9d) and are quite different from the observed trends. Model-derived trends for partial melting (PM1% to 20% on figs. 4.9c

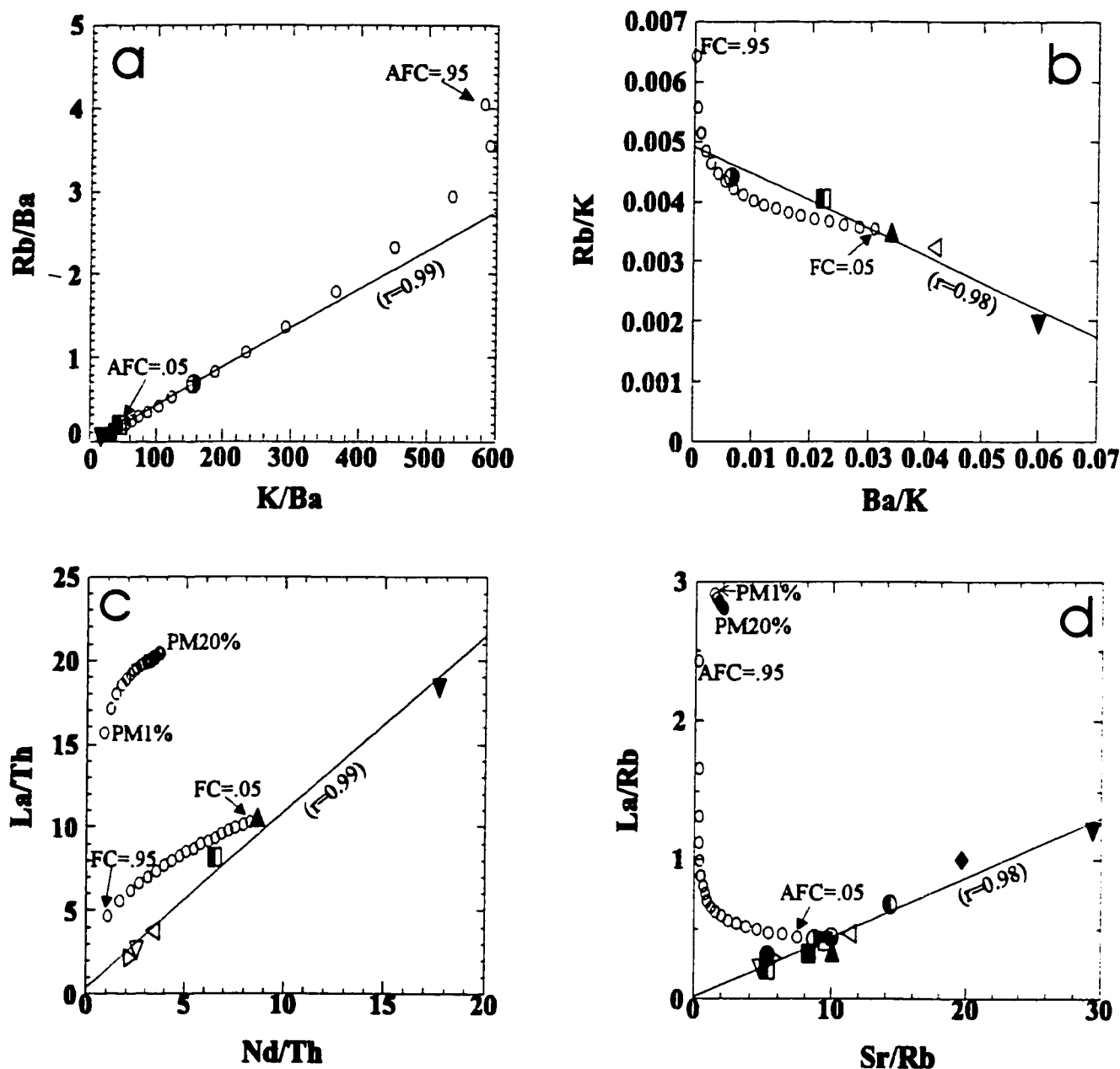


Fig. 4.9. Ratio-ratio plots of samples from the LWGB potassic suite with best fit lines and correlation coefficients (r) (symbols as on Fig. 4.2). Model derived trends as follows: Figure 4.9a assimilation fractional crystallization (AFC) model calculated using equation of DePaolo (1981) and Powell (1984); using sample # 32 (Table 1) as the parental composition, the mineral distribution coefficients of Rollinson (1993); fractionation of 20% clinopyroxene, 20% biotite, 20% plagioclase, 20% alkali feldspar, 15% amphibole and 5% magnetite; the composition of average continental crust as the assimilant (Rollinson 1993) and rate of assimilation $r = 0.2$. Figure 4.9b, fractionation (FC) model calculated using the equation for Rayleigh fractionation (Rollinson 1993), sample # 32 as parental composition and the same proportions of fractionating minerals as above. Figure 4.9c, FC model parameters as above, partial melting (PM) model calculated using the equilibrium fractional melting equation (Rollinson 1993), the primitive mantle composition of Sun and McDonough (1989) with a garnet lherzolite assemblage consisting of 55% olivine, 25% orthopyroxene, 15% clinopyroxene and 5% garnet. Figure 4.9d as above.

& 4.9d) of a garnet lherzolite source of primitive mantle composition are also very different from the observed trend in the suite. As expected none of the modelled trends suggest a unique solution for all the ratio-ratio plots, but are consistent with petrogenesis of the suite by a combination of fractionation (with or without small degrees of crustal assimilation) and magma mixing.

4.4.4 Crustal contamination

Mixing can also be the result of crustal contamination. The crystallization products of plutons, or the extrusive products of magma chambers, contaminated by melting and assimilation of material from the surrounding country rock can be tested for assimilation fractional crystallization. The country rock surrounding the Falcon Island pluton includes mafic and ultramafic metavolcanic rocks which have relatively low and flat REE abundances at around 10 times chondrite (Fig. 2.5). Beside the compositional dissimilarity, thermodynamic consideration of the temperatures required to melt ultramafic and mafic rocks excludes them from being contaminants. The metasedimentary country rocks are however very similar in trace element patterns to the units of the pluton (Fig 4.5). This chemical similarity, in conjunction with an evolutionary trend towards increasing Al content and silica saturation in the pluton, is consistent with contamination.

Radiogenic isotopes are a very useful tool for determining if crustal contamination is a viable process and have been used in documenting the crustal contamination of alkaline magmas (e.g., Laflèche et al., 1991; Chen et al., 1994; Landoll and Foland, 1996). Archean metasedimentary rocks are typically derived from ancient crustal sources and thus have more

enriched Nd isotopes than rocks derived from depleted mantle such as tholeiitic, komatiitic and potassic magmas. The metavolcanic suites of the LWGB have broadly similar depleted ϵ_{Nd} values (i.e. +1.2 to +2.1) to those of the pluton (avg. +2.2), but wacke in the metasedimentary unit intruded by the pluton has a much lower ϵ_{Nd} of -0.8 (Chapter 5). Thus assimilation of the metavolcanic rocks would not change the ϵ_{Nd} values but assimilation of significant amounts of the metasedimentary rocks would result in lowering the ϵ_{Nd} values in progressively fractionated units of the pluton. Figure 4.10 shows that ϵ_{Nd} values in the Falcon Island pluton progressively increase with fractionation and thus mixing by crustal assimilation is precluded. The Electrum Assemblage volcanics are a more plausible candidate for having experienced crustal contamination as they have considerably lower ϵ_{Nd} values of +0.4 for the shoshonite (and -0.4 for calc-alkaline rhyolite; Table 5.1) and plot between the values for the pluton and the wacke (Fig. 4.10).

4.5 Discussion and conclusions

Late-tectonic potassic suite of the LWGB include silica-undersaturated ultrapotassic, silica-saturated monzodioritic and silica-oversaturated shoshonitic units. These mantle-derived magmas have been collectively identified as a “sanukitoid” suite in the Superior Province (Stern et al., 1989). The rocks of the Falcon Island pluton are ultrapotassic on the basis of $K_2O > 3$ wt %, $MgO > 3$ wt % and $K_2O/Na_2O > 2$ (Foley et al., 1987), while Don Lake monzodiorite and the Electrum Assemblage shoshonite samples are lower in K_2O , contain normative quartz, and thus belong to the shonshinite clan (Fig 4.3a). The similar geochemical characteristics, including

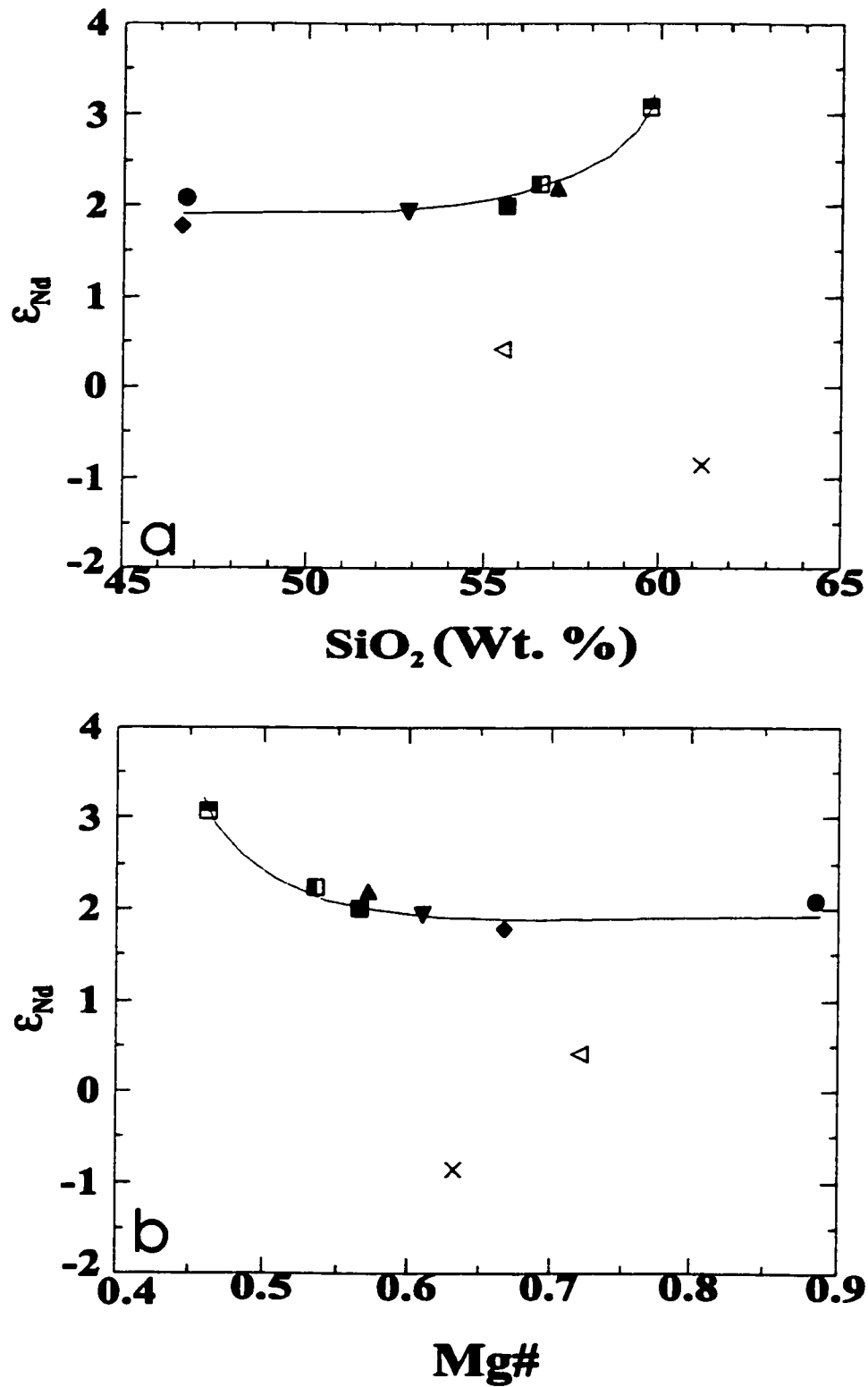


Fig. 4.10. Epsilon Nd vs. (a) SiO_2 and (b) Mg# (mol. Wt % $MgO/MgO + Fe_2O_{3(T)}$) (symbols as on Figs. 4.2 & 4.5).

elevated Al, K, Sr, Rb, Ba, LREE, depleted HREE and HFSE of the ultrapotassic, monzodioritic and shoshonitic units are attributed to a late subduction event which occurred from about 2695 to 2700 Ma across the western Wabigoon Subprovince (Ayer and Davis, 1997).

Ultrapotassic rocks are subdivided into three principal groups according to geochemistry and tectonic setting (Foley et al., 1987). Group-I ultrapotassic rocks consist of lamproite and other alkaline rocks in continental environments. Group-II rocks are represented by kamafugite and occur in rift zones (e.g. the East African rift). Group III rocks include the leucitites in island-arc environments and are characterised by relatively higher Al, Ca and K and lower Mg and high field strength elements (HFSE) contents with pronounced depletion in Nb, Ta, P and Ti.

Modern potassic igneous rocks occur in a wide range of tectonic settings that can be broadly subdivided into subduction-related and within-plate associations. The three subduction-related tectonic associations are oceanic arcs such as the Mariana Islands and Fiji, continental arcs such as the Andes and the Roman Province and post-collisional arcs such as the Alps and Papua New Guinea (Müller et al. 1992). Müller and Groves (1997) have devised a hierarchical set of discrimination diagrams to classify the tectonic setting of potassic rocks based on a world-wide lithogeochemical database of over 2200 potassic rocks. The classification is based on simple ratios of immobile elements revealed by discriminant analysis as contributing most effectively to separation of the potassic igneous rocks from the various tectonic settings. Figure 4-11a demonstrates the LWGB potassic rocks occur in the field of postcollision arcs and continental arcs, and are further subdivided in Figure 4-11b to an geochemical affinity with modern continental arcs.

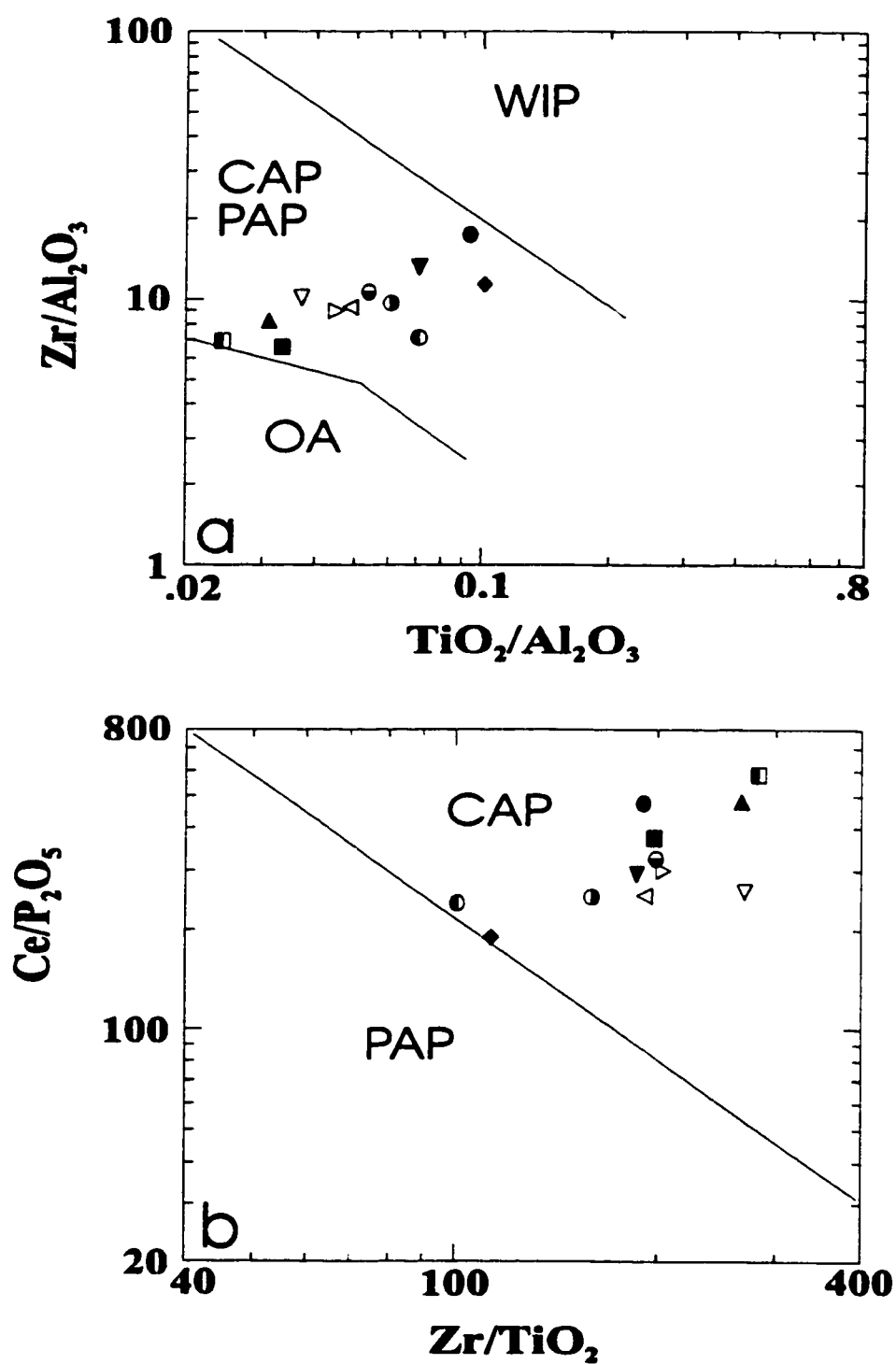


Fig. 4.11. LWGB potassic suite plotted on tectonic discrimination diagrams of Müller and Groves (1997). WIP - within plate, CAP - continental arc, PAP - postcollisional arc and OA - oceanic arc (symbols as on Fig. 4.2).

The petrogenetic processes producing the various types of modern potassic igneous rocks are still debated (Peccerillo 1992). However, there is general consensus that potassic magmas cannot be derived by melting of normal mantle peridotite, but require heterogeneous mantle sources which have been metasomatically enriched in LILE and LREE (Müller and Groves 1997). Enrichment of these elements, volatiles and halogens are mainly concentrated in hydrous minerals such as phlogopite and amphibole (Aoki et al. 1981). Thus those portions of the mantle rich in phlogopite are considered to be important sources in the genesis of potassic melts (Foley 1992). Shoshonites from modern arc environments are considered to be generated primarily in the mantle wedge overlying the Benioff zone (Gill, 1981). Meen (1987) considers that potassic melts form by low degrees of partial melting of upper mantle lherzolite that has been metasomatically enriched in the LILE and LREE under hydrous conditions in a relatively low heat-flow environment. In arc settings, the partial melting may be achieved by modifications of mantle geotherms in proximity to the cold subducted slab, while the metasomatic enrichment may be achieved by overprinting and veining of the mantle wedge by either volatile and LILE-enriched fluids and/or by actual alkalic, low temperature, LILE and LREE-enriched partial melts derived during dehydration of the subducted oceanic slab (Müller and Groves 1997).

Documented occurrences of shoshonitic volcanic rocks are increasingly common throughout the Superior Subprovince as the volcanic components of the Neoarchean successor basins are studied in greater detail. Potassic volcanic rocks of shoshonitic affinity have been reported in the central part of the western Wabigoon Subprovince (Teal, 1979) and the Lake Shebandowan area of the Wawa Subprovince (Shegelski, 1980). Shoshonites of striking geochemical similarity to those of the Electrum assemblage also occur in the Oxford Lake area of

northeastern Manitoba (Brooks et al., 1981) and the Chibougamau area of Northwestern Quebec (Dostal and Mueller, 1992). Shoshonitic volcanic rocks ranging to more undersaturated alkaline compositions also occur in the Timiskaming Group in northeastern Ontario (Ujike, 1985; Ben Othman et al., 1990).

Stern et al. (1989) have defined the late-tectonic monzodioritic rocks in the Superior Province as a sanukitoid-like suite based on chemical characteristics such as 55-60 wt % SiO_2 , $\text{Mg\#} > 0.6$, Ni & Cr both > 100 ppm, $\text{Na}_2\text{O} + \text{K}_2\text{O} > 6$ wt %, Sr & Ba both 600-1800 ppm and elevated light rare earth elements (LREE) without Eu anomalies. Members of the Superior Province sanukitoid suite exhibit pronounced depletions in the HFSE elements (e.g. Brooks et al., 1982; Ujike, 1985; Laflèche et al., 1991, Stern and Hanson, 1991; Dostal and Mueller, 1992; Rowins et al., 1993) and are thus very similar to the K-rich alkaline, shoshonitic and calc-alkaline magmatic association of the Sunda arc in Indonesian (Wheller et al., 1987). The potassic suite of the LWGB fit these criteria although the more evolved end members of the Falcon Island suite are less primitive than the typical sanukitoids having relatively low Mg# and Ni, Cr and Co abundances (Figs. 4.2 & 4.4). Shoshonitic to ultrapotassic magmatism is attributed to late stage subduction events in the Neoproterozoic Ottawa orogeny (ca. 1080 Ma) of the Grenville Province (Corriveau et al., 1993) and the Paleozoic Caledonian orogeny (ca. 410 Ma) of NW Scotland (Thompson and Fowler, 1986) and thus spans over 2.7 Ga of Earth history

Geochemical patterns within the LWGB potassic suite suggest crystal fractionation was an important petrogenetic processes, while variation in the degrees of partial melting appears to be of less significance (Figs. 4.6, 4.7, 4.8 & 4.9). Clinopyroxene, biotite, magnetite and possibly amphibole were accumulated in the mafic units and fractionated along with feldspar in the evolved

units. The differing trends on Figures 4.7 and 4.8 provide vectors pointing to a common parental composition which is similar to that of the Don Lake stock monzodiorite and the Falcon Island melasyenite. The unfractionated, relatively uniform composition of Don Lake monzodiorite, in conjunction with its close proximity to the Falcon Island pluton (Fig. 4.1), supports a common genesis and is consistent with the parental monzodiorite composition suggested from a number of studies of Superior Province sanukitoids (e.g., Shirey and Hanson, 1984; Sutcliffe et al., 1990; Stern and Hanson, 1991; Laflèche et al., 1991; Rowins et al., 1993).

Compatible and incompatible ratio plots (Fig. 4.9) suggest that magma mixing rather than source mixing better explains the chemical trends. In addition, the variation of Nd isotopes with the compositional evolution at Falcon Island pluton indicates mixing between end-members with variably depleted mantle signatures, and did not involve assimilation of the surrounding metasedimentary rocks (Fig. 4.10). However, the considerably lower ϵ_{Nd} values for the Electrum Assemblage shoshonite (and calc-alkaline rhyolite; Table 5.1) indicates mixing by crustal assimilation may have occurred in the silica oversaturated end-members. This is similar to the conclusion reached for the silica saturated end-members of a number of other alkaline complexes (e.g. Laflèche et al., 1991; Chen et al., 1994; Landoll and Foland, 1996).

The post 2.70 Ga potassic suite of the LWGB have broadly similar trace element patterns to the Upper Keewatin Assemblage calc-alkaline suite (ca. 2.72 Ga), both of which are attributed to small degrees of partial melting of depleted mantle sources. However, this calc-alkaline suite has somewhat less fractionated REE, lower LILE, and less pronounced HFSE depletion (Chapter 2), and more homogeneous, depleted mantle Nd isotopic signatures (Chapter 5). Upper Keewatin Assemblage turbidite (ca. 2.71 Ga) samples have very similar LILE, REE, and HFSE depletions

to those of the potassic suite (Fig. 4.5). The geochemical patterns of Abitibi Subprovince turbidite and alluvial fan fluvial sandstone samples (ca. 2675-2690 Ma), which are very similar to those of the Upper Keewatin turbidites, have been attributed to bimodal mixing between end-members averaging about 75% felsic, and 25% mafic and ultramafic detritus (Feng and Kerrich, 1990; Born, 1995). Archean sedimentary rocks typically have primitive compositions consisting of higher Mg#s, Ni and Cr, and more fractionated REE without Eu depletion attributed by McLennan and Taylor (1991) to a higher proportion of mantle-derived rocks in the source, than the dominantly intracrustal sources evident in the geochemistry of post Archean sediments.

Chemical similarity between samples of Quetico Subprovince turbidite (ca. 2.69 Ga) and Superior Province sanukitoid have been attributed to a source terrain with a large component of monzodioritic rocks (Zaleski, 1997). However, this hypothesis has a volume problem as the estimated amount of sanukitoids is only about 5% (Stern et al., 1989), considerably less than the estimated 25-30% of the southern Superior Province underlain by the metasedimentary subprovinces. In the LWGB, the potassic suite rocks are 10-15 million years younger than the Upper Keewatin Assemblage turbidites and thus could not represent a provenance source for the sediments. In addition, the Nd isotope data and U-Pb zircon ages indicate the Upper Kewatin Assemblage sediments were derived by depositional mixing of locally-derived greenstone belt sources and an ancient crustal source (Chapters 2 & 5).

Nd, Pb and Sr isotopic data from a variety of igneous units across the southern Superior Province indicates a change in the character of the mantle around 2.7 Ga (Chapter 5). Pre 2.7 Ga, mantle-derived, igneous rocks have relatively homogeneous isotopes, while the post 2.7 Ga mantle-derived igneous rocks have more variable isotope ratios (Chapter 5). The post 2.7 Ga

mantle rocks range from depleted isotopic signatures evident at Falcon Island, the Otto Stock, the Clèricy pluton ultrapotassic units and the carbonatite dikes at Lac Shortt and Dolodau, to enriched signatures evident in the Electrum Assemblage shoshonites, the Clèricy nordmarkites, the Roaring River monzodioritic complex and the Sturgeon Lake and Poohbah Lake alkaline complexes (Stern and Hansen, 1991; Laflèche et al., 1991; Tilton and Bell 1994; Chapter 5). The post 2.7 Ga data suggests mixing between depleted mantle and an enriched mantle end-member component than may be analogous to present day EMII. The depleted component is similar to the pre 2.7 Ga “greenstone belt” mantle and the enriched component having characteristics more akin to the “ancient crustal” signatures of the Winnipeg River, Opatica and Pontiac subprovinces. This type of heterogeneity can be explained by mantle which has been variably metasomatized by fluids derived by melting of sediments with a component of ancient crustal provenance. The sediments were probably on the upper part of a oceanic crustal slab subducted beneath the post 2.7 Ga crust as a result of Neoarchean collisional tectonic processes similar to those of island arcs and continental arcs in the Phanerozoic.

U-Pb geochronological studies of the late-tectonic potassic rocks and their associated sedimentary successor basins suggests that the magmatism became younger by 5-10 M.y. increments towards the southern margin of the Superior Province. For example, U-Pb zircon geochronology of Electrum Assemblage and associated potassic plutons in the LWGB and a number of other localities across the western Wabigoon Subprovince indicate ages from 2695 to 2700 Ma (Blackburn et al., 1991; Chapter 2). In the Wawa Subprovince, late-tectonic sedimentation and potassic volcanism and plutonism occurred around 2685 to 2690 Ma (Corfu and Sage, 1992; Corfu and Stott, in press), while further to the south in the Abitibi Subprovince,

Timiskaming Group sedimentation and alkaline magmatism ranges from 2675 to 2680 Ma (Corfu, 1993). This southerly progression of late-tectonic subduction-related potassic magmatism over a 15 million year interval suggests that the southern Superior Province experienced a late continental arc cratonization event possibly associated with terrain accretion to the south.

Chapter 5

Nd and Pb Isotopes from the Lake of the Wood Greenstone Belt, Northwestern Ontario: Implications for Mantle Evolution and the Formation of Crust in the Southern Superior Province.

5.1 Introduction

Similar to the approach that has led to the recognition of modern day mantle heterogeneities in terms of the radiogenic Sr, Nd, and Pb isotope compositions (e.g. Zindler and Hart, 1986), the isotopic characteristics of the Archean mantle can be evaluated by analysis of mantle-derived igneous rocks. Temporal and spatial variations in mantle composition permit a better understanding of crust-mantle interaction and aid in the characterisation of different geodynamic environments. Our data from the Neoproterozoic rocks of the Lake of the Woods greenstone belt (LWGB) confirm isotopic homogeneity in the pre 2.7 Ga mantle and heterogeneities in the post 2.7 Ga mantle, patterns which are widespread across the southern Superior Province. However, interpretation of the Archean results is more complex because the rocks are typically more altered and metamorphosed and it is therefore more difficult to evaluate the extent to which the isotopic systems have remained closed since the time of crystallization (Machado et al. 1986). Additionally, the Superior Province isotopic reservoirs were shorter lived

than modern day mantle reservoirs and thus because of less time for radiogenic growth, the fractionation trends are less pronounced and closer to the margin of error of the measurements.

Radiogenic isotopes are particularly useful as petrogenetic tracers when the composition of potential end-members are well characterised and when a precise time frame is available. The Lake of the Woods greenstone belt (LWGB) has a well established stratigraphy and U-Pb geochronology has provided a highly precise framework for the timing of geological events. In addition, the spectrum of geochemical compositions and the petrogenesis of the igneous suites has been studied in detail (Chapters 2, 3, & 4). Thus a Nd and Pb isotopic study of this belt provides additional insight into the characteristics of the Neoarchean crust and mantle and the basis for comparisons of crust-mantle reservoirs across the southern Superior Province.

Previous studies of Sr, Nd and Pb isotopes from mafic-ultramafic rocks and volcanic associated sulphide minerals from the Abitibi and Wabigoon subprovinces have shown the existence of isotopically homogeneous pre 2.7 Ga depleted Neoarchean mantle (i.e. 2.75 to 2.70 Ga) across the southern Superior Province (Tilton, 1983; Morrison et al., 1985; Shirey and Hanson, 1986; Machado et al., 1986; Dupré and Arndt, 1990; Vervoort et al., 1994; Carignan et al., 1995). A greater degree of isotopic heterogeneity is evident in the post 2.7 Ga mantle-derived shoshonite, monzodiorite, syenite and carbonatite suites (Shirey and Hanson, 1986; Tilton and Kwon, 1990; Stern and Hanson, 1991; Tilton and Bell, 1994, Chapter 5). Plutonic and gneissic rocks having isotope data indicating derivation from reservoirs with long-term radiogenic enrichment are documented from predominantly granitic and metasedimentary terrains such as the Winnipeg River, Opatica and Pontiac subprovinces (Gower et al., 1983; Corfu, 1988; Beakhouse and McNutt, 1991; Carignan et al., 1993).

5.2 Regional setting

The Lake of the Woods greenstone belt (LWGB) is situated along the northwestern margin of the Wabigoon Subprovince, a northeast-trending metavolcanic-plutonic belt about 150 km wide exposed over a length of about 900 km in the Superior Province. The subprovince is a collage of metamorphosed, supracrustal assemblages intruded by granitic batholiths.

Paleoarchean platformal supracrustal rocks and granitic bodies representing a 2.9-3.0 Ga microcontinent divides the Neoarchean (2.7 Ga) arc-dominated supracrustal rocks of the eastern and western parts of the subprovince (Blackburn et al., 1991). The subprovince is bounded to the north by the Winnipeg River and English River Subprovinces, and to the south, by the Quetico Subprovince (Fig 2.1).

Extensive detailed and synoptic geological mapping, geochemistry and geochronological studies by the Ontario Geological Survey have resulted in subdivision of the LWGB into three supracrustal assemblages (Fig. 5.1): (1) the Lower Keewatin Assemblage, (2) the Upper Keewatin Assemblage and (3) the Electrum Assemblage.

The Lower Keewatin Assemblage consists of geographically isolated mafic volcanic groups correlated on the basis of stratigraphy, lithological and geochemical similarities, and age. Lithogeochemistry of the tholeiitic basalts and rare komatiites of this assemblage indicate flat rare earth element (REE) patterns (Fig. 2.5a) suggestive of derivation from undepleted mantle and zircons from the upper part of the Bigstone Bay group yield an upper age limit of 2737.9 ± 2.0 Ma (Chapter 2). The Upper Keewatin Assemblage is the most areally extensive assemblage within the LWGB. It comprises a diversity of stratigraphically and structurally interdigitated

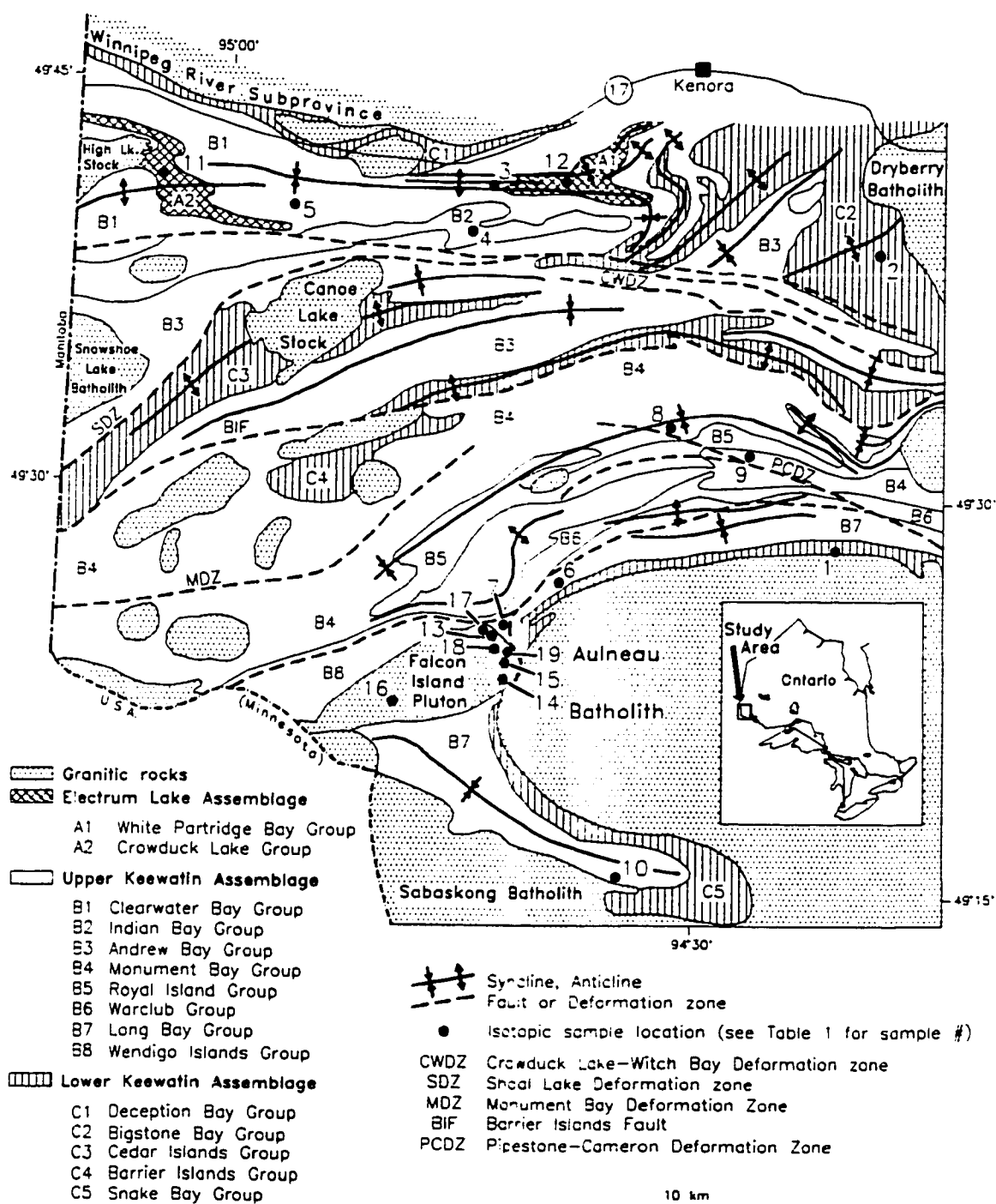


Fig. 5.1. General geology of the Lake of the Woods greenstone belt with isotope sample locations.

rock units having an estimated stratigraphic thickness of about 10 km. Three distinctive rock associations comprise the assemblage: (1) mafic to felsic volcanic rocks of calc-alkaline affinity which have fractionated REE patterns (Fig. 2.5b) suggestive of derivation by small degree of partial melting from relatively deep mantle sources (Ayer and Davis, 1997). U-Pb zircon ages from felsic volcanic rocks in the lower stratigraphy range from 2719 to 2723 Ma; (2) tholeiitic mafic volcanic rocks and komatiites in sequences up to 2 km thick (intercalated with the more extensive calc-alkaline sequences) are more primitive than the Lower Keewatin Assemblage tholeiites with higher Mg#s and more depleted LREE patterns (Fig. 2.5d); and (3) turbiditic sedimentary sequences in the upper stratigraphy of the assemblage. Zircons from felsic pyroclastic rocks intercalated with the turbidites have a U-Pb age of about 2712 ± 2 Ma (Davis, 1996). The Electrum Assemblage consists of the Crowduck and the White Partridge groups and is restricted to the northern part of the LWGB (Fig. 5.1). The groups unconformably overlie the Lower and Upper Keewatin assemblages and consist of clastic sedimentary rocks with subordinate shoshonitic to calc-alkaline volcanic units with highly fractionated REE patterns (Fig. 2.5e). Detrital zircons indicate deposition after 2699 ± 2 Ma (Chapter 2).

There are a variety of U-Pb zircon determined ages of intrusive events in and around the LWGB (Blackburn et al., 1991). An early event consists of tonalite to granodiorite in stocks and batholiths with ages that range from 2717 to 2724 Ma, coeval with early volcanism in the Upper Keewatin Assemblage. Syn- to late-tectonic granodiorite intrusions post date Upper Keewatin volcanism, ranging in age from 2709 to 2711 Ma. A final intrusive event is represented by late to post tectonic alkaline to granite plutons coeval with the Electrum Assemblage volcanism. The Falcon Island pluton is a member of this group and has an age of 2695 ± 3 Ma (Chapter 2). The

ultrapotassic rocks of the pluton have highly fractionated REE patterns similar to the Electrum Assemblage shoshonitic volcanic rocks (Fig. 2.5f).

The Winnipeg River is a predominantly granitic subprovince located immediately to the north of the LWGB (Fig. 2.1). Granodiorite intruded along the contact with the Winnipeg River Subprovince has an age of 2709 ± 4 Ma providing the maximum age for juxtaposition of the two terrains (Corfu, 1988). The subprovince consists of gneissic and plutonic rocks spanning a large range of ages, subdivided into two groups: an older (2830-3170 Ma) gneissic sodic group intruded by post tectonic potassic plutons (2660-2709 Ma) (Beakhouse, 1991).

5.3 Sm-Nd isotopes

Samples for Sm-Nd isotope analysis were selected to represent the range of ages and compositions of the supracrustal rocks in the LWGB, and the range of composition of the Falcon Island pluton. The sample locations are shown on Figure 5.1 and results are presented in Table 5.1. Analytical techniques, the precision and accuracy and the sample descriptions are provided in the Appendix. Linear regressions for the Sm-Nd and Pb-Pb data were calculated using "ISOPLOT" a program of Ludwig (1992) utilizing the least squares regression techniques of York (1968).

Isotope data for the metavolcanic rocks of the study area are plotted on a Sm-Nd diagram in Figure 5.2. The samples are moderately well correlated and have a slope equivalent age of 2754 ± 65 Ma (m.s.w.d. = 39.3), an initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.509146 ± 31 and initial $\epsilon_{\text{Nd}} = +1.7$.

TABLE 5.1. WHOLE ROCK ND ISOTOPE DATA, LAKE OF THE WOODS GREENSTONE BELT AND FALCON ISLAND PLUTON

Location #	Sample #	Rock Type	Strata Unit	Sm ^a (ppm)	Nd ^a (ppm)	¹⁴⁷ Sm/ ¹⁴⁴ Nd +/-0.02%	¹⁴³ Nd/ ¹⁴⁴ Nd +/-0.002%	ε _{Nd} (T) +/-0.5	T ^b (Ma)	f _{SmNd}	T _{DM} ^d (Ga)
<u>Supracrustal Rocks</u>											
1	87-839	tholeiitic basalt	LKA	1.930	5.590	0.2086	0.512937	1.69	2738	0.06	>4.55
2	93-23	tholeiitic basalt	"	1.571	4.612	0.2059	0.512921	2.28	2738	0.05	>4.55
3	93-36	calc-alkalic andesite	UKA	3.173	15.100	0.1270	0.511456	1.34	2720	-0.35	2.79
4	93-37	tholeiitic basalt	"	1.495	4.399	0.2055	0.512881	1.68	2720	0.04	>4.55
5	93-18	high silica rhyolite	"	6.869	43.82	0.0947	0.510864	1.10	2719	-0.52	2.80
6	93-29	calc-alkalic rhyolite	"	1.988	11.95	0.1005	0.511019	2.11	2719	-0.48	2.72
7	93-30	komatiite	"	0.719	1.902	0.2285	0.513276	1.32	2720	0.16	2.38
8	93-26	komatiitic basalt	"	1.391	4.354	0.1931	0.512634	1.19	2720	-0.02	>4.55
9	93-27	evolved basalt	"	3.548	18.76	0.1143	0.511242	1.62	2720	-0.41	2.76
10	89-1116	wacke	"	4.600	26.27	0.1057	0.510963	-0.86	2712	-0.46	2.94
11	93-17	shoshonitic andesite	EA	5.440	28.00	0.1174	0.511249	0.41	2699	-0.40	2.84
12	93-35	calc-alkalic rhyolite	"	1.884	13.840	0.0823	0.510584	-0.38	2699	-0.58	2.85
<u>Falcon Island Pluton</u>											
13	88-182	clinopyroxenite		11.78	62.79	0.1134	0.511263	2.12	2695	-0.42	2.70
14	88-190	syenite		9.76	58.40	0.1010	0.511040	2.08	"	-0.48	2.71
15	88-195	monzodiorite		22.75	124.15	0.1108	0.511201	1.82	"	-0.43	2.73
16	88-261	syenite		7.36	55.07	0.0808	0.510734	3.12	"	-0.58	2.64
17	93-31	syenite		8.53	50.45	0.1022	0.511072	2.28	"	-0.48	2.69
18	93-32	melasyenite		20.82	121.26	0.1038	0.511085	1.98	"	-0.47	2.71
19	93-33	melasyenite		12.30	74.23	0.1002	0.511034	2.23	"	-0.49	2.70
Falcon Island pluton average values								2.23		-0.48	2.70

Units; LKA = Lower Keewatin Assemblage, UKA = Upper Keewatin Assemblage, EA = Electron Assemblage (Ayer and Davis 1997)

^a Isotope dilution determinations

^b Estimated age based on correlation with U-Pb geochronology (Ayer & Davis, 1997)

^c $f_{SmNd} = \{[(^{147}Sm/^{144}Nd_{sample}) / (^{147}Sm/^{144}Nd_{CHUR})] - 1\}$

^d T_{DM} calculated from ε_{Nd}(T) = 0.25 T² - 3T + 8.5 (Bennet & DePaolo, 1987)

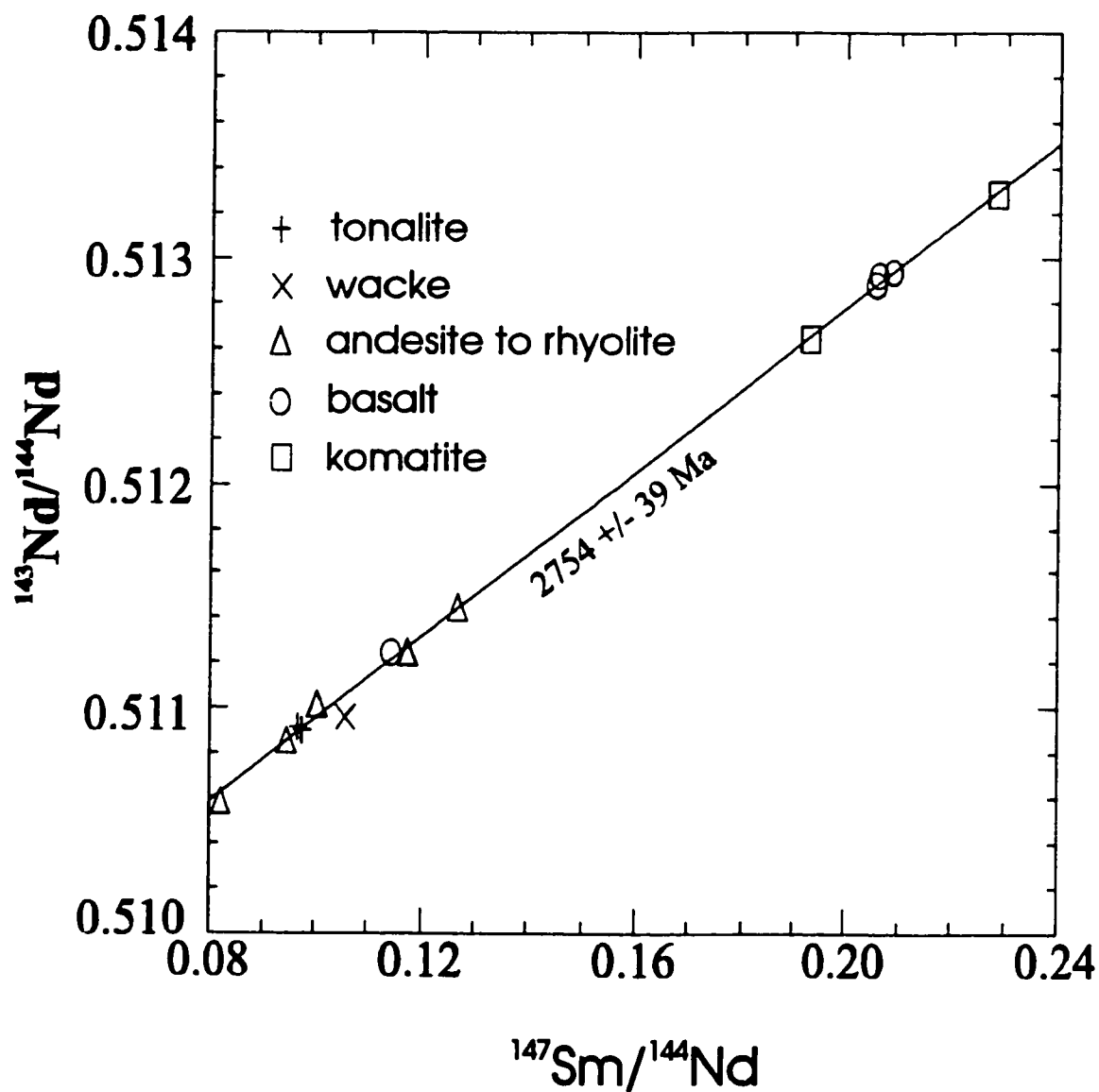


Fig. 5.2. Sm-Nd isochron for igneous samples from the Lake of the Woods greenstone belt supracrustal samples (Table 5.1). Aulneau batholith tonalite sample data is from Beakhouse and McNutt (1991). The wacke sample was not utilized in the isochron calculation.

The correlation is not a true isochron because differences in the U-Pb crystallisation ages indicate the rocks are not all coeval and their trace element characteristics (Fig. 2.5 & 2.6) indicate they come from different sources (Chapter 2). It is however, within the error of the U-Pb ages and is broadly similar to the 2737 ± 42 Ma Sm-Nd age and initial $\epsilon_{Nd} = +1.9$ for a Sm-Nd isotopic correlation of the igneous rocks in the Rainy Lake greenstone belt (Shirey and Hanson, 1986). The similarity in isotopic data of the two greenstone belts 150 km apart, with similar U-Pb ages (Davis et al., 1989), suggests large scale continuity for the Archean mantle beneath the Western Wabigoon Subprovince and that the Nd isotopic system remained largely closed to resetting.

The Nd isotopic ratios of variable evolved samples, representative of the fractionation trend of the Falcon Island pluton, yield a slope equivalent age of 2418 ± 150 Ma (m.s.w.d. = 3.5), an initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.509440 ± 102 , and an initial $\epsilon_{Nd} = -1.2$ (Fig. 5.3). The considerably younger age of the Falcon Island Sm-Nd isochron than the U-Pb determined crystallization age of 2695 ± 3 Ma for the pluton suggests that the correlation may represent a mixing line between end members with differing Sm/Nd ratios (discussed further below).

Figure 5.4 displays the secular variation of the Nd isotope values of individual samples and suggests the presence of heterogeneous sources. Tholeiitic basalt samples from the Lower Keewatin Assemblage (Bigstone Bay and Snake Bay groups) (Fig. 5.1) have positive ϵ_{Nd} values of +1.7 and +2.3 respectively. These values in conjunction with their flat REE patterns (Fig 2.5a), indicates that the tholeiites were derived from juvenile depleted mantle with no evidence for crustal interaction. These ϵ_{Nd} values are similar to mafic sequences with the same range of

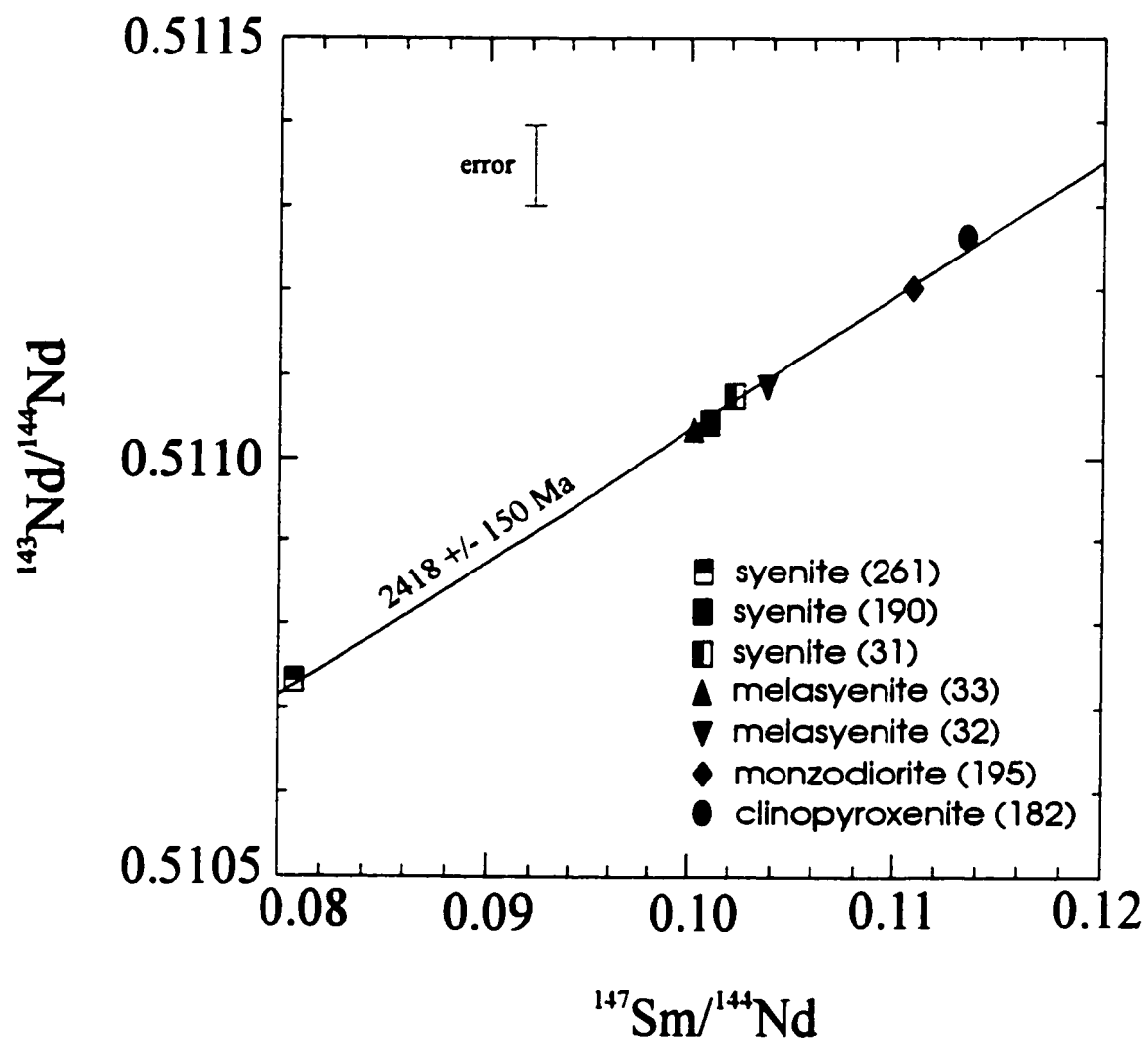


Fig. 5.3. Sm-Nd isochron for samples from the Falcon Island pluton.

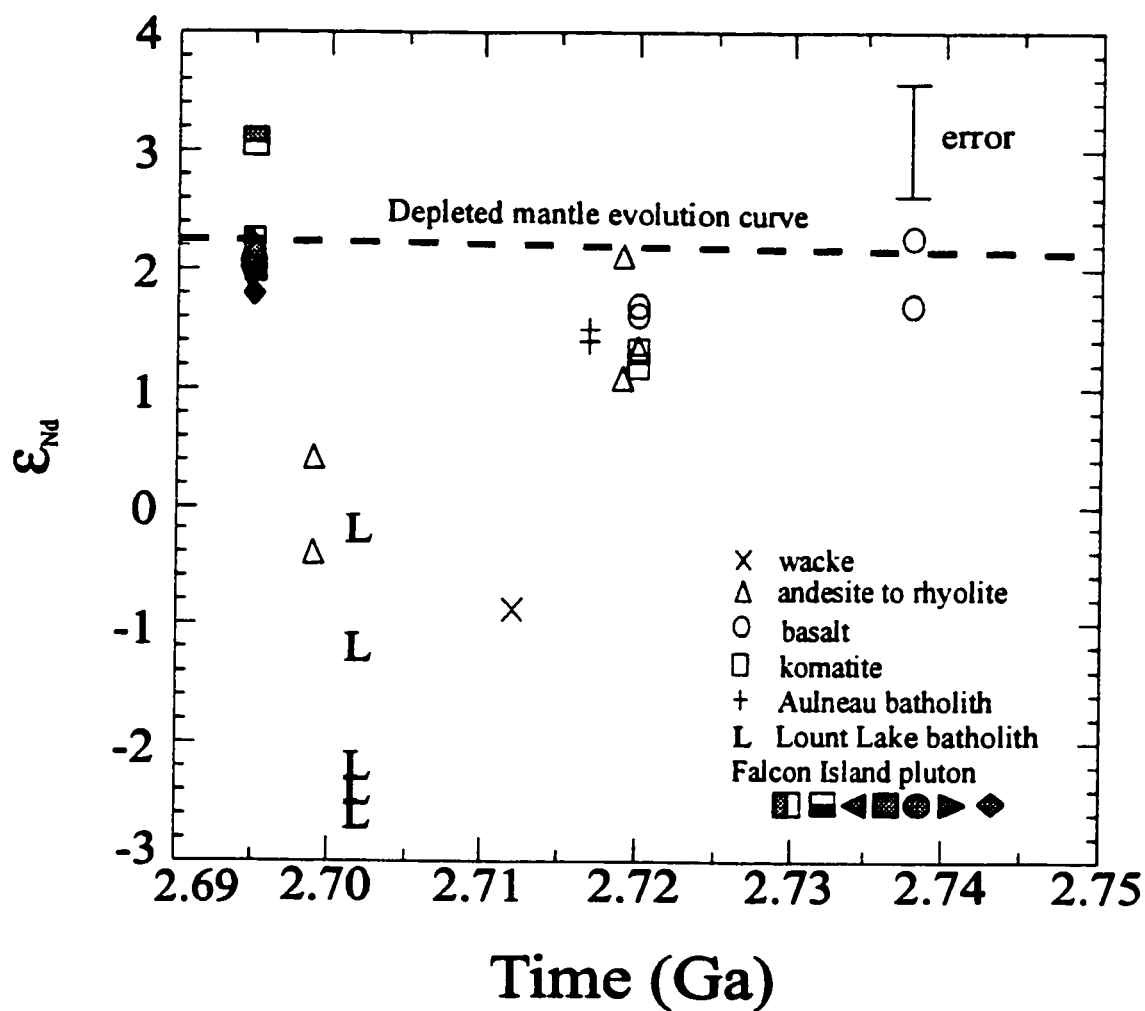


Fig. 5.4. An ϵ_{Nd} vs. time diagram for samples from the Lake of the Woods greenstone belt, Falcon Island pluton (symbols as in Fig. 5.3), and the Lount Lake batholith, from the Winnipeg River Subprovince (Beakhouse and McNutt, 1991). Depleted mantle evolution curve after Bennett and DePaolo (1987).

ages from other parts of the Western Wabigoon Subprovince, including tholeiites and anorthosites of the Rainy Lake greenstone belt (Ashwal et al., 1983; Shirey and Hanson, 1986) and the Mulcahy intrusion ($\epsilon_{Nd} = +2.6 \pm 1.2$) (Morrison et al., 1985). The values are also similar to slightly younger (i.e. 2700 to 2720 Ma) tholeiitic volcanic rocks and komatiites from the Abitibi Subprovince (Zindler, 1982; Dupré et al., 1984; Cattell et al., 1984; Muchado et al., 1986; Lehay et al., 1995).

The Upper Keewatin Assemblage metavolcanic samples (ca. 2719-2723 Ma) have ϵ_{Nd} values of +1.1 to +2.1 and are within error (± 0.5 epsilon unit) of the Lower Keewatin Assemblage tholeiites (Table 5.1). The Upper Keewatin samples include komatiite, basaltic komatiite, tholeiite, evolved tholeiite (i.e. with fractionated REE patterns), calc-alkaline andesite and rhyolite and high silica rhyolite with a broad range of compositions and highly variable trace element patterns (Fig. 2.5). The lack of any systematic change in ϵ_{Nd} values with SiO_2 (or variable degrees of REE fractionation) for the Lower and Upper Keewatin volcanic samples (Fig. 5.5) is consistent with a relatively homogeneous mantle reservoir.

Beakhouse and McNutt (1991) indicate that Aulneau batholith (Fig. 5.1) tonalite samples (ca. 2717 Ma) yield ϵ_{Nd} values of +1.4 and +1.5. These values are very similar to those of the Upper Keewatin volcanic samples and further indicate homogeneous isotopic characteristics for the pre 2.7 Ga LWGB magmatism. They attributed the tonalite to melts derived by partial melting of metamorphosed (hydrated) basalts with a short residence time in the lower crust. Based on ϵ_{Nd} values, the tonalite could have been derived by partial melting of Lower or Upper Keewatin Assemblage mafic volcanic rocks. The older mafic sequences may

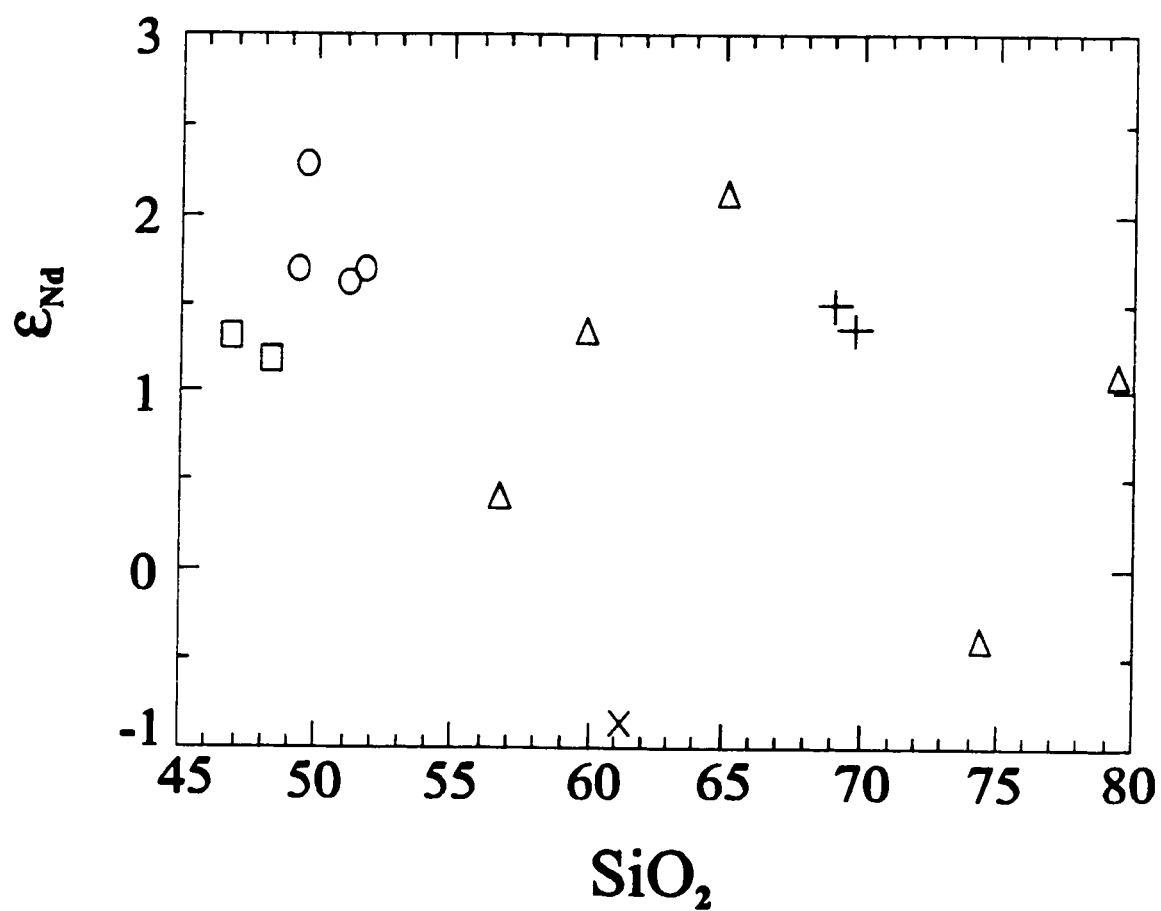


Fig. 5.5. An ϵ_{Nd} vs. SiO_2 diagram for samples from the Lake of the Woods greenstone belt and Aulneau batholith. Symbols as on Figure 5.2.

represent a more likely source however as units such as the Snake Bay Group of the Lower Keewatin Assemblage are thicker and occur draped around an extensive dome cored by the Batholith (Fig. 5.1).

Warclub Group turbiditic sedimentary rocks deposited at about 2712 Ma (Davis 1996) occur within the upper stratigraphy of Upper Keewatin Assemblage. Wacke the Upper Keewatin Assemblage in the southern part of the belt (Fig. 5.1) yields an $\epsilon_{Nd} = -0.86 (\pm 0.5)$. This value is considerably lower than the values of the volcanic samples from either the Lower or Upper Keewatin assemblages, or from the Aulneau batholith tonalite. A negative ϵ_{Nd} value for the wacke indicates a substantial input from a long-term LREE enriched source. Davis (1996) has determined that the Warclub Group turbidites contain detrital zircons with a wide variety of U-Pb ages ranging from 2.7 to 3.0 Ga. He attributes the older zircons to the Winnipeg River Subprovince which is located immediately north of the LWGB (Fig. 5.1), and which has U-Pb ages as old as 3.2 Ga (Beakhouse, 1991). The younger zircon population is attributed to local greenstone belt provenance on the basis of age and low U/Pb ratios (Chapter 2).

A relatively enriched Nd isotopic sources is also evident in shoshonitic andesite ($\epsilon_{Nd} = +0.4$) and calc-alkaline rhyolite ($\epsilon_{Nd} = -0.4$) (± 0.5) of the Electrum Assemblage (Table 5.1). The shoshonites are part of a suite of mantle-derived, late tectonic potassic rocks as indicated by elevated Mg#s (65-75) (Chapter 4). The Electrum Assemblage clastic rocks have detrital zircons ranging from 2699 to 2825 Ma which are consistent with a mixed provenance from the Winnipeg River and local volcanic sources (Chapter 2). Considerably lower ϵ_{Nd} values for the Electrum Assemblage volcanic samples, than those of the Lower and Upper Keewatin volcanic samples, is

consistent with input from a long-term LREE-enriched isotopic reservoir. An enriched Nd isotopic reservoir is also evident in the Lount Lake batholith, intruding the central part of the Winnipeg River Subprovince. Lount Lake granitic samples yield ϵ_{Nd} values which range from -0.4 to -2.6 and are derived by partial melting of a crustal source 200–400 Ma older than the crystallization age of the batholith ($2702 \pm 4/-3$ Ma) (Beakhouse and McNutt, 1991). The negative ϵ_{Nd} values (avg. = -1.9) of Lount Lake batholith also could suggest that its source might also have represented a contaminant source for mixing with depleted mantle magmas to form the Electrum Assemblage samples (avg. $\epsilon_{Nd} = 0.0$) (Fig. 5.4).

Evidence for depleted mantle at the time of deposition of the Electrum Assemblage (i.e. post 2.7 Ga) is demonstrated by the positive ϵ_{Nd} values for the Falcon Island pluton samples (avg. $\epsilon_{Nd} = +2.2$) (Table 5.1). However, when considered individually, the successively evolved units of the pluton display asymptotic increases in ϵ_{Nd} values from a low of $+1.8$ to a high of $+3.1$ (± 0.5) with increasing SiO_2 and decreasing Mg#s (Fig. 4.10). These hyperbolic trends are consistent with a petrogenesis by mixing between end-member components with differing Nd isotopic ratio (Langmuir et al., 1978) and degrees of ϵ_{Nd} depletion. Other lines of evidence which are also consistent with mixing include: (1) a Sm-Nd isochron with a substantially lower initial ϵ_{Nd} value (i.e. $\epsilon_{Nd} = -1.2$) (Fig. 5.3) than the values of the samples calculated individually (Table 5.1); and (2) a slope-derived isochron age of 2418 ± 150 Ma (Fig. 5.3), which is significantly younger than the U-Pb determined crystallization age of 2695 ± 3 Ma, indicating a lack of age correspondence for different isotopic systems in the same rocks (Faure 1986). Mixing by assimilation fractional crystallization could also produce these results if the assimilated country

rocks were more depleted than the initial magma. This is considered unlikely because the metavolcanic and metasedimentary rocks of the LWGB have equivalent or lower ϵ_{Nd} values than those of the pluton (Fig. 5.4).

The T_{DM} ages of the Falcon Island samples also suggests that there is little likelihood for crustal interaction as they display a relatively limited range from 2.62 to 2.72 Ga (avg. = 2.70 Ga; Table 5.1) and are similar to the U-Pb age of the pluton (2695 Ma). This is in distinct contrast to the wide range of T_{DM} ages for the supracrustal rocks, which are typically older than their U-Pb crystallization age. Correspondence of T_{DM} and crystallization ages, in conjunction with positive ϵ_{Nd} values, is considered evidence for little interaction between crustal components and depleted mantle magmas (Jalsma et al., 1996). Thus mixing of successive magma batches from a heterogeneous depleted mantle source is considered to be the most likely explanation.

5.4 Common Pb isotopes of the Falcon Island pluton

Lead isotope analyses on alkali feldspar separates from four locations within the Falcon Island pluton are tabulated in Table 5.2. Alkali feldspar has a low uranium content and consequently very low abundances of radiogenic Pb. Thus analysis of the Pb isotopes of alkali feldspar can provide a useful tool to estimate the initial Pb isotopic ratios of magma sources. Acid leaching removes surface contaminants and loosely bound Pb (presumably radiogenic in origin). Any remaining sources of radiogenic Pb will most likely be from fine-grained mineral or fluid inclusions that contain U or Th as well as Pb.

TABLE 5.2. LEAD ISOTOPES OF ALKALI FELDSPAR, FALCON ISLAND PLUTON

	<u>Rock type</u>	<u>$^{206}\text{Pb}/^{204}\text{Pb}$</u>	<u>$^{207}\text{Pb}/^{204}\text{Pb}$</u>	<u>$^{208}\text{Pb}/^{204}\text{Pb}$</u>	<u>206/207 Model Age</u>
88-190	syenite	13.533	14.576	33.368	2618 Ma
93-31	syenite	15.897	15.004	36.150	981 Ma
93-32	melasyenite	13.687	14.590	33.565	2452 Ma
93-33	melasyenite	13.748	14.603	33.417	2403 Ma
93-33 (duplicate)		13.856	14.618	33.528	2302 Ma
93-33 (biotite-rich fraction)		14.229	14.689	33.913	2031 Ma
93-33 (unleached fraction)		14.235	14.708	34.122	2044 Ma

Four different splits of sample 93-33 (Table 5.2) were run to check the analytical uncertainty, and the efficiency of acid leaching in removal of radiogenic Pb. Two splits free of visible contamination were acid leached. They yielded results which are slightly above the quoted 2σ error (Appendix III). This could be the result of analytical error, or error caused by small inclusions. An estimate of the true uncertainty (analytical and geological) is thus about ± 0.02 for $^{207}\text{Pb}/^{204}\text{Pb}$ and ± 0.1 for both $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$. A split with visible biotite inclusions yielded more radiogenic Pb than the clean splits, but is still slightly less radiogenic than the unleached split. Differing results from the differently treated alkali feldspar splits of sample 93-33 indicate that there is a radiogenic Pb component in the feldspars, thus the initial Pb isotopic composition can not be directly determined by leaching. That the four different splits of sample 93-33 and the leached alkali feldspars from three other rock units all fall along a secondary isochron with a slope yielding an age of 2700 ± 100 Ma (m.s.w.d. = 95.2) (Fig. 5.6), similar to the crystallization age of the pluton (2695 ± 3 Ma), indicates that the radiogenic Pb component in the feldspars was produced by in situ decay of U and Th and that the Pb isotopic system has remained closed since crystallization (Gariépy and Allègre, 1985). The $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ ratios of the feldspars form an array which is parallel to the Stacey and Kramers (1975) Pb evolution curve (S-K on Fig. 5.7).

The $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ values of the feldspar samples all plot well below the S-K crustal evolution curve on Figure 5.6 indicating derivation from a LILE depleted mantle source. In distinct contrast, Pb isotope values from a number of leached alkali feldspar samples from gneissic and plutonic samples from the Winnipeg River Subprovince all plot well above the S-K curve, indicating that they were derived from an enriched crustal source (Corfu, 1988).

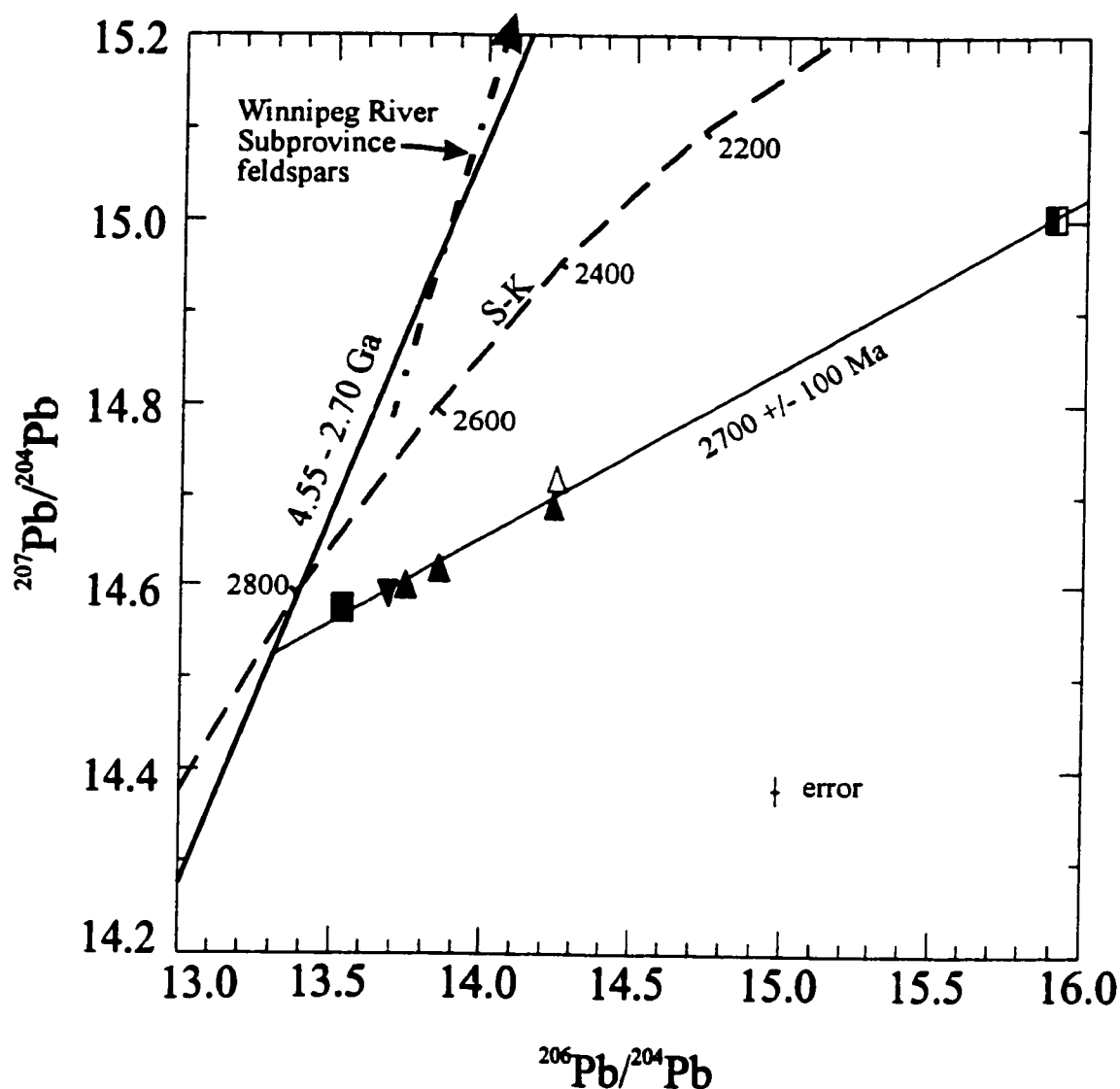


Fig. 5.6. A $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagram showing the isochron (2700 ± 100 Ma) for feldspar separates from the Falcon Island pluton. Leached feldspars data from rock samples symbols as indicated on Figure 5.3, the open triangle and lightly shaded triangle represent an unleached fraction and a biotite-bearing, leached fraction of sample 93-33 respectively. An isochron for single stage evolution from 4.55 to 2.70 Ga with Canyon Diablo meteorite compositions (Tatsumoto et al. 1973) and the S-K (Stacey and Kramers, 1975) crustal evolution curve are also illustrated, as is the range of values for leached feldspar separates from Winnipeg River Subprovince plutonic rocks (Corfu 1988).

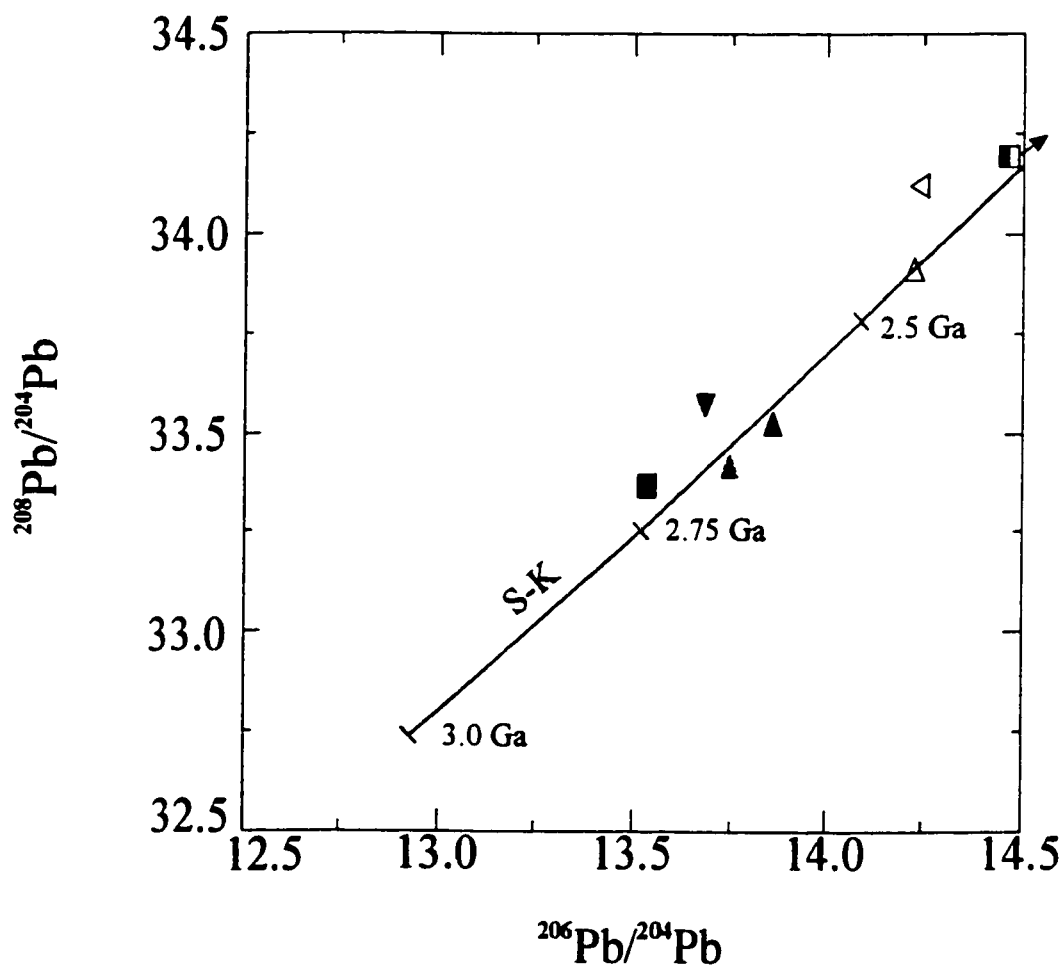


Fig. 5.7. A $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagram for feldspar separates from the Falcon Island pluton. Symbols as on Figure 5.6. S-K represent the Stacey and Kramers (1975) crustal evolution curve.

Feldspar from syenite sample 88-190 yields the least radiogenic Pb values with a single stage $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ model age of 2618 Ma closest to the crystallization age of the pluton (Table 5.2). The leached feldspar from melasyenite samples 93-32 and 93-33 yield slightly more radiogenic Pb isotopes, while the leached feldspar from syenite sample 93-31 is significantly more radiogenic than all of the other samples, including the biotite-bearing and unleached fractions. It is displaced a considerable distance from the S-K curve on Figure 5.6 and has a model age of 981 Ma, which is considerably younger than the crystallization age. Elevated $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ values suggests that inclusions in the feldspar may contain significant Th as well as U. The sample is from the margin of the pluton (Fig. 5.1) and unlike the massive internal units of the pluton, it is foliated and contains epidote and andradite overgrowths indicative of late stage alteration. This suggests the excess radiogenic Pb may have been acquired by contamination or during a multistage history (Faure, 1986). However, it is likely that the contamination occurred around the time of emplacement of the pluton as the $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios of this feldspar sample lie on the same 2.7 Ga isochron as the less radiogenic feldspars. Doe and Hart (1963) have documented scavenged radiogenic Pb in feldspars up to 300 meters from the contact of the Eldora stock in Colorado indicating that radiogenic Pb can be extensively remobilized during contact metamorphism.

Regressing the Falcon Island pluton feldspar isochron to intersection with a single stage growth curve geochron from 4.55 to 2.70 Ga with Canyon Diablo troilite Pb isotopic composition (Fig. 5.6) (Tatsumoto et al., 1973) yields initial $^{207}\text{Pb}/^{204}\text{Pb} = 14.52$ and $^{206}\text{Pb}/^{204}\text{Pb} = 13.29$, corresponding to a reservoir with a $\mu_1 = 7.86$.

5.5 Discussion

5.5.1 Nd Isotopes

The secular variation in Nd isotopes displayed on Figure 5.4 indicates there is no consistent linear trend in mantle isotopic evolution over 45 millions years of Neoarchean crustal evolution in the LWGB. Rather, variations are best explained by mixing between two end-member components: (1) juvenile depleted mantle with ϵ_{Nd} values ranging from about +1 to +3; and (2) a crustal source with long-term enrichment in LREE and ϵ_{Nd} values ranging to from about -1 to -3. Intermediate values between those in of the end-member ranges are most probably the result of mixing in various proportions of the two components, either by depositional processes in the sedimentary rocks, or by mixing of magmas in the Electrum Assemblage volcanic suite.

Figure 5.8 is a diagram of $f^{Sm/Nd}$ vs. ϵ_{Nd} for all LWGB samples. This type of plot was introduced by Shirey and Hanson (1986) to show the duration for development of enriched crustal versus depleted mantle reservoirs in the Rainy Lake area of the Wabigoon Subprovince. $f^{Sm/Nd}$ is defined as the enrichment factor of ^{147}Sm to ^{144}Nd in a sample relative to the Chondritic Uniform Reservoir (CHUR) (DePaolo and Wasserburg, 1976). Rocks with LREE enriched patterns plot in the lower two quadrants and depleted samples plot in the upper quadrants. Depletion or enrichment of the LREE by geological processes such as partial melting, fractionation or metasomatism can be considered instantaneous when compared to the half life of ^{147}Sm and thus would be represented by a vertical array. As ^{144}Nd appears in the denominator of

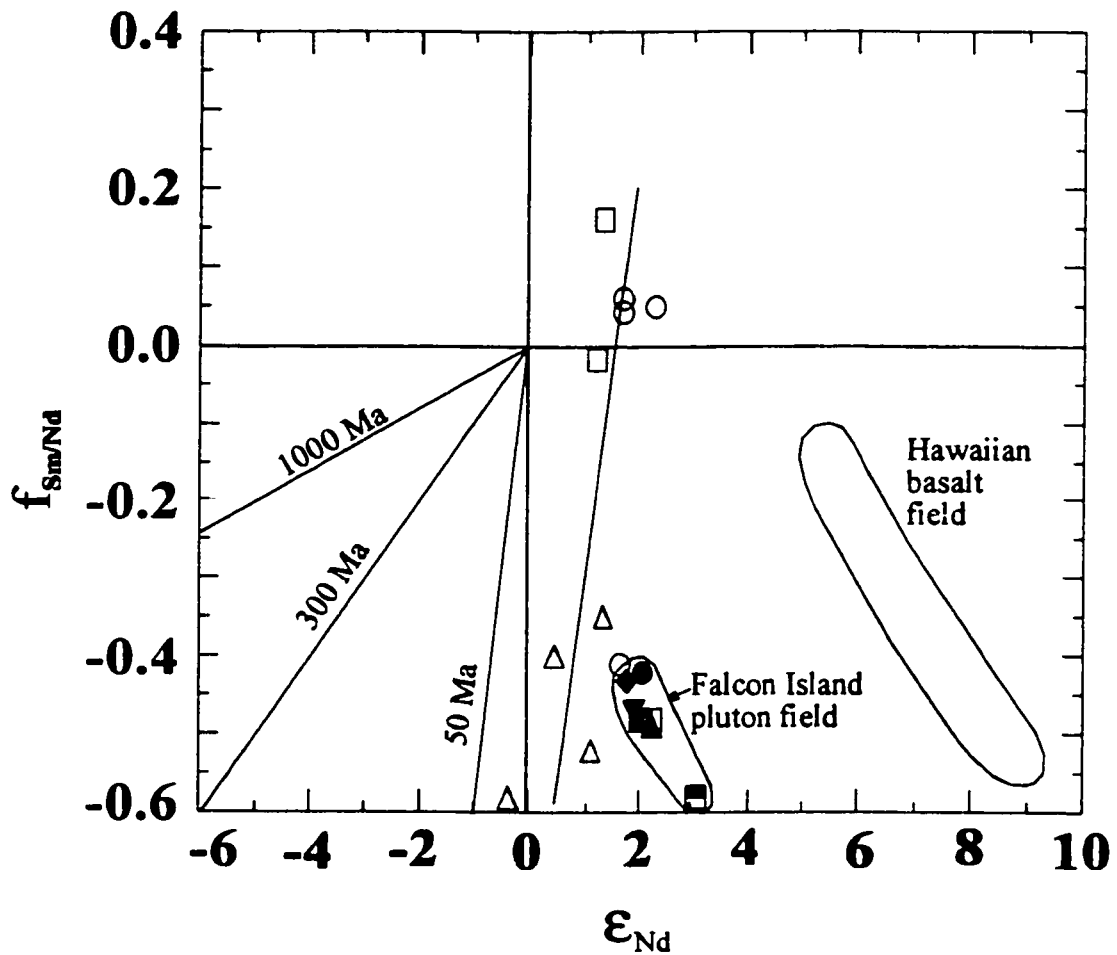


Fig. 5.8. A $f^{\text{Sm/Nd}}$ vs. ϵ_{Nd} for the volcanic rocks from the Lake of the Woods and the Falcon Island pluton. Symbols as on Figures 5.2 and 5.3. The volcanic samples are weakly correlated around a line with a slope equivalent age of 50 Ma. Reference 1000 Ma, 300 Ma and 50 Ma isochrons are illustrated in the lower left quadrant. The Falcon Island samples defines a field with a negative correlation which suggests an origin by source mixing. The field of Hawaiian basalts (O'Nions et al., 1977) is shown for comparison.

both axes, and if the system has remained closed from the point at which a vertical array was created, a straight line relating different samples will be an isochron with a slope which is equivalent to the time required for separation of the differing source reservoirs. The slopes of isochrons representing various hypothetical time spans are shown in the lower left quadrant. Although diffuse, the array of the LWGB samples suggests a steep slope representing a duration of about 50 M.y. This time span is similar to the range of crystallization ages as determined by the U-Pb isotopic system and confirms the duration for development of the crust and the juvenile nature of the initial mantle source. Time spans of about 70 my have been determined by similar techniques for the Rainy Lake and Abitibi greenstone belts (Shirey and Hanson, 1986).

Partial melting of mantle which is heterogeneous at the time of depletion, or enrichment by melting or a fluid related process, will form roughly linear arrays on an ϵ_{Nd} vs. $f^{Sm/Nd}$ diagram. The slope of positive trends will be a function of the time since the homogenization of an earlier heterogeneity and negative slopes indicate mixing of sources with differing isotopic ratios as is indicated for the field of Hawaiian Oceanic Island basalts (OIB) samples (Fig. 5.8) (Shirey, 1986; Shirey and Hanson, 1986). The negative slope represented by the field of the Falcon Island pluton samples (Fig. 5.8), although not as extensive as the modern mantle mixing trend evident in the Hawaiian field, is also most likely the result of mixing

A number of other post 2.7 Ga potassic igneous suites throughout the Superior Province yield similarly positive Nd isotopes values to those of the Falcon Island pluton. They include ϵ_{Nd} values of +1.0 to +1.9 (avg. = +1.4) for monzodiorite, hornblendite and trachyandesite from the Rainy Lake area (Shirey and Hanson, 1986), $+2.3 \pm 0.6$ for the Otto Stock (Tilton and Kwon, 1990), $+1.9 \pm 1.6$ for the Timiskaming Group alkalic volcanic suite (Basu et al. 1984) and +2.1

to + 3.2 (avg. = +2.7) in the Houghton pluton (Bédard, 1996). These positive ϵ_{Nd} values (similar to those of the pre-2.7 Ga subalkaline volcanic suites), in conjunction with elevated LILE and LREE, is consistent with derivation from depleted mantle sources which have been subjected to a metasomatic event not experienced by pre-2.7 Ga mantle (Shirey and Hanson, 1986; Stern et al., 1989; Stern and Hanson, 1991).

5.5.2 Pb isotopes

Lead isotopic data from igneous suites are useful for determining the chemical nature of the source reservoirs. Determining the initial lead compositions of Archean rocks is a difficult task as the various methods often yield differing results for the same samples and some are more appropriate to the specific materials utilized. The four main methods include: (1) the technique used for the Falcon Island feldspars in which the initial ratios are based on intersection of an isochron determined from leached minerals with low to negligible radiogenic Pb contents and a single stage geochron for the crystallization age of the magma (Gariépy and Allègre, 1985; Carignan et al., 1993; 1995); (2) calculation of the initial values from Pb isotopic ratios determined from leached mineral residues corrected for in situ decay of U by utilizing the crystallization age of the magma and precise U/Pb isotope ratios or U and Pb concentrations determined by isotope dilution (Kwon et al., 1989; Tilton and Kwon, 1990; Tilton and Bell, 1994), (3) isochrons based on whole rock isotopic analysis of volcanic samples and silicate mineral separates, or analysis of sulphide mineral separates genetically associated with volcanic rocks (Tilton, 1983; Dupré et al., 1984; Brévart et al., 1986; Dupré and Arndt, 1990; Vervoort et

al., 1994; Carignan et al., 1995), (4) Pb isotopic ratios of galena associated with VMS deposits (Tilton, 1983; Vervoort et al., 1994).

Detecting crustal contamination using Pb isotopes strongly depends on the U/Pb ratios of the sources and the crustal residence times. Significant variations in the production of ^{207}Pb compared to ^{206}Pb occurs in Archean reservoirs having different U/Pb ratio as a result of the short half life and higher abundance of ^{235}U relative to ^{238}U . Thus melts derived from uncontaminated mantle, crust-contaminated mantle and older crust should be detectable by progressively higher $^{207}\text{Pb}/^{204}\text{Pb}$ values (Gariépy and Allègre, 1985; Carignan et al., 1993; 1994). The Pb isotopes of the Falcon Island pluton have a similar $^{206}\text{Pb}/^{204}\text{Pb}$, but slightly higher $^{207}\text{Pb}/^{204}\text{Pb}$ values than for fields determined for the Wabigoon and Abitibi mantle but are similar to the tonalite-trondhjemite plutons and associated calc-alkaline volcanic suites of the southern Abitibi Subprovince (Fig. 5.9). The more elevated values in the calc-alkaline suites is attributed to minor sediment contamination of the mantle source (Carignan et al., 1993; 1995). Progressively higher $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ are evident in the LREE enriched plutons and syn- to late-tectonic tholeiites of the Abitibi, Wawa and Wabigoon subprovinces; gneisses and late granitoids of the Pontiac, Opatica and Abitibi subprovinces; and gneisses and late granitic rocks of the Winnipeg River Subprovince respectively (Fig. 5.9).

5.5.3 Combined Nd-Pb isotopes

Pb and Nd isotopes of tonalite from a number of pre 2.7 Ga batholiths in the Wabigoon

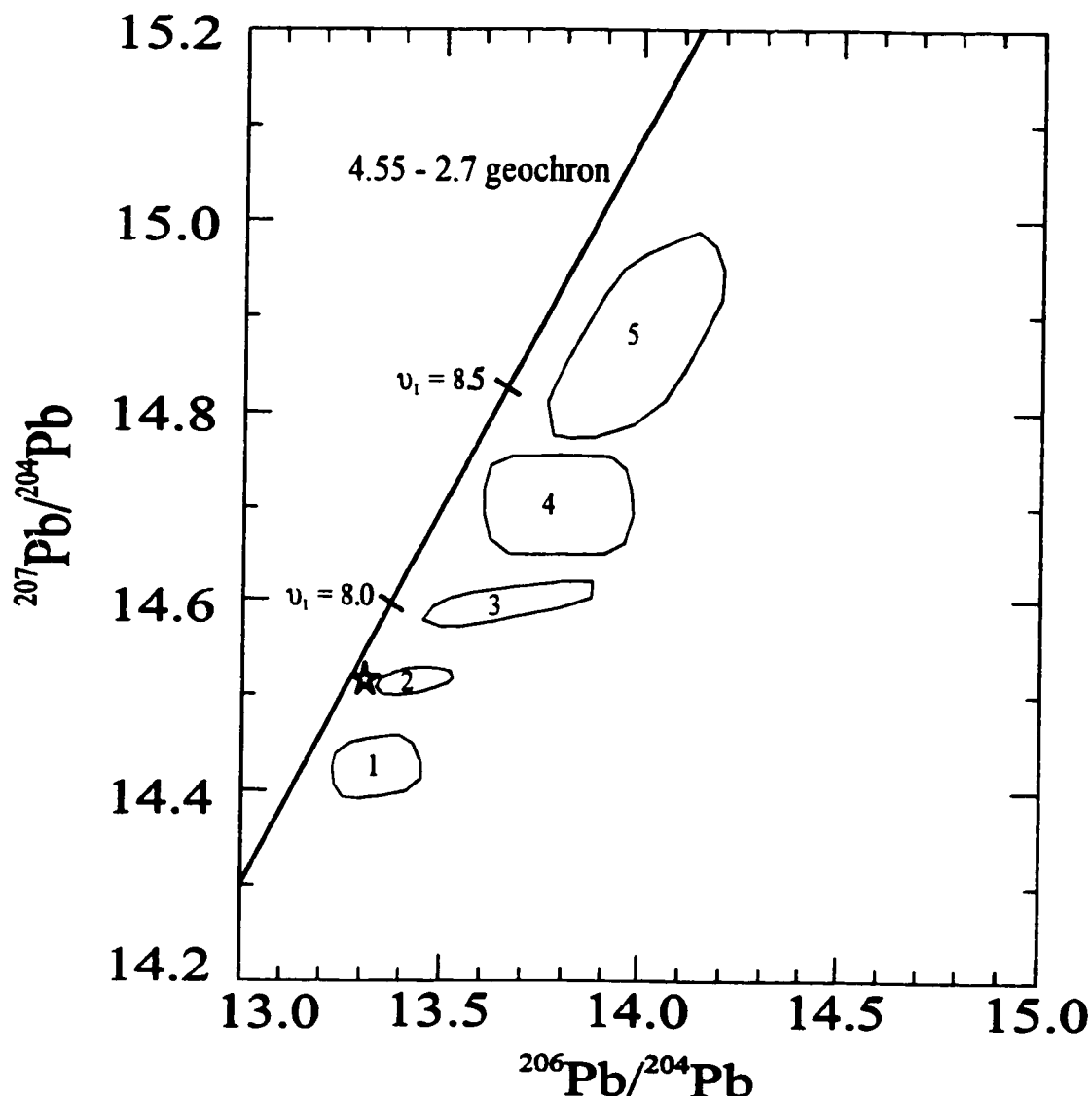


Fig. 5.9. Initial $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ isotopic compositions for various suites from the southern Superior Province (see Carignan et al., 1995). The star represents the initial Pb composition of the Falcon Island pluton; (1) is the field of southern Superior Province depleted mantle; (2) calc-alkaline tonalite-trondhjemite suites from the Abitibi Subprovince; (3) syn- to late-tectonic tholeiites and alkaline and LREE-enriched plutons in the Abitibi, Wabigoon and Wawa subprovinces; (4) gneisses and late granitic plutons from the Pontiac, Abitibi and Opatica subprovinces; (5) gneisses and granitic plutons from the Winnipeg River Subprovince. Also shown is a single stage 2.7 Ga isochron calculated using 4.55 Ga as the age of the earth and the composition of Canyon Diablo meteorite (Tatsumoto et al. 1973) with μ_i reservoir values of 8.0 and 8.5.

Subprovince have heterogeneous compositions. Batholiths intruded into greenstone belts such as the Lawrence Lake (2.73 Ga), the Lewis Lake (2.74 Ga) and the Revel, have mantle-like isotopic compositions with positive ϵ_{Nd} values of up to +2 and $^{207}Pb/^{204}Pb$ ratios between 14.5 to 14.6 (Kwon, 1986; Henry et al., 1996). Similar to the petrogenesis of the Aulneau batholith tonalite, they are best explained by the absence of significantly older crustal material and relatively fast recycling of crustal material after extraction from the mantle. The Indian Lake batholith (2.73 Ga), which intrudes Paleoproterozoic (2.9 to 3.0 Ga) granitic rocks of the central Wabigoon Subprovince, has a more crustal-like ϵ_{Nd} value (+0.4) and $^{207}Pb/^{204}Pb$ ratio (14.77) indicating contamination by older crust (Henry et al., 1996).

Combined initial Pb and Nd isotopic data for well-documented mantle-derived igneous suites from the southern Superior Province are shown on Figure 5.10. The Superior Province depleted mantle field is derived from a variety of sources including the Mulcahy intrusion (Morrison et al., 1985; Carignan et al., 1995), U-poor sulphides in Abitibi Subprovince komatiite flows (Machado et al., 1986; Br  vard et al., 1986) and galena from volcanogenic massive sulphides (VMS) (Burgeon et al., 1989; Deloule et al., 1989; Vervoort et al., 1994). A crustal field is defined by gneissic and plutonic suites from the Winnipeg River Subprovince (Corfu, 1989; Beakhouse and McNutt, 1991) and the Pontiac and Opatika subprovinces (Carignan et al., 1993). The data suggests a negative hyperbolic correlation between ϵ_{Nd} and $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ between the mantle and crustal end-members.

Vervoort et al. (1994) interpreted the negative correlation between Nd and Pb isotopes of many of these suites as the result of mixing between mantle components variously depleted by

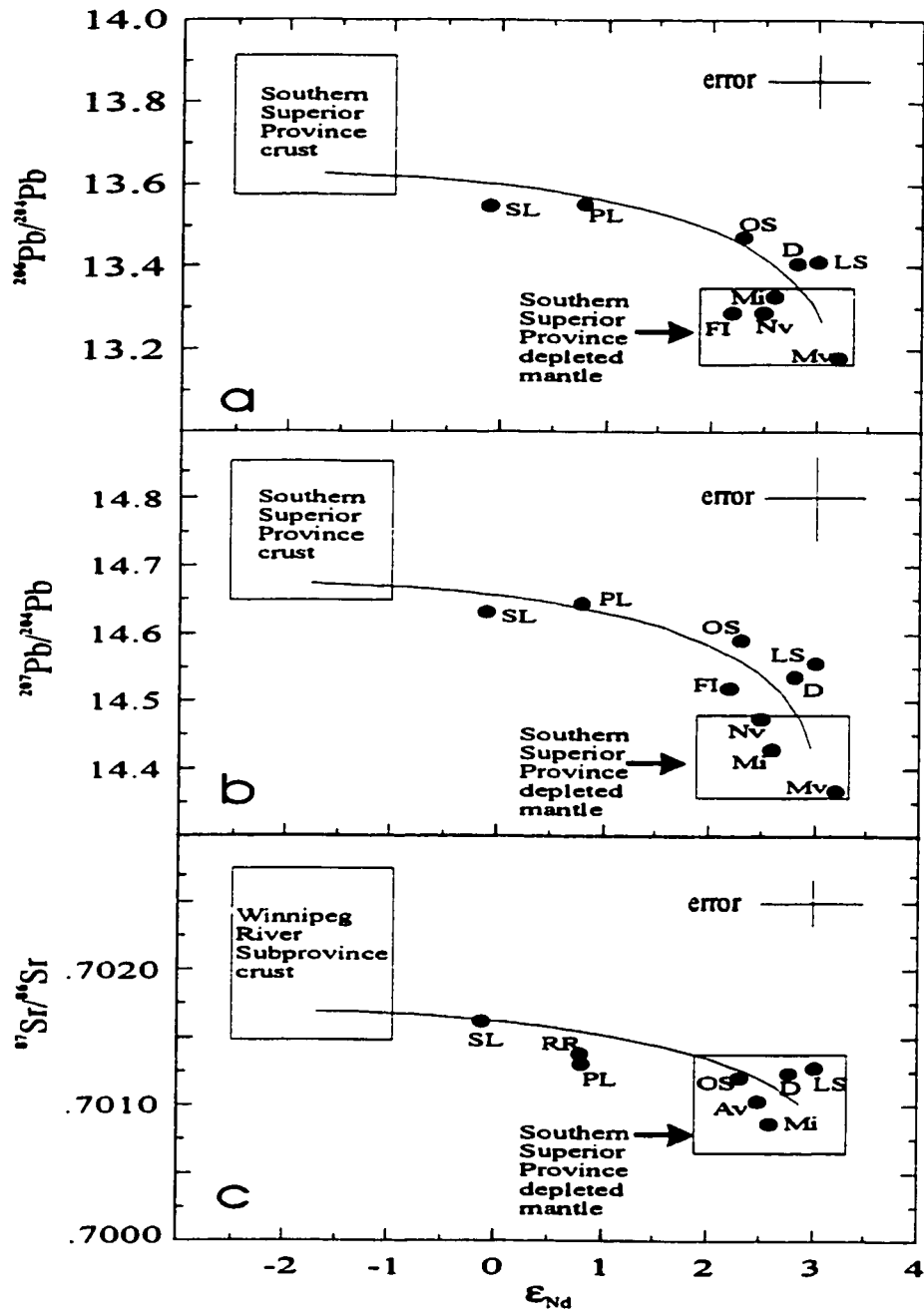


Fig. 5.10. Comparison of initial Nd, Pb and Sr isotopic values from a variety of sites across the southern Superior Province. Roaring River (RR) monzodiorite complex data from Stern and Hanson, (1991); Sturgeon Lake (SL), Poohbah Lake (PL), Otto stock (OS), Lac Shortt (LS) and Dolodau (D) alkaline complexes from Tilton and Bell (1994); Mulcahy intrusion (Mi) from Carignan et al. (1995); Noranda VMS galena (Nv) and Matagami VMS galena (Mv) from Vervoort et al. (1994); southern Superior Province depleted mantle fields derived from (Morrison et al., 1985; Machado et al., 1986; Brèvaried et al., 1986; Burgeon et al., 1989; Deloule et al., 1989; Vervoort et al., 1994; Carignan et al., 1995); southern Superior Province crustal field derived from (Gower et al., 1983; Corfu, 1989, Beakhouse and McNutt, 1991; Carignan et al., 1993).

an early melting event. However, the correlation of suites illustrated on Figure 5.10 is consistent with mixing of juvenile mantle melts with an enriched crustal component. The curvature of a mixing line between M (mantle) and C (crust) end-members in an ϵ_{Nd} vs. $^{207}Pb/^{204}Pb$ (or $^{206}Pb/^{204}Pb$) diagram is controlled by the $[Nd]/[Pb]$ ratio of each end-member, where $[Nd]$ and $[Pb]$ are the concentrations of Nd and Pb respectively and $r = ([Nd]/[Pb])_M / ([Nd]/[Pb])_C$ (Langmuir et al., 1978), where the numerical value of r is a measure of the degree of curvature between two end-members. If $r = 1$, then the mixing curve is a straight line. As r becomes progressively greater or less than 1, the hyperbolic curvature becomes more concave up, or down respectively. For an estimated average Archean crustal concentrations of $[Nd]/[Pb] = 20/10 = 2$ (Taylor and McLennan, 1985) and estimated primitive mantle concentrations $[Nd]/[Pb] = 1.35/0.185 = 7$ (Sun and McDonough, 1989); $r = 3.5$, yielding a convex hyperbola with a shape similar to the correlation curves in Figure 5.10. Comparing isotopic data from widespread geological environments can be problematic and may explain relatively poor correlation.

The enriched portion of the mixing line is defined by the Sturgeon Lake and Poohbah Lake alkaline complexes, both of which have relatively elevated $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ and low ϵ_{Nd} values (Figs 5.10a & 5.10b). The Eye Dashwa and the Smirch Lake plutons which are part of the post 2.7 Ga potassic suites, have ϵ_{Nd} values of +0.4 to +1.1 and elevated $^{207}Pb/^{204}Pb$ values of 14.64 to 14.68 which Henry et al. (1996) attributed to the participation of an older crustal component.

Crustal contamination may have occurred at the Poohbah complex which intrudes turbidite of Quetico Subprovince (Kwon, 1986) or the Electrum Assemblage with its proximity to the older crust of the Winnipeg River Subprovince, but this type of contamination is less likely

in plutons intruding the “greenstone” portions of the interior of the Wabigoon Subprovince, such as at Sturgeon Lake, Eye Dashwa and Smirch Lake. Relatively large variations in Nd and Pb isotopes for the post 2.7 Ga potassic suites, in conjunction with the systematic change to more positive ϵ_{Nd} values in the progressively evolved units of the Falcon Island pluton, are best explained by a heterogeneous mantle source region with variable Nd and Pb isotope characteristics. Phanerozoic potassic magmas with similar trace element geochemical characteristics are attributed to mantle sources metasomatized by the influx of LILE-rich fluids by partial melting of the sediment-enriched upper portions of subducting oceanic crust slab into an overlying depleted mantle wedge (Foley and Peccerillo, 1992; Hawkesworth et al., 1993).

5.5.4 Combined Nd-Pb-Sr isotopes

Using Pb-Sr-Nd isotopic studies of present-day oceanic island groups, Zindler and Hart (1986) interpreted correlations between isotopic systems as mixing curves between different components in the mantle. They identified the following four end-member components: (1) depleted mid oceanic ridge basalt (MORB) mantle (DMM), characterised by low Rb/Sr, low U/Pb and high Sm/Nd ratios considered to represent the source of MORB; (2) high U/Pb (HIMU) with elevated U/Pb ratios and no parallel increase in Rb/Sr and Sm/Nd is considered to have been derived from ancient, altered subducted oceanic crust; (3) enriched mantle I (EM I) possessing low U/Pb coupled with a low Sm/Nd and moderate Rb/Sr attributed to mixing of HIMU with a small amount of ancient pelagic sediments; and (4) enriched mantle II (EM II)

marked by high Rb/Sr, low Sm/Nd and moderate to high U/Pb with the isotopic composition of upper crustal rocks or modern pelagic sediments.

Tilton and Bell (1994) attribute the Sr, Nd and Pb isotopic patterns of young carbonatites (<0.2 Ga) to mantle processes involving mixtures between EM I and HIMU components with little involvement of DMM. Their patterns are thus similar to those of OIB (Zindler and Hart, 1986). They further suggest that EM I, EM II, HIMU and DMM as defined today were absent, or poorly developed in the Archean. This interpretation is based on data from Archean alkaline complexes which suggests a positive correlation between the Pb and Sr isotopes, but which is largely controlled by the elevated Sr values of the Sturgeon Lake pluton which may contain crustal components (Tilton and Bell 1994). However, mixing between distinctive reservoirs in an Archean mantle would yield less fractionated trends because the long half lives of these isotopic systems has allowed for more radiogenic growth in the present mantle. In addition, the Sr isotopic system appears to be less fractionated than Nd or Pb in the Archean mantle, merging with the bulk Earth evolution trend at about 2.85 Ga (Kwon et al., 1989; Tilton and Kwon, 1990; Tilton & Bell, 1994).

Figure 5.10c displays the initial $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} values for well documented, mantle-derived Neoproterozoic intrusions in the southern superior Province. Of the post 2.7 Ga intrusions, only the Otto stock, Lac Shortt and Dolodau carbonatites (Tilton and Bell, 1994) are coincident with the field for depleted mantle in southern Superior Province (Machado et al., 1986; Carignan et al., 1995). Other complexes such as Sturgeon Lake, Poobah Lake and Roaring River have more radiogenic Sr, Nd and Pb isotopic ratios (Tilton and Kwon, 1990; Stern and Hanson, 1991) and lie between the depleted mantle field and an enriched “crustal” field defined by the Sr

isotopes of plutonic suites from the Winnipeg River Subprovince (Gower, 1983; Beakhouse and McNutt, 1991).

Thus there appears to have been at least three distinctive isotopic reservoirs in the Neoproterozoic of the southern Superior Province: (1) a long-term enriched ancient source (ca. 3.0 Ga crust); (2) a pre-2.7 Ga, relatively homogeneous depleted mantle source; and (3) a post 2.7 Ga, heterogeneous mantle source. The Nd, Pb and Sr isotope data from igneous suites of the southern Superior Province are consistent with mixing of different isotopic reservoirs with characteristics that are similar (but less radiogenic) to those observed in the present-day mantle. An Archean equivalent to EM II may be represented by subducted sediments derived from long-term enriched crustal components such as the Winnipeg River, Opatica and Pontiac Subprovince with their higher initial Pb and Sr, and low initial Nd isotopic ratios. The pre-2.7 Ga subalkaline komatiite, tholeiite, calc-alkaline and granitic suites are derived from homogeneous isotopically depleted mantle with low Pb and Sr, and elevated Nd isotopic ratios and thus have characteristics analogous to present day DMM sources. The post-2.7 Ga potassic suites are derived from mantle which has variable initial isotopic ratios, ranging from depleted with ratios that are low for Pb and Sr, and elevated for Nd, to relatively high ratios for Pb and Sr and low for Nd. This is consistent with the potassic magmas being derived from mantle with an isotopic diversity representing mixing between the DMM- and EM II-like components (Shirey and Hanson, 1986; Stern et al., 1989; Henry et al., 1996).

The EM II-like enriched component may represent a Neoproterozoic mantle wedge metasomatized by the influx of isotopically enriched fluids derived from partial melting of sediments on a slab subducted beneath a Neoproterozoic crust thickened by over 30 My of

collisional tectonics (Ayer and Davis, 1997). The mixing evident in the post 2.7 Ga potassic suites thus contrasts the mixing between EM I and HIMU evident in modern OIB and carbonatite suites (Zindler and Hart, 1986; Tilton and Kwon 1991), but an EM I-like component appears to be missing from the southern Superior Neoproterozoic mantle (Tilton and Bell, 1994). The geodynamic implications of these differences are uncertain at this time and further detailed studies utilizing the Sr, Nd and Pb isotopic systems in both the subalkaline and alkaline suites of the Archean will be required to further refine the characteristics and to better understand why differences exist.

5.5 Conclusions

Nd isotopes of the igneous rocks of the LWGB greenstone belt yield an “isochron” of 2754 ± 65 Ma with an initial $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.509146 ± 31 and $\epsilon_{\text{Nd}} = +1.7$. The similarity of this age (within error) to independent U-Pb zircon ages of 2.70 to 2.74 Ga from the individual units demonstrates closure of this system since the Neoproterozoic. Agreement with the isochron for similar aged rocks in the Rainy Lake greenstone belt (2737 ± 42 Ma, $^{143}\text{Nd}/^{144}\text{Nd} = 0.509178 \pm 33$ and $\epsilon_{\text{Nd}} = +1.9$; Shirey and Hanson, 1986) suggests a broad-scale continuity of Neoproterozoic mantle sources across a 150 km wide section across the western Wabigoon Subprovince. The overall similarity of positive ϵ_{Nd} values for tholeiites, komatiites, and calc-alkaline volcanic rocks (+1.1 to +2.1) of the Lower and Upper Keewatin assemblages indicates they were derived from isotopically homogeneous mantle with a time averaged Sm/Nd depletion relative to CHUR.

Matching ϵ_{Nd} values in similar aged tonalite of the Aulneau Batholith (+1.5; Beakhouse and McNutt, 1991) suggests a short residence time for crustal recycling of partial melts of depleted lower crust.

A long time-averaged Sm/Nd enriched source is evident in the ϵ_{Nd} value of $-0.9 (\pm 0.5)$ for wacke within the upper stratigraphy of the Upper Keewatin Assemblage. Pronounced negative ϵ_{Nd} values also occur in the Lount Lake batholith (avg. -1.9) within the interior of the Winnipeg River Subprovince (Beakhouse and McNutt, 1991). These values in conjunction with a wide range of detrital zircon ages from 2.7 to 3.0 Ga in the turbidite sequences of the western Wabigoon (Davis 1996), suggest that enriched crustal sources in the Winnipeg River and the depleted igneous rocks of the greenstone belt provided a mixed provenance for the sediments.

Heterogeneity in the post-2.7 Ga mantle is demonstrated by low Nd values in the shoshonitic and calc-alkaline volcanic rocks of the Electrum Assemblage (-0.4 & $+0.4$) and depleted values in the Falcon Island pluton ultrapotassic rocks ($+1.8$ to $+3.1$). Hyperbolic correlation of increases in ϵ_{Nd} values with increases in SiO_2 and decreases in Mg#s for progressively evolved units of the pluton indicates petrogenesis by magma mixing of mantle components with variable time averaged Sm/Nd depletion. Mantle mixing is also supported by an errorchron with an age of 2418 ± 150 Ma, significantly younger than the U-Pb age of 2695 ± 3 Ma and an initial ϵ_{Nd} value of -1.2 (initial $^{143}Nd/^{144}Nd = 0.509440 \pm 102$) which is significantly lower than the calculated ϵ_{Nd} values of the individual units of the pluton. The T_{DM} ages of the individual samples are similar to the U-Pb crystallization age suggesting the mixing is not related to crustal contamination.

Pb isotopes of leached alkali feldspars from a number of units of the Falcon Island pluton plot on a $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ isochron with an age of 2700 ± 100 Ma. The isochron plots well below the Stacey and Kramers (1975) crustal evolution curve and has an initial $^{207}\text{Pb}/^{204}\text{Pb}$ of 14.52 and an initial $^{206}\text{Pb}/^{204}\text{Pb}$ of 13.29, corresponding to derivation from a 4.55 to 2.70 Ga single stage geochron reservoir with $\mu_1 = 7.86$. This indicate a mantle Pb source slightly more enriched in $^{207}\text{Pb}/^{204}\text{Pb}$, but similarly depleted in $^{206}\text{Pb}/^{204}\text{Pb}$, than the source of mafic and ultramafic magmas in the Wabigoon and Abitibi subprovinces.

Comparison of the Falcon Island Nd and Pb isotope data with other well-documented potassic and alkaline suites supports the suggestion that the post-2.7 Ga southern Superior Province mantle was isotopically heterogeneous. The isotopic data lie along an hyperbolic curve and is thus consistent with mixing between a pre-2.7 Ga depleted mantle component and an enriched component with isotopic characteristics similar to those of the ancient crustal rocks such as the gneissic and plutonic suites of the Winnipeg River, and an ancient detrital component in the Opatoca and Pontiac metasedimentary subprovinces. This type of heterogeneity is most readily explained by metasomatism of a mantle wedge by the influx of fluids derived by partial melting of ancient crustal sediments subducted beneath the Neoarchean crust.

Chapter 6

Conclusions

The main objective of this thesis is to explore the origin and evolution of the crust and mantle of the Lake of the Woods greenstone belt (LWGB) and to compare and contrast it with the development of similar-aged greenstone belts across the southern Superior Province. The secular evolution and geodynamic settings are constrained with precise geochronological, lithogeochemical and radiogenic isotopic data. Extensive detailed mapping in conjunction with precise U-Pb geochronology indicate the stratigraphy can be subdivided into 3 major lithostratigraphic units: a lowermost mafic volcanic assemblage; a compositionally diverse, predominantly volcanic, middle assemblage; and a predominantly sedimentary uppermost assemblage.

Although not perfect analogues, the lower Keewatin assemblage tholeiites bear some geochemical similarities to tholeiites from modern arcs and back-arc basins suggesting they may have been derived by destructive margin magmatism involving averaging about 20-30% partial melting of primitive mantle at the relatively shallow depth of less than 70 km. They lie at the base of the stratigraphic section in the LWGB and have been dated at 2738 Ma.

The Upper Keewatin calc-alkaline suite rocks are about 15 Ma years younger and disconformably overlie the Lower Keewatin assemblage. They are geochemically and petrologically similar to the calc-alkaline volcanic rocks forming in modern mature island arc settings. Intercalated tholeiitic and komatiitic suite rocks are geochemically similar but somewhat more primitive than the older Lower Keewatin tholeiites and may be the result of

second stage melting of the refractory mantle source from which the Lower Keewatin tholeiites were derived. They may represent the product of a subduction hydrated mantle and a higher Archean geothermal gradient.

The Electrum assemblage occurs as localised successor basins unconformably overlying the Upper Keewatin assemblage in the northern part of the LWGB. Detrital zircon ages indicate deposition after 2699 Ma. The volcanic rocks are of shoshonitic and calc-alkaline affinity and are most similar to those of modern shoshonitic suites forming in the late stages of development of mature island arcs and continental arcs.

Late-tectonic potassic intrusions are coeval with the Electrum volcanism and include the Falcon Island pluton (2695 Ma) and monzodiorite in the Don Lake stock. The rocks of the Falcon Island pluton are ultrapotassic on the basis of $K_2O > 3$ wt %, $MgO > 3$ wt % and $K_2O/Na_2O > 2$ (Foley et al., 1987), while Don Lake monzodiorites and the Electrum Assemblage shoshonites are lower in K_2O , contain normative quartz, and thus belong to the shonshonitic clan (Fig 4.3a). Similar geochemical characteristics, including elevated Al, K, Sr, Rb, Ba, LREE, depleted HREE and HFSE of the ultrapotassic, monzodioritic and shoshonitic units are attributed to a late subduction event which occurred from about 2695 to 2700 Ma across the western Wabigoon Subprovince (Ayer and Davis, 1997).

Stern et al. (1989) have defined the mantle-derived potassic rocks of the Superior Province as a sanukitoid suite based on chemical characteristics such as 55-60 wt % SiO_2 , $Mg\# > 0.6$, Ni & Cr both > 100 ppm, $Na_2O + K_2O > 6$ wt %, Sr & Ba both 600-1800 ppm and elevated light rare earth elements (LREE) without Eu anomalies. Members of this suite exhibit pronounced depletions in the HFSE elements (e.g. Brooks et al., 1982; Ujike, 1985; Laflèche et al., 1991, Stern and Hanson, 1991; Dostal and Mueller, 1992; Rowins et al., 1993) and are thus

very similar to the K-rich alkaline, shoshonitic and calc-alkaline magmatic association of the Sunda arc in Indonesian (Wheller et al., 1987). The potassic suite of the LWGB fit these criteria although the more evolved end members of the Falcon Island suite are less primitive than the typical sanukitoids having relatively lower Mg# and Ni, Cr and Co abundances (Figs. 4.2 & 4.4). Elsewhere, shoshonitic to ultrapotassic magmatism is attributed to late stage subduction events in the Neoproterozoic Ottawa orogeny (ca. 1080 Ma) of the Grenville Province (Corriveau et al., 1993) and the Paleozoic Caledonian orogeny (ca. 410 Ma) of NW Scotland (Thompson and Fowler, 1986) and thus spans over 2.7 Ga of Earth history

Geochemical patterns within the LWGB potassic suite suggest crystal fractionation and accumulation were important petrogenetic processes, while variation in the degrees of partial melting appears to be of less significance (Figs. 4.6, 4.7, & 4.8). Clinopyroxene, biotite, magnetite and possibly amphibole were accumulated in the mafic units and fractionated along with plagioclase in the evolved units. The differing trends on Figures 4.6 and 4.8 provide vectors pointing to a common parental composition which is similar to that of the Don Lake stock monzodiorite and the Falcon Island melasyenite. The unfractionated, relatively uniform composition of Don Lake monzodiorite, in conjunction with its close proximity to the Falcon Island pluton (Fig. 4.1), supports a common genesis and is consistent with the parental monzodiorite composition suggested from a number of studies of Superior Province sanukitoids (e.g., Shirey and Hanson, 1984; Sutcliffe et al., 1990; Stern and Hanson, 1991; Laflèche et al., 1991; Rowins et al., 1993).

U-Pb geochronological studies of the late-tectonic alkaline rocks and their associated sedimentary successor basins suggests that the alkaline magmatism became younger by 5-10 my increments towards the southern margin of the Superior Province. For example, U-Pb

zircon geochronology of Electrum Assemblage and associated alkaline plutons in the LWGB and a number of other localities across the western Wabigoon Subprovince indicate ages which range from 2695 to 2700 Ma (Blackburn et al., 1991). In the Wawa Subprovince, late-tectonic sedimentation and alkaline volcanism and plutonism occurred around 2685 to 2690 Ma (Corfu and Sage, 1992; Corfu and Stott, in press), while further to the south in the Abitibi Subprovince, Timiskaming Group sedimentation and alkaline magmatism ranges from 2675 to 2680 Ma (Corfu, 1993). This southerly progression of subduction-related alkaline magmatism over a 15 million year interval suggests that the southern Superior Province experienced a late-tectonic, cratonization event possibly associated with terrain accretion to the south.

The Falcon Island plutonic complex is an ultrapotassic end-member of the late-tectonic potassic suite. Its magma evolved with increasing $P(H_2O)$, as indicated by the transition from early crystallization of clinopyroxene to amphibole and biotite, and from an early hypersolvus texture to a later subsolvus assemblage. An "aegirine-enrichment trend" is evident in the clinopyroxene as progressive depletion in Mg and Ca, and enrichment in Fe and Na (both as zoning in individual grains and in the successively fractionated units) (Figs. 3.3, 3.4 and 3.5). Mg-rich diopside in clinopyroxenite was followed by edenite and phlogopite. In monzodiorite, diopside was followed by biotite and normally zoned plagioclase (An_{32} to An_{20}). In alkali feldspar melasyenite and alkali feldspar syenite, Fe-rich diopside cores evolved to aegirine augite and hastingsite rims and was followed by albite ($An_{1.6}$ to $An_{0.7}$). Elevated Al is evident in whole rock compositions and is reflected in the Al content of clinopyroxene, amphibole and biotite, characteristics which are typical of the Group-III ultrapotassic suites from modern collisional orogenic areas (Foley et al. 1987). Based on its mineral assemblage, and those of metasedimentary enclaves, the Falcon Island pluton was emplaced at mid-crust. The relatively

elevated ^{VI}Al -content of the clinopyroxene is similar to that of the alkaline Murdock Creek pluton in the Kirkland Lake camp of the Abitibi Subprovince, emplaced at a similar crustal depth (Rowins et al. 1991). However, differences in Al evolution with fractionation is evident in the biotite trends of the two intrusions. Biotite at Murdock Creek shows a depletion in ^{IV}Al at constant $\text{Fe}/(\text{Fe} + \text{Mg})$, whereas at Falcon Island the biotite displays increasing ^{IV}Al and $\text{Fe}/(\text{Fe} + \text{Mg})$ with fractionation. The differing trends are probably the result of a variety of factors including: differing $f(\text{O}_2)$ paths, differing relative ASI values, the types and relative abundance of precipitating minerals, and differing crustal contamination sources.

The redox conditions of alkaline complexes may have economic significance. Late Archean alkali-rich intrusive rocks have been implicated in models for gold mineralization on the basis of their close spatial and temporal association (Rock & Groves 1988, Wyman & Kerrich 1989, Sutcliffe et al. 1990, Robert 1997). Fyon et al. (1989) suggested that gold mineralization and alkaline magmatism may be a manifestation of the cratonization process involving mantle-derived melts and CO_2 . Cameron & Hattori (1987) have proposed that gold mineralization in the Kirkland Lake camp originates from oxidized hydrothermal fluids derived from magmatic sources. Many of the syenites associated with mineralization in the Kirkland Lake gold camp represent oxidized magmas with possible links to the gold-bearing fluids (Cameron and Carignan 1987; Rowins et al. 1991, Levesque 1994),

Gold mineralization environments in the Lake of the Woods greenstone belt have both porphyry and shear-zone associations (Davis & Smith 1991). There is, however, no evidence of anomalous gold mineralization in the vicinity of the Falcon Island pluton, despite its alkaline nature, and the proximity of potentially favorable shear-zones (Ayer 1991). This fact suggests a negative correlation between gold mineralization and reduced alkaline plutons. The Falcon

Island complex is comparable in array of rock types and depth of emplacement to syenitic complexes associated with gold mineralization in the Kirkland Lake camp, but there are significant differences in the redox state of the evolving magmas. It would therefore appear that alkaline complexes with progressively reduced magmatic fluids, such as the Falcon Island complex, have less potential to produce the oxidized hydrothermal fluids conducive to gold mineralization. Thus analysing the composition of ferromagnesian minerals in a differentiated complex, and in particular the biotite, may provide an additional tool for evaluating their potential to produce gold deposits.

Nd isotopes of the igneous rocks of the LWGB greenstone belt yield a crustal “isochron” of 2754 ± 65 Ma with an initial $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.509146 ± 31 and $\epsilon_{\text{Nd}} = +1.7$. The similarity of this age (within error) to independent U-Pb zircon ages of 2.70 to 2.74 Ga from the individual units demonstrates closure of this system since the Neoarchean. Agreement with the isochron for similar aged rocks in the Rainy Lake greenstone belt (2737 ± 42 Ma, $^{143}\text{Nd}/^{144}\text{Nd} = 0.509178 \pm 33$ and $\epsilon_{\text{Nd}} = +1.9$; Shirey and Hanson, 1986) suggests a broad-scale continuity of Neoarchean mantle sources over a 150 km wide section across the western Wabigoon Subprovince. The overall similarity of positive ϵ_{Nd} values for tholeiites, komatiites, and calc-alkaline volcanic rocks (+1.1 to +2.1) of the Lower and Upper Keewatin assemblages indicates they were derived from isotopically homogeneous mantle with a time averaged Sm/Nd depletion relative to CHUR. Matching ϵ_{Nd} values in similar aged tonalite of the Aulneau Batholith (+1.5; Beakhouse and McNutt, 1991) suggests a short residence time for crustal recycling of partial melts of depleted lower crust.

A long time averaged Sm/Nd enriched source is evident in the ϵ_{Nd} value of -0.9 for wacke within the upper stratigraphic part of the Upper Keewatin Assemblage. Pronounced negative ϵ_{Nd} values also occur in the Lount Lake batholith (avg. -1.9) within the interior of the Winnipeg River Subprovince (Beakhouse and McNutt, 1991). These values in conjunction with a wide range of detrital zircon ages from 2.7 to 3.0 Ga in the turbidite sequences of the western Wabigoon (Davis 1996), indicate depositional mixing between an enriched “ancient crustal” source such as the Winnipeg River Subprovince and local sources dominated by the isotopically depleted juvenile igneous rocks of the greenstone belt.

Heterogeneity in the post 2.7 Ga mantle is demonstrated by low Nd values in the shoshonitic and calc-alkaline volcanic rocks of the Electrum Assemblage (-0.4 & $+0.4$) and depleted values in the Falcon Island pluton ultrapotassic rocks ($+1.8$ to $+3.1$). Asymptotic increase in ϵ_{Nd} values with fractionation in the pluton suggest mixing between mantle components with variable time averaged Sm/Nd depletion. Mantle mixing is also supported by a “mixing” isochron with an age of 2418 ± 150 Ma, significantly younger than the U-Pb crystallization age of 2695 ± 3 Ma; an initial ϵ_{Nd} of -1.2 (initial $^{143}Nd/^{144}Nd = 0.509440 \pm 102$), significantly lower than the ϵ_{Nd} values of the individual samples of the pluton; and T_{DM} ages which are similar to the U-Pb age of the pluton.

Pb isotopes of leached alkali feldspars from a number of units of the Falcon Island pluton plot on a $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ isochron with an age of 2700 ± 100 Ma. The isochron plots well below the Stacey and Kramers (1975) crustal evolution curve and has an initial $^{207}Pb/^{204}Pb$ of 14.52 and an initial $^{206}Pb/^{204}Pb$ of 13.29, corresponding to derivation from a 4.55 to 2.70 Ga single stage geochron reservoir with $\mu_1 = 7.86$. This indicate a mantle Pb

source slightly more enriched in $^{207}\text{Pb}/^{204}\text{Pb}$, but similarly depleted in $^{206}\text{Pb}/^{204}\text{Pb}$, than the source of mafic and ultramafic magmas in the Wabigoon and Abitibi subprovinces.

Comparison of the Falcon Island Nd and Pb isotope data with other well documented potassic suites supports the suggestion that the post 2.7 Ga southern Superior Province mantle was isotopically heterogeneous. The data is correlative along an asymptotic curve that suggests mixing between a pre 2.7 Ga depleted mantle component and an enriched component with isotopic characteristics similar to those of the ancient crustal rocks such as the gneissic and plutonic suites of the Winnipeg River, Opatika and Pontiac subprovinces. This type of heterogeneity is most readily explained by metasomatism of a mantle wedge by the influx of fluids derived by partial melting of ancient crustal sediments subducted beneath the Neoproterozoic crust.

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Appendix

Appendix I. Analytical procedures for major and trace element lithogeochemical analysis

Analyses were performed by the Geoscience Laboratories of the Ontario Geological Survey. Major elements were determined by X-ray fluorescence (XRF) on fused discs. Loss on ignition (LOI) was by gravimetric methods. Trace elements were determined by XRF and Inductively Coupled Plasma-Optical Emission spectroscopy (ICP/OES), the rare earth elements (REE) were analyzed by Inductively Coupled Plasma-Mass Spectroscopy (ICP/MS). Duplicates and standards submitted with the samples indicate precision and accuracy values were generally within 5% (at 2σ) for values greater than 10 times detection limits.

Selected samples were also analyzed by Instrumental Neutron Activation for Sc, Cr, La, Ce, Nd, Sm, Eu, Tb, Dy, Ho, Yb, Lu, Hf, Ta, Th and U. For all of the above elements, with the exception of Dy, the samples were irradiated for 2 hours in a SLOWPOKE reactor at an energy flux of 5×10^{11} neutrons/cm/sec, let stand for differing periods determined as optimum for the specific element being measured and then measured for a period of 2 hours. The predicted accuracy of the INA technique is from 5-10% for most of the REE. Ho is somewhat more difficult to detect and has a predicted accuracy of 10-15%. Dy, having a very short half-life of 2.33 hours requires a different procedure involving irradiation with a flux of 1×10^{11} neutrons/cm/sec, a wait period of 20 minutes and a counting time of 1 minute. Blind duplicates submitted with samples indicate precision is generally within 10%, except for Cr, Ta and U, which can be up to 30% off at low concentrations levels. Blind analysis of USGS reference standard, BCR-1, demonstrates a very good accuracy of within 5% of the recommended values

for all analyzed elements with the exception of a 10% accuracy for Eu and a 20% accuracy for Cr. International reference standards SY-2 and NIM-G were generally within 10% of the recommended values, with the exception of considerably lower accuracy for Cr. However, Cr was accurately analyzed by XRF.

Appendix II. Analytical procedures for Sm-Nd isotopes

Samarium and neodymium concentrations and isotope ratios for the LWGB greenstone belt samples (Table 5.1) were determined at the Geoservices Laboratory of the Ministry of Northern Development and Mines, Sudbury, Ontario. Approximately 50 to 200 mg of -170 mesh rock powder and the appropriate amount of a mixed ^{148}Nd - ^{149}Sm spike are sealed into SavillexTM bombs and dissolved in an oven at 125°C. The sample is then dried and redissolved in 1 ml of 2.5 N HCl and loaded onto a cation exchange column which allows the separation of Rb, Sr and the REE. The REE aliquot is evaporated and redissolved in 0.25 ml of 0.18N HCl and loaded onto a anion exchange column of Biobeads coated with a phosphate activator (HDEHP). Nd is eluted using 0.18N HCl, followed by Sm in 0.35N HCl.

Each sample separate is loaded onto a clean, degassed Ta side filament in 1 μl of an acid solution. The filament block is set with a Re centre filament in a triple filament configuration. The Nd sample is analysed in dynamic collector mode in a VG Sector 54 Thermal Ionization Mass Spectrometer.

Total procedural blanks are 0.5 ng or better for each of Nd and Sm. $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic analyses are measured to $\pm 0.0015\%$ (2σ). Actual reproducibility is considered to be $\pm 0.0020\%$ (2σ) or better, based on reproducibility of USGS rock standard BCR-1 ($n = 7$) which

yields Sm = 6.61 ± 0.46 ppm, Nd = 28.74 ± 1.65 ppm and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512643 \pm 15$, and an internal rock standard (MRB-29). Concentrations are precise to $\pm 2\%$, but $^{147}\text{Sm}/^{144}\text{Nd}$ ratios are reproducible to 1%. Seven runs of the La Jolla Nd standard average $^{143}\text{Nd}/^{144}\text{Nd} = 0.511839 \pm 19$ are within error of the literature value of 0.511860 ± 10 (Shirey & Hanson 1986) so no correction was applied.

The Falcon Island pluton samples (Table 5.1) were determined by isotope dilution procedures at Carleton University, Ottawa, Ontario. Whole rock powder was dissolved in 50% HF-12N HNO₃, 8N HNO₃ and finally 6N HCl. The residue was taken up in 2.5 N HCl and the rare earth elements (REE) are separated in a Rb-Sr chromatography column. The REE solution is dried and the residue is dissolved in 0.26N HCl and loaded into a 10-ml borosilicate glass chromatographic columns containing a 2 cm-high bed of Teflon powder coated with orthophosphoric acid. Nd is eluted using 0.26N HCl, followed by Sm in 0.5N HCl.

Total procedural blanks for Nd are < 75 picograms. Samples are spiked with a mixed ^{148}Nd - ^{149}Sm spike prior to dissolution. Concentrations are precise to $\pm 2\%$, but $^{147}\text{Sm}/^{144}\text{Nd}$ ratios are reproducible to 1%. Samples are loaded with 0.25N HCl on one side of a Re double filament, and run at temperatures of 1780-1820°C. Isotope ratios are normalised to $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$. Periodic analyses of the USGS standard BCR-1 have yielded Nd = 29.02 ppm, Sm = 6.68 ppm and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512668 \pm 20$ (n = 4). Thirty four runs of the La Jolla standard average $^{143}\text{Nd}/^{144}\text{Nd} = 0.511875 \pm 18$ (Sept., 1992- July, 1996).

Appendix III. Analytical procedures for Pb isotopes

Lead isotope ratios were determined at Carleton University. Alkali feldspar grains from 4 samples of the Falcon Island pluton were hand picked by binocular microscope after crushing and separation using a Frantz magnetic separator. The selected grains were acid washed to remove the outer layer of the feldspar and surface contaminants, and to preferentially remove as much labile radiogenic Pb in the feldspar structure as possible. The fractions were washed in 6N HCl for one hour at 120°C, then partially dissolved in a mix of HBr and HF for one hour. The remaining feldspar grains were dissolved in a mixture of 50% Hf-12N HNO₃, then attacked with 8 N HNO₃ and finally 6N HCL. The residue was taken up in 1N HBr. Pb is separated in Bio-Rad 10 ml polyethylene columns and Dowex AG2-8X anion resin, using 1N HBr to elute other elements and 6N HCL to elute Pb. The collected Pb solution is dried, redissolved in 1N HBr, and the above procedure is repeated with a small volume resin bed. Total procedural blanks for Pb are < 150 picograms. The samples were run using a Finnigan Mat 261 thermal ionization mass spectrometer using the static mode of analysis. Samples were loaded onto single Re filaments with H₃PO₄ and silica gel, and are run at filament temperatures of 1250-1350 °C. Analysis of USGS Standard BCR-1 yields $^{206}\text{Pb}/^{204}\text{Pb} = 18.818$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.633$, and $^{208}\text{Pb}/^{204}\text{Pb} = 38.633$ (average of two runs, see Tatsumoto et al. 1972). All mass spectrometer runs are corrected for fractionation using NIST SRM981. The average ratios measured for SRM981 are $^{206}\text{Pb}/^{204}\text{Pb} = 16.890 \pm .012$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.429 \pm .014$, and $^{208}\text{Pb}/^{204}\text{Pb} = 386.502 \pm .048$ (2 σ), based on 38 runs between September 1992 and February 1996. The fractionation correction, based on the values of Todt et al. (1984) is + 0.13%/amu.

Appendix IV. Sample descriptions

Sample numbers, rock names and description, stratigraphic unit / intrusion affiliation for the sample locations indicated on Figure 5.1.

- 1) 87JAA-0839: Magnesium tholeiitic basalt; pillowed, plagioclase megaphyric mafic flow; Snake Bay Group, Lower Keewatin Assemblage.
- 2) 93JAA-0023: Magnesium tholeiitic basalt; massive, fine-grained massive mafic flow; Bigstone Bay Group, Lower Keewatin assemblage.
- 3) 93JAA-0036: Calc-alkaline andesite; pillowed, plagioclase microlitic mafic flow, Clearwater Bay Group, Upper Keewatin Assemblage.
- 4) 93JAA-0037: Iron tholeiitic basalt; pillowed mafic flow, Indian Bay Group, Upper Keewatin Assemblage.
- 5) 93JAA-0018: High silica rhyolite (FII type); fine-grained, laminated, feldspar and quartz porphyritic flow; Zigzag Island Formation, Clearwater Bay Group, Upper Keewatin Assemblage.
- 6) 93JAA-0029: Calc-alkaline rhyodacite; felspar and quartz crystal tuff, Sunset Channel Formation, Long Bay Group, Upper Keewatin Assemblage.
- 7) 93JAA-0030: Peridotitic komatiite; pillowed ultramafic flow, Cliff Island Formation, Long Bay Group, Upper Keewatin Assemblage.
- 8) 93JAA-0026: Basaltic komatiite; fine-grained, massive ultramafic flow, Cochrane Island Formation, Monument Bay Group, Upper Keewatin Assemblage.

- 9) 93JAA-0027: Evolved tholeiitic basalt (fractionated REE patterns): pillowed, clinopyroxene- and plagioclase-phyric mafic flow. Bishops Bay Formation, Monument Bay Group, Upper Keewatin Assemblage.
- 10) 89JAA-1116: Wacke; fine-grained and biotitic with garnet porphyroblasts; Long Bay Group, Upper Keewatin Assemblage.
- 11) 93JAA-0017: Shoshonitic andesite; fine-grained, clinopyroxene-phyric, massive mafic flow; Crowduck Lake Group, Electrum Assemblage.
- 12) 93JAA-0035: Calc-alkaline rhyodacite; Fine-grained quartz and feldspar crystal tuff; White Partridge Bay Group, Electrum Assemblage.
- 13) 88JAA-0182: Biotite clinopyroxenite; Falcon Island pluton.
- 14) 88JAA-0190: Biotite-clinopyroxene syenite; Falcon Island pluton.
- 15) 88JAA-0195: Biotite-clinopyroxene monzodiorite; Falcon Island pluton.
- 16) 88JAA-0261: Biotite-muscovite syenite; Falcon Island pluton.
- 17) 93JAA-0031: Biotite-clinopyroxene syenite; Falcon Island pluton.
- 18) 93JAA-0032: Biotite-clinopyroxene melasyenite; Falcon Island pluton.
- 19) 93JAA-0033: Biotite-clinopyroxene melasyenite; Falcon Island pluton.

LAKE OF THE WOODS SYNOPTIC MAP (1:50 000 SCALE)

LEGEND

PHANEROZOIC

CENOZOIC

QUATERNARY

PLEISTOCENE AND RECENT

Glacial, glacial-fluvial, bog, lake, and
stream deposits.

UNCONFORMITY

PRECAMBRIAN

PROTEROZOIC

MAFIC INTRUSIVE ROCKS

11a Diabase.

11b Diorite.

INTRUSIVE CONTACT

ARCHEAN

POST-TECTONIC FELSIC INTRUSIVE ROCKS (massive-textured)

10a Potassium feldspar porphyritic biotite granite.

10b Potassium feldspar porphyritic biotite
granodiorite.

10c Biotite granite.

10d Biotite granodiorite.

10e Feldspar +/- quartz porphyry.

10f Felsite, aplite.

10g Pegmatite.

INTRUSIVE CONTACT

LATE TECTONIC INTRUSIVE ROCKS (foliated to massive)

**Quartz Saturated Intermediate to Ultramafic Intrusive
Rocks**

9a Hornblende +/- biotite syenite, monzonite.

9b Hornblende +/- biotite monzodiorite, diorite.

9c Hornblende +/- clinopyroxene +/- biotite
monzogabbro, gabbro.

9d Hornblendite, pyroxenite.

9e Amphibole +/- clinopyroxene +/- biotite
lamprophyre.

INTRUSIVE CONTACT

CALC-ALKALINE SERIES METAVOLCANIC ROCKS^b

Felsic to Intermediate Metavolcanic Rocks

- 3 Unsubdivided.
- 3a Massive flows.
- 3b Laminated flows.
- 3c Pillowed flows.
- 3d Amygdaloidal flows.
- 3e Brecciated Flows.
- 3f Tuff, bedded tuff.
- 3g Lapilli-tuff, tuff-breccia.
- 3h Pyroclastic-breccia.

Intermediate to Mafic Metavolcanic Rocks

- 2 Unsubdivided.
- 2a Massive flows.
- 2b Pillowed flows.
- 2c Pillow breccia, flow breccia.
- 2d Amygdaloidal flows.
- 2e Tuff, bedded tuff.
- 2f Lapilli-tuff, tuff-breccia.
- 2g Pyroclastic-breccia.

THOLEIITIC TO KOMATIITIC SERIES METAVOLCANIC ROCKS^b

Mafic to Ultramafic Metavolcanic Rocks

- 1 Unsubdivided.
- 1a Fine-grained massive flows.
- 1b Medium to coarse-grained massive flows.^c
- 1c Pillowed flows.
- 1d Plagioclase megaphyric flows.
- 1e Amygdaloidal flows.
- 1f Variolitic flows.
- 1g Pillow breccia, flow breccia.
- 1h Tuff, lapilli-tuff, tuff-breccia.

^aMay be in part extrusive.

^bNumerical order does not imply age relationships among metasedimentary and metavolcanic rocks.

^cMay be in part intrusive.

METAL AND MINERAL REFERENCE

Ag.....silver	Mo.....molybdenum
Au.....gold	Pb.....lead
Cu.....copper	Sb.....antimony
fe.....feldspar	talc.....talc
mi.....mica	Zn.....zinc

Quartz Oversaturated Felsic to Intermediate Intrusive Rocks

- 8a Potassium feldspar porphyritic biotite +/- hornblende granodiorite, quartz monzodiorite.
- 8b Biotite +/- hornblende granodiorite, quartz monzodiorite.
- 8c Hornblende +/- biotite tonalite, quartz diorite.
- 8d Hornblende +/- biotite diorite.
- 8e Quartz +/- plagioclase porphyritic tonalite, trondhjemite.
- 8f Felsite, aplite.
- 8g Pegmatite.

INTRUSIVE CONTACT

PRE-TECTONIC FELSIC TO INTERMEDIATE INTRUSIVE ROCKS (gneissic to foliated)

- 7a Biotite +/- hornblende tonalite, quartz diorite.
- 7b Biotite +/- hornblende granodiorite, quartz monzodiorite.

INTRUSIVE CONTACT

METAMORPHOSED FELSIC TO INTERMEDIATE HYPABYSSAL INTRUSIVE ROCKS^a

- 6a Quartz +/- feldspar porphyry.
- 6b Feldspar porphyry.
- 6c Felsite, aplite.
- 6d Quartz and feldspar porphyritic granodiorite.

INTRUSIVE CONTACT

METAMORPHOSED MAFIC AND ULTRAMAFIC INTRUSIVE ROCKS^a

- 5 Unsubdivided.
- 5a Gabbro.
- 5b Melagabbro.
- 5c Leucogabbro, anorthositic gabbro, quartz diorite.
- 5d Quartz gabbro.
- 5e Pyroxenite, hornblendite.
- 5f Periodotite, dunite.

INTRUSIVE CONTACT

CLASTIC AND CHEMICAL METASEDIMENTARY ROCKS^b

- 4 Unsubdivided.
- 4a Wacke.
- 4b Arenite.
- 4c Siltstone, mudstone.
- 4d Heterolithic conglomerate.
- 4e Monolithic conglomerate.
- 4f Siliceous siltstone, chert.
- 4g Oxide ironstone.
- 4h Sulphide +/- graphite ironstone.

PAST PRODUCING MINES (recorded past production figures)

1. Cedar Island (4941 oz Au, 3884 oz Ag)
2. Duport (4672 Au, 1143 oz Ag)
3. Gold Hill (1090 oz Au)
4. Kenricia (2533 oz Au, 521 oz Ag)
5. Mikado (28 335 oz Au, 41 oz Ag)
6. Olympia (332 oz Au, 58 oz Ag)
7. Ophir (6089 oz Au)
8. Sultana (15 977 oz Au)
9. Wendigo (67 423 oz Au, 14 762 Ag, 1 886 246 lbs Cu)

MINERAL RESOURCES AND PRODUCTION

The map area has a long history of gold exploration with numerous occurrences discovered dating back to the 1880s. Occurrences indicated on the map face are those with anomalous assay results and a record of past exploration work or with surface indications of unrecorded past work. Table 1 lists nine of the past producing gold mines with a recorded production of greater than 100 ounces (Davies and Smith 1988). Two of the past producers, the Duport Deposit (2) and the Cedar Island Extension Deposit (1) have significant reserves. The Duport Deposit has undergone extensive underground development with a decline extending to a depth of 660 ft and total reserves estimated at 2.0 million tons averaging 0.36 ounce gold per ton, and the Cedar Island Extension Deposit has estimated reserves based on diamond drilling of 1.5 million tons averaging 0.25 ounce gold per ton (Canadian Mines Handbook 1989-90).

Gold mineralization is restricted to vein systems hosted by permeable ductile and brittle fault zones. They appear to have formed in response to the regional deformation events. Many are in close proximity to syn- to late-tectonic felsic intrusions near the greenschist-amphibolite facies transition boundary of their contact metamorphic aureoles. Detailed studies around the Canoe Lake and High Lake stocks have shown that gold mineralization appears to have formed at the same time as, but peripheral to, porphyry-style copper and molybdenum mineralization and associated potassic alteration (Smith et al. 1988).

In addition to the strong association with felsic intrusions a significant number of gold occurrences are hosted by deformation zones which transect the map area. Many of these occurrences are far removed from the contact aureoles of late-tectonic intrusions. Three factors appear to determine the locus of the more significant deformation zone occurrences, these are: (1) at abrupt dilational jogs in the trends of a deformation zones; (2) in smaller subsidiary splays off the main part of a deformation zone (Ayer and Buck 1989); and (3) where a deformation zone is

intersected at a high angle by the greenschist-amphibolite facies transition boundary (Morrice 1989)..

Copper and zinc mineralization is reported with iron sulphides as disseminations and veins in mafic metavolcanics and as disseminations in interbedded cherty and graphitic metasediments on the Eastern Peninsula. Molybdenum, copper and gold mineralization in widely spaced fractures associated with complex zonal alteration patterns is associated with the High Lake stock and Canoe Lake stock (porphyry copper-style mineralization). The Evenlode prospect on the east side of High Lake has diamond drill estimated reserves of 126 000 tons grading 0.68% MoS₂ and 0.015 ounce gold per ton (Davies and Smith 1988).

Minor iron occurs in magnetite ironstone interbedded with epiclastic metasedimentary and pyroclastic metavolcanic rocks on a small island in Andrew Bay and several small islands south of East Allie Island and sulphidic ironstone and gossans zones scattered throughout the map area.

Industrial mineral exploration consists of the evaluation of the talc content quarried from ultramafic intrusive rocks in Moore Bay and from ultramafic volcanic rocks east of Coste Island. Minor feldspar and mica production are reported from a pegmatite dike on the southwestern tip of Falcon Island.

SOURCES OF INFORMATION

Geology from published maps of the Ontario Geological Survey, Ministry of Northern Development and Mines and the Geological Survey of Canada, unpublished assessment file maps and theses maps, and supplementary mapping by J.A. Ayer in 1989 and 1990.

Assessment files, Resident Geologist's Office, Ministry of Northern Development and Mines, Kenora.

Base map derived from 1:50,000 topographic maps produced by the Canada Centre for Mapping (Ottawa), Department of Energy, Mines and Resources.

Magnetic Declination approximately 5° East, 1990.

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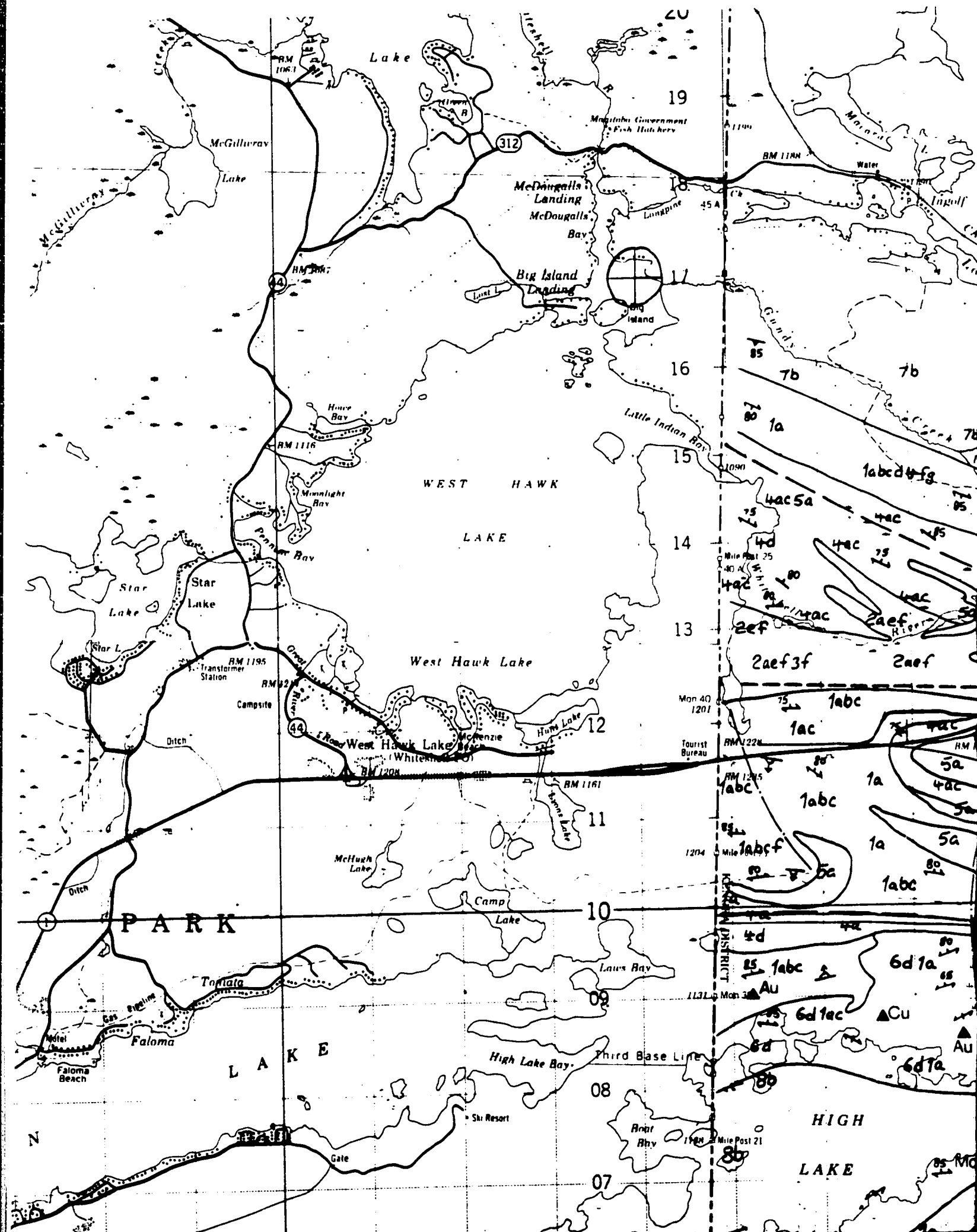
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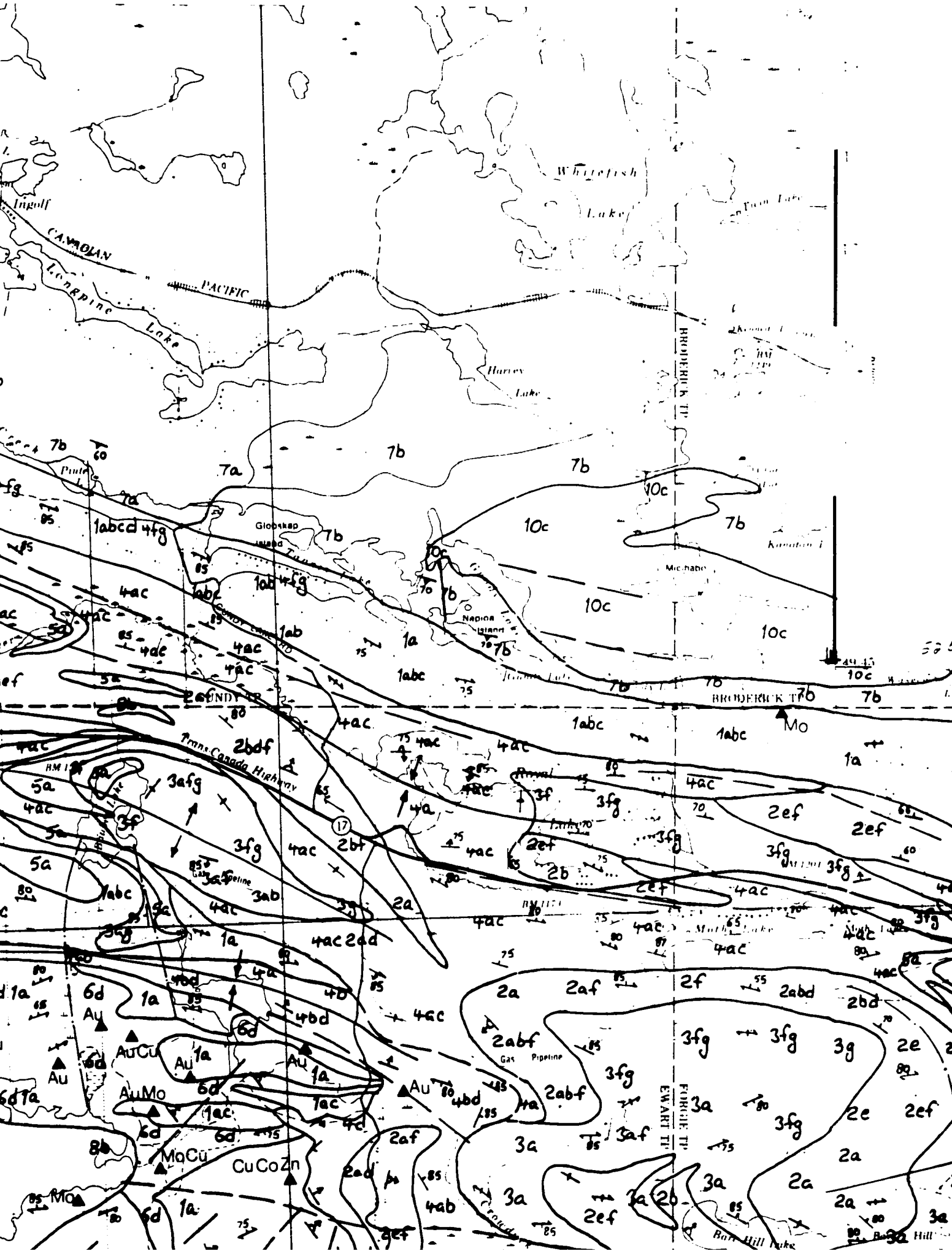
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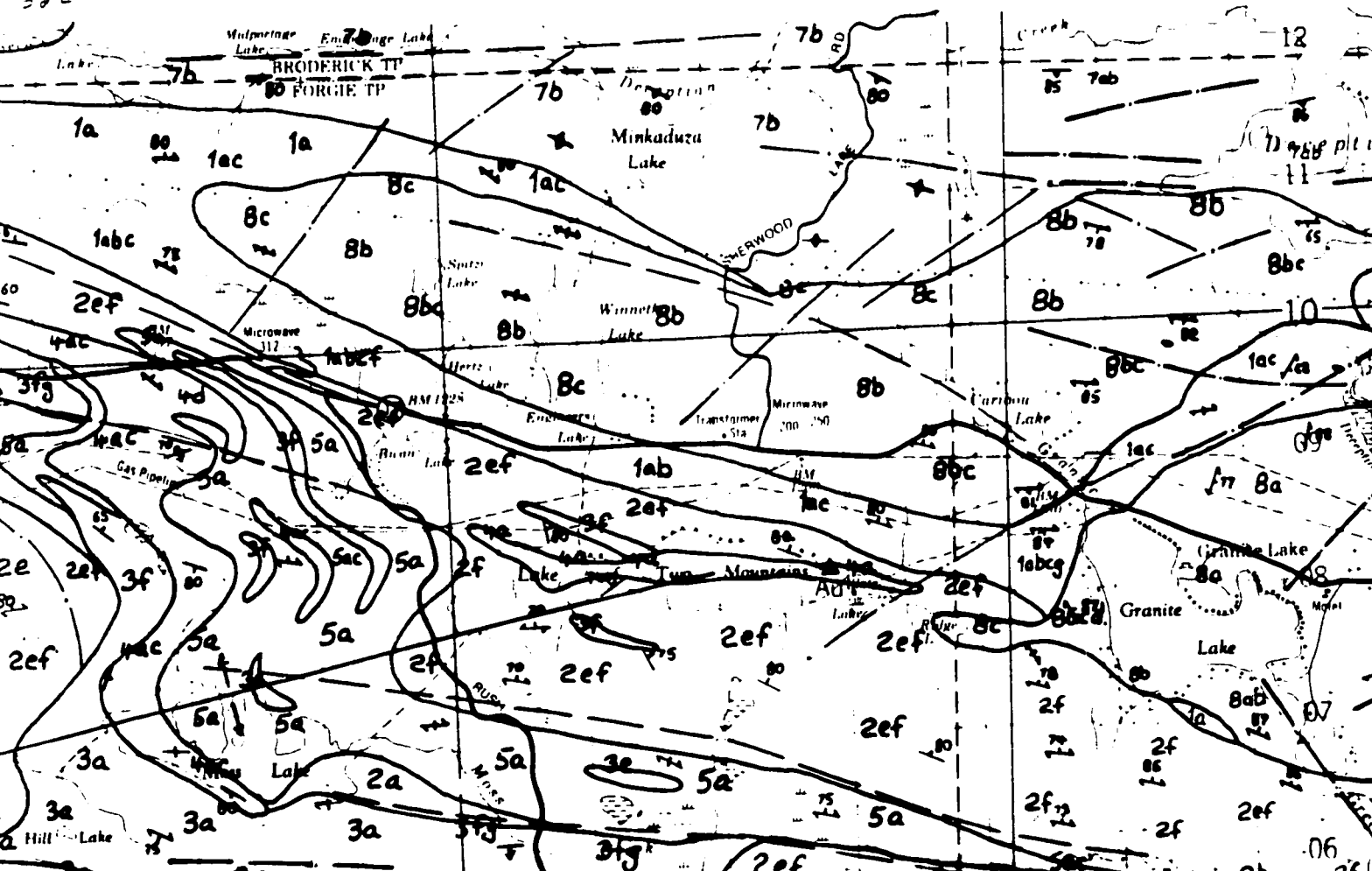
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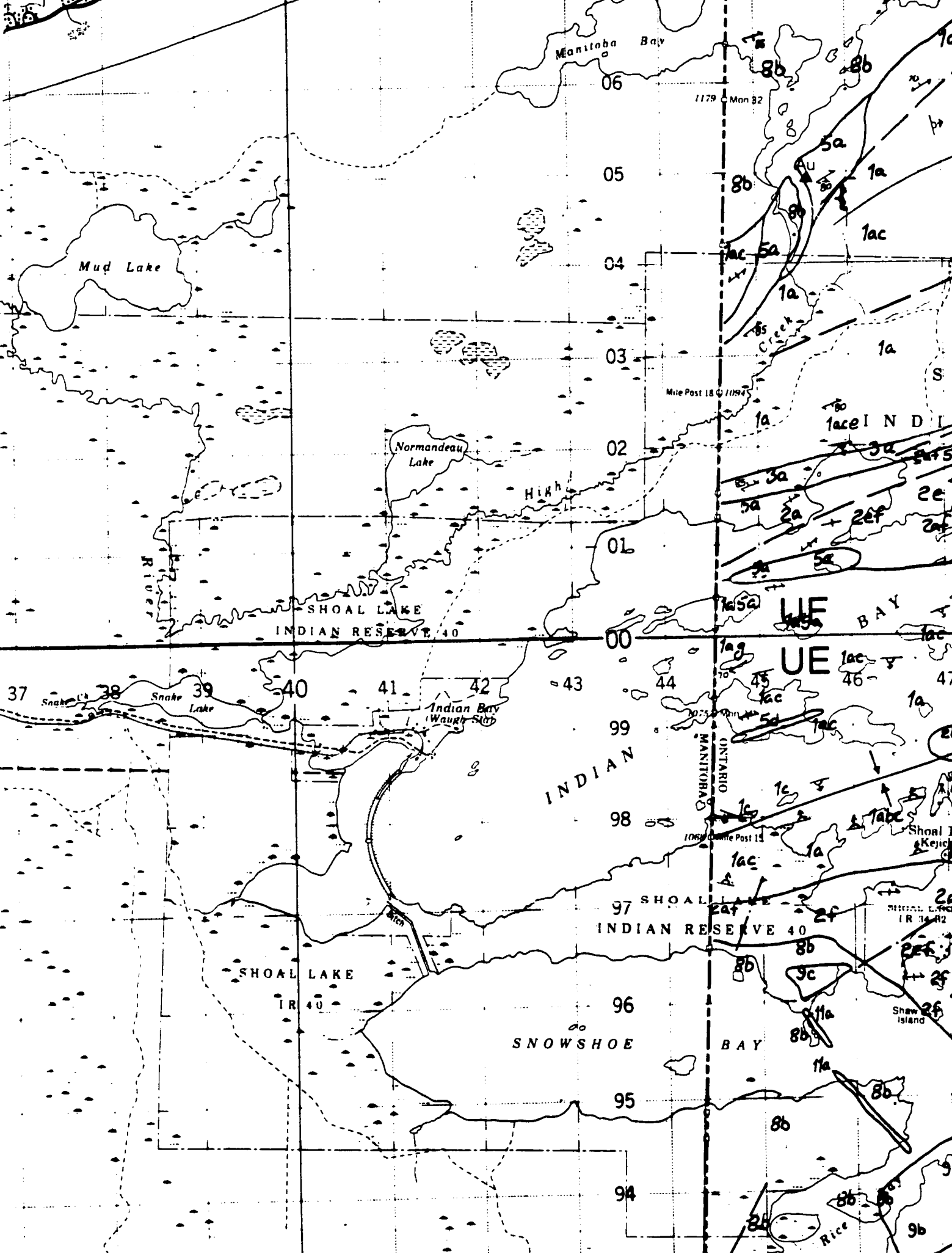


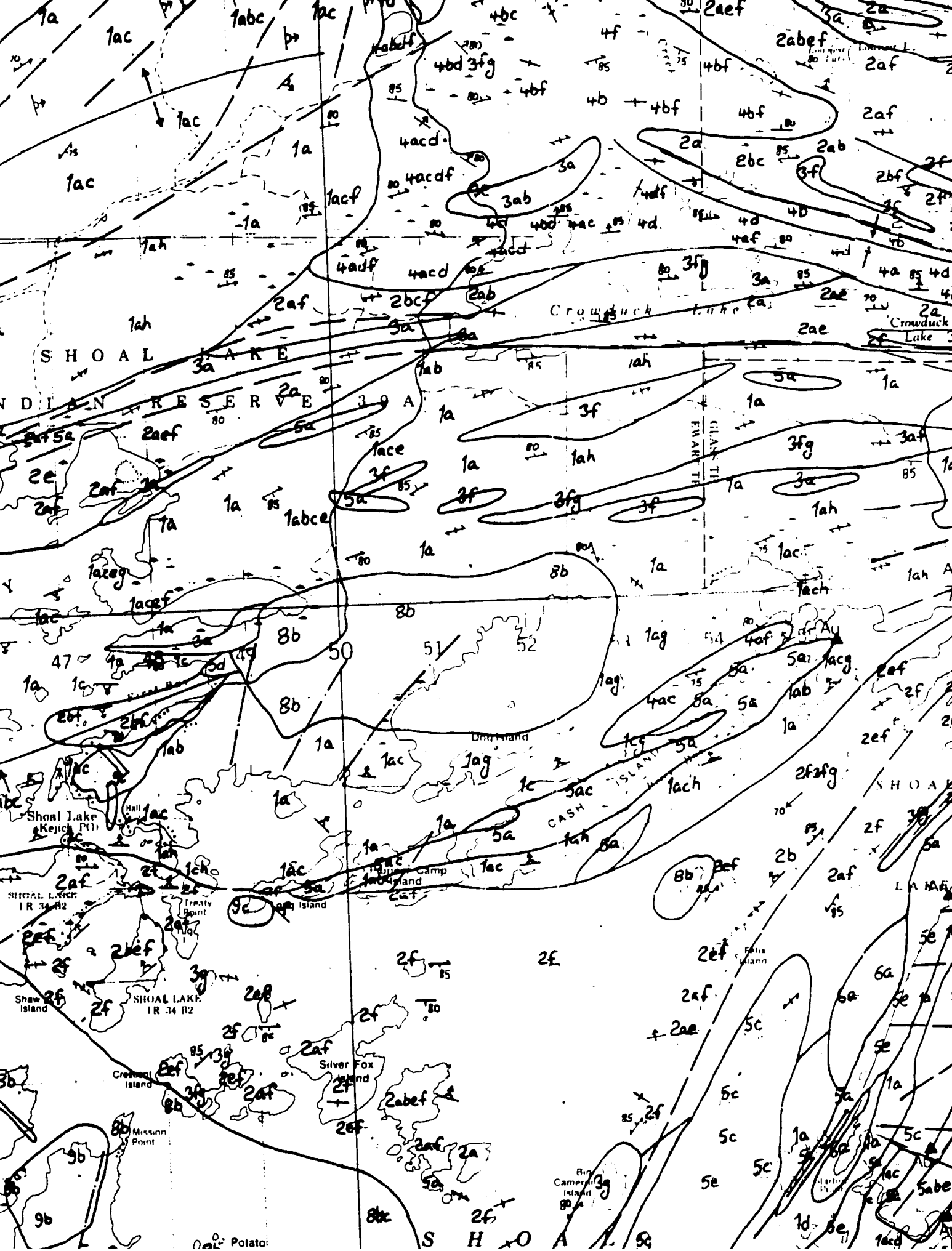


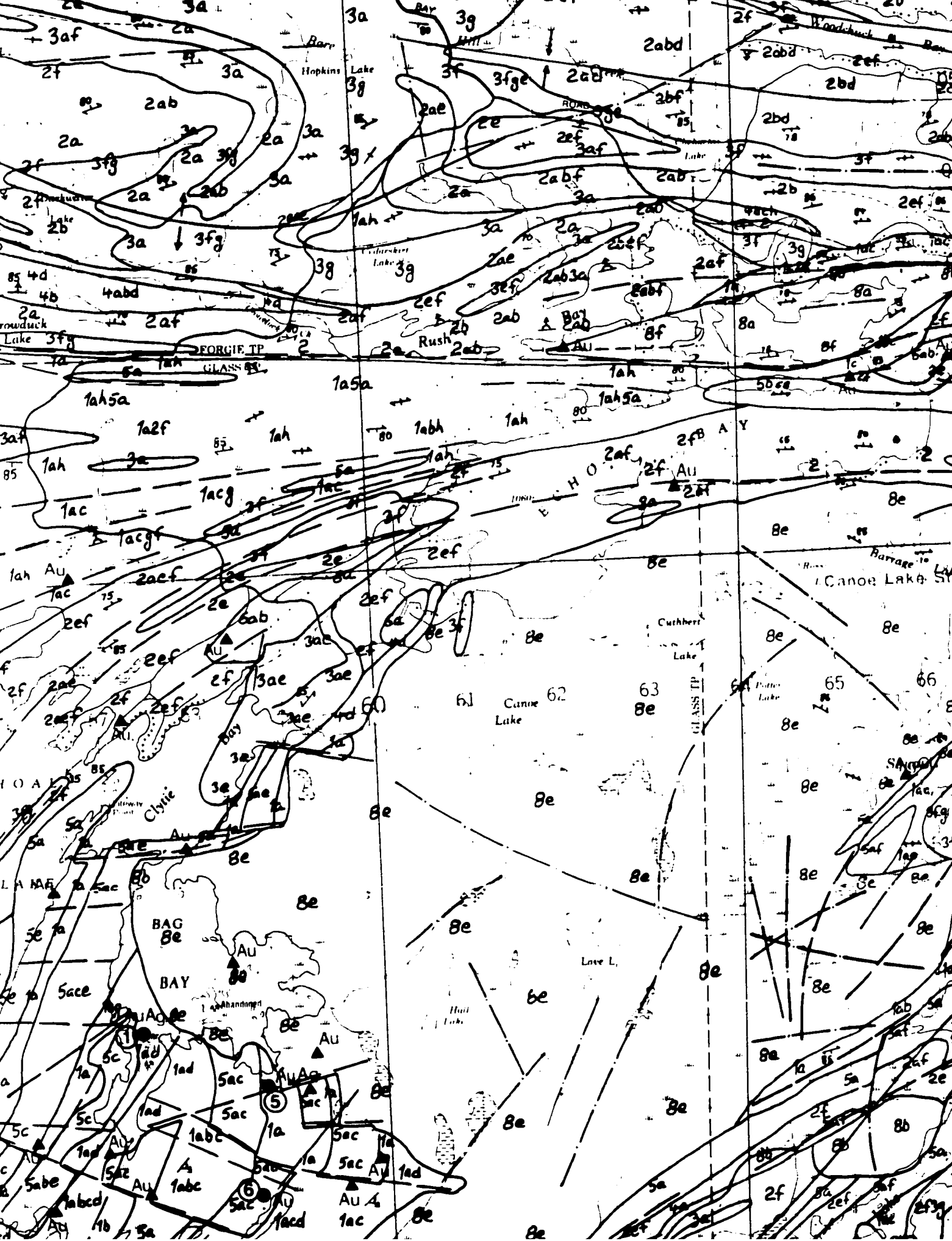
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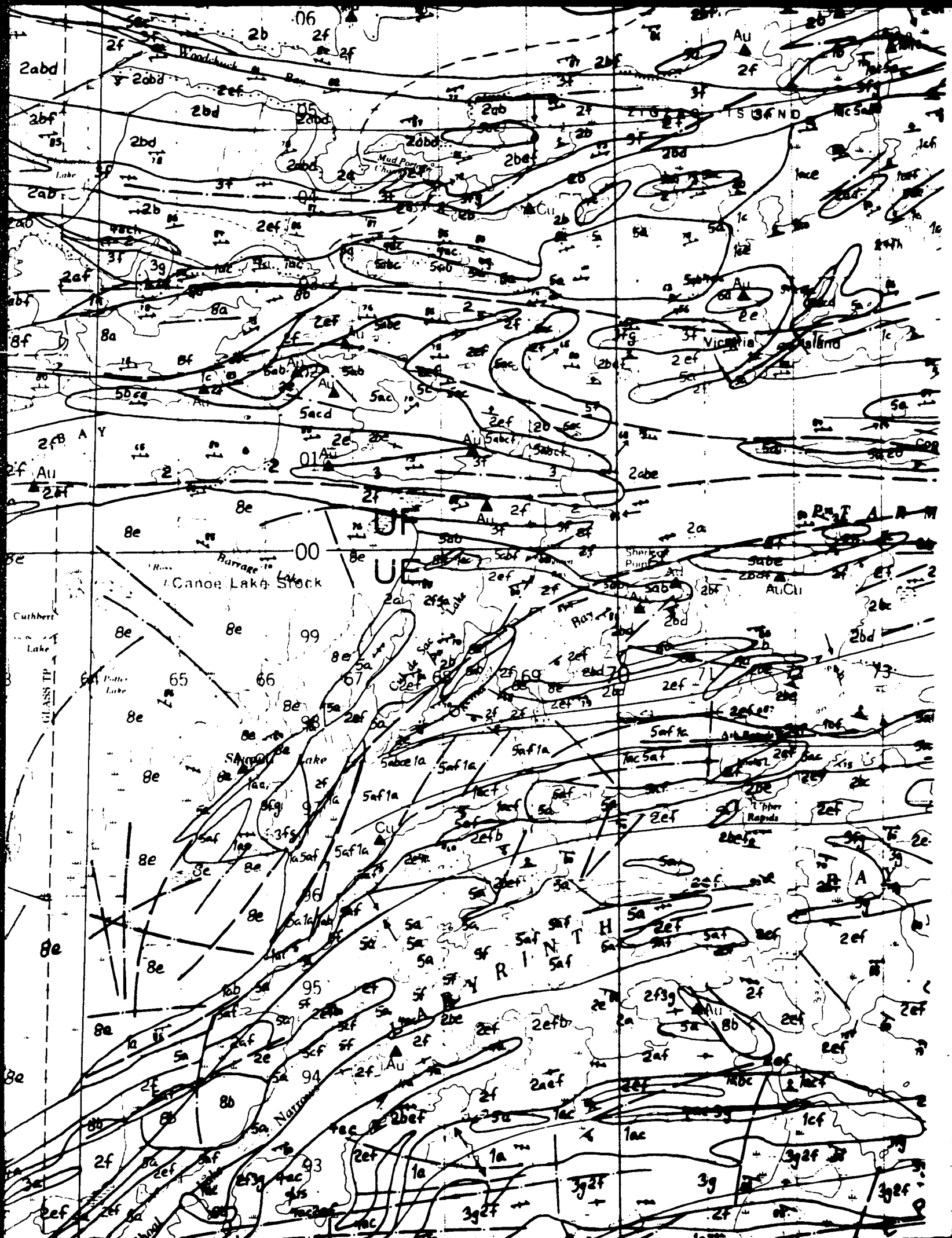


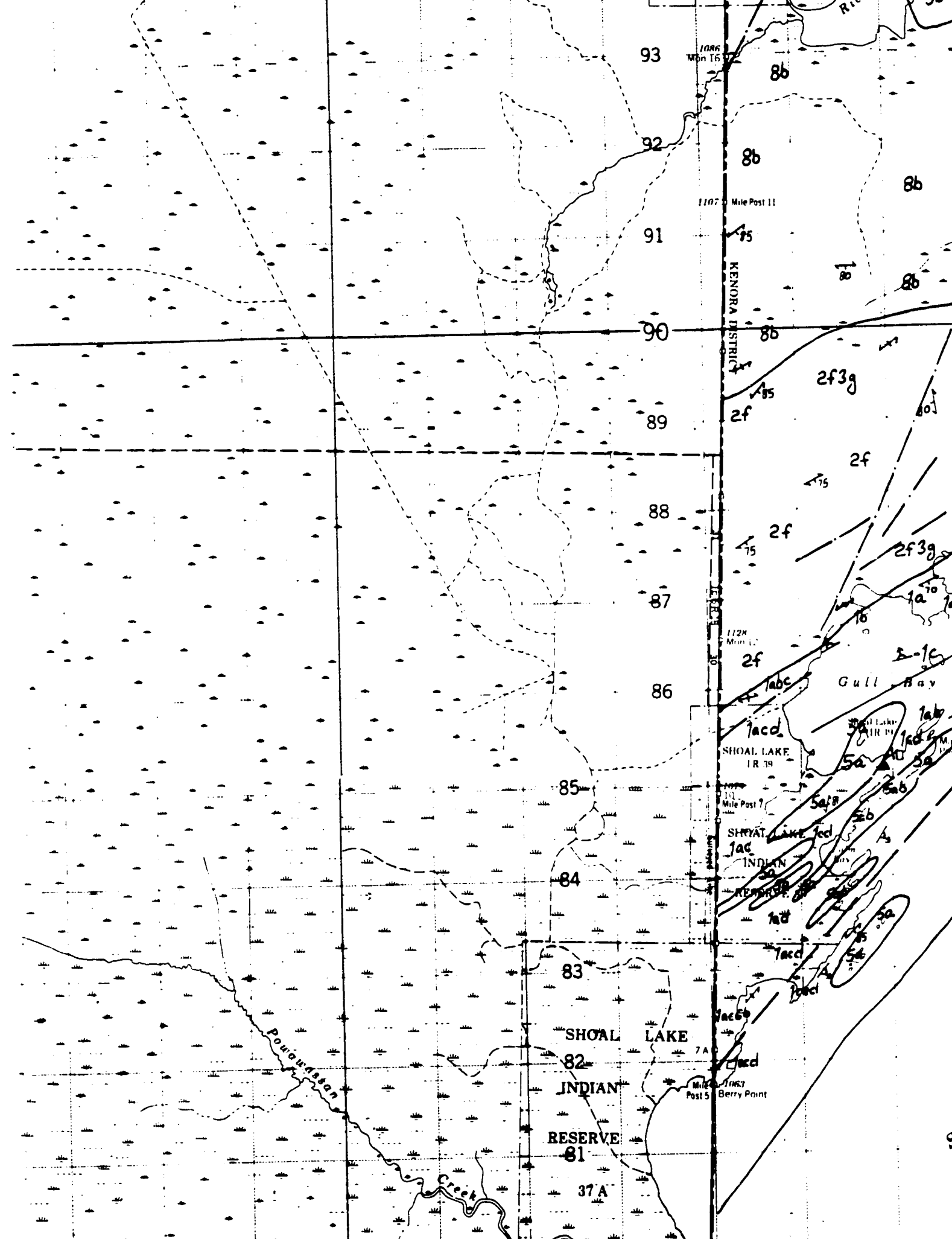


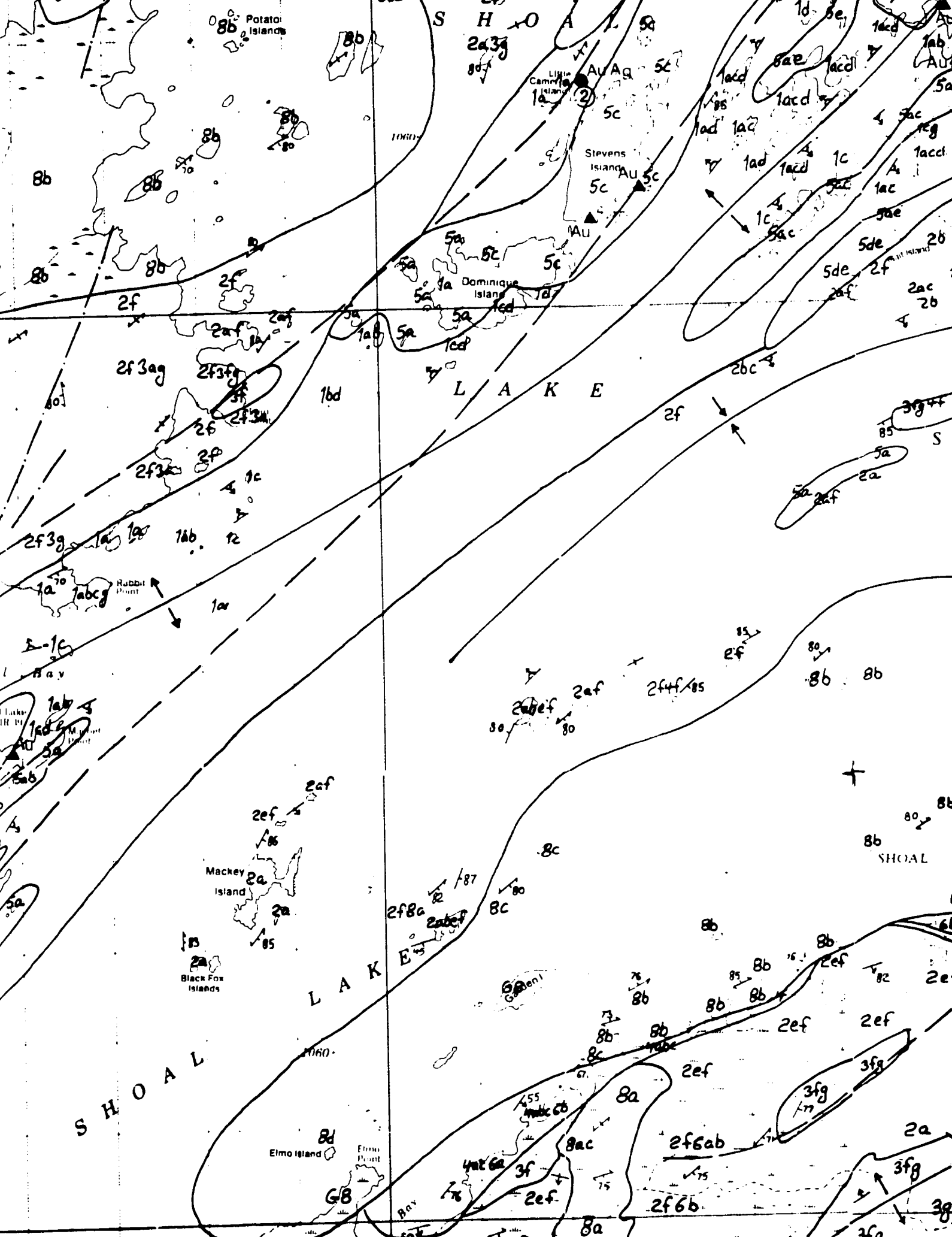


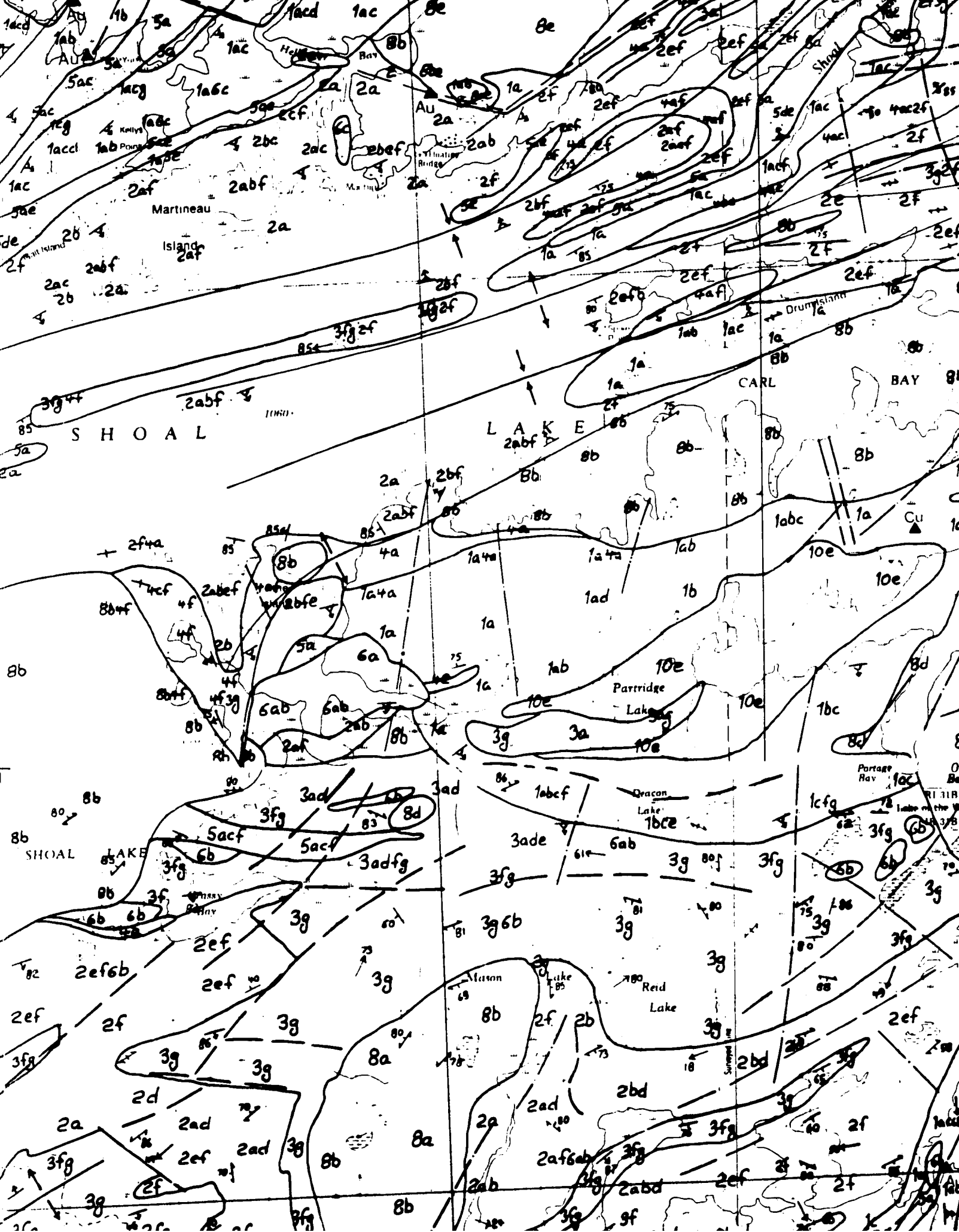


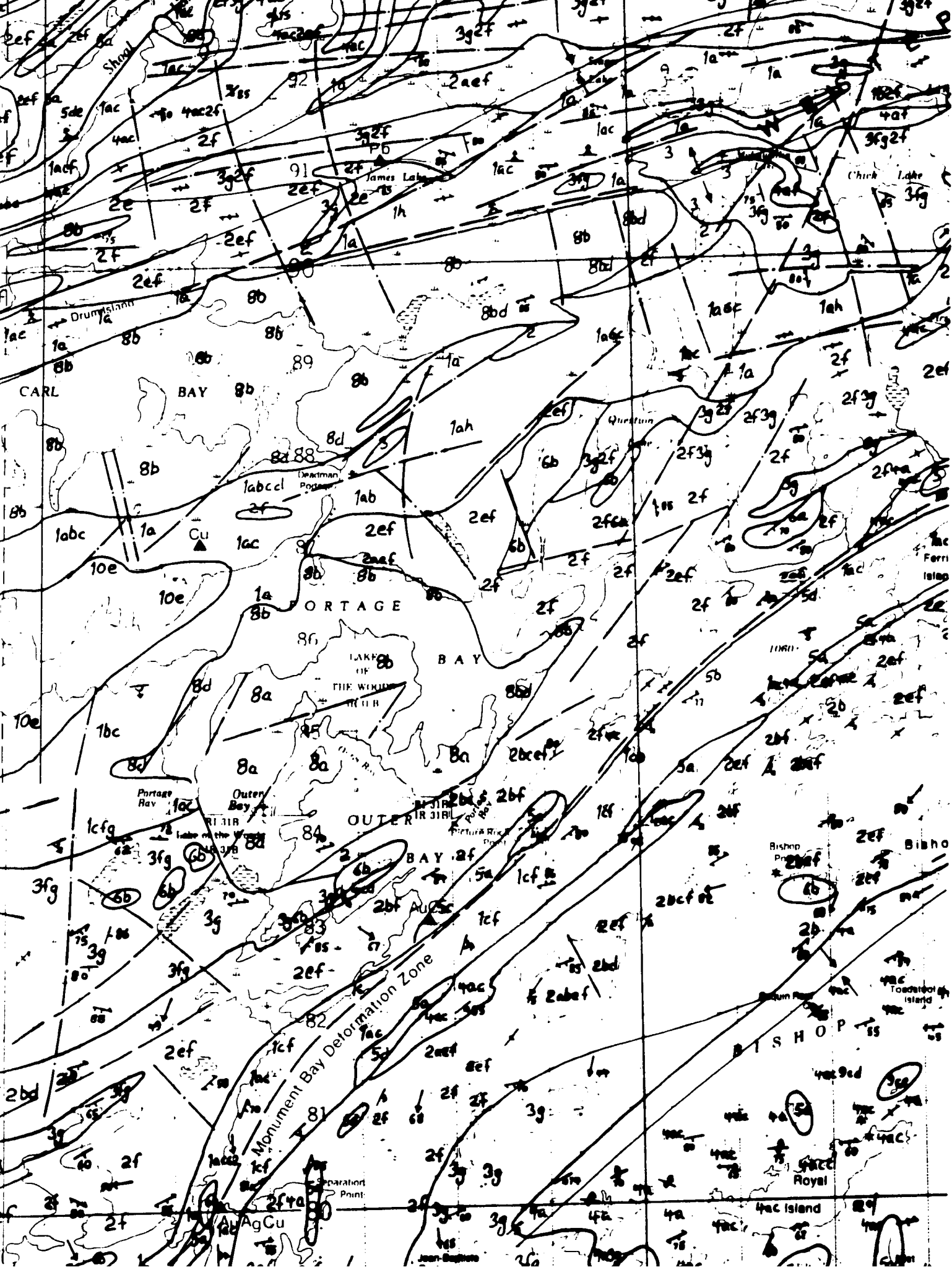


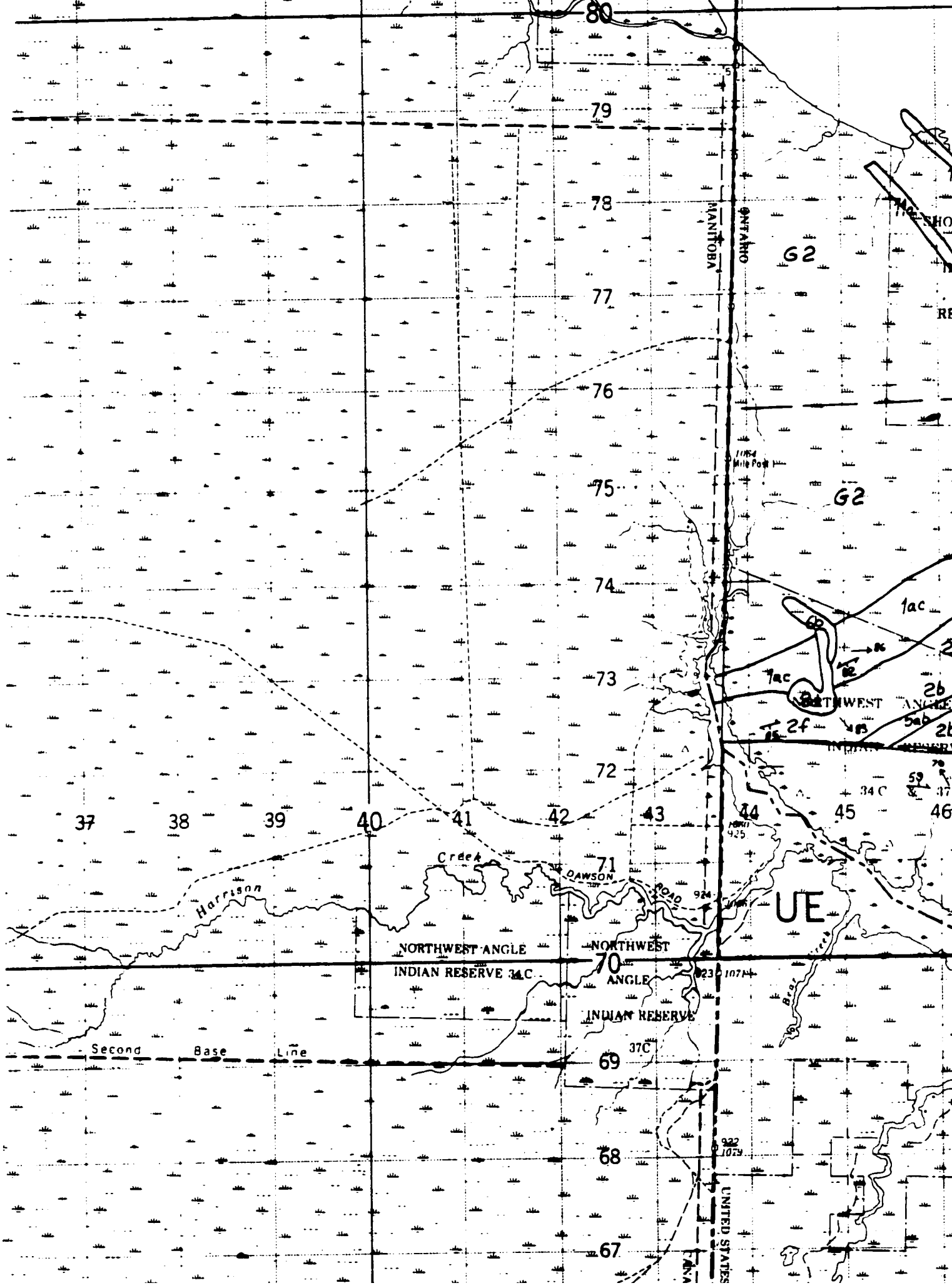


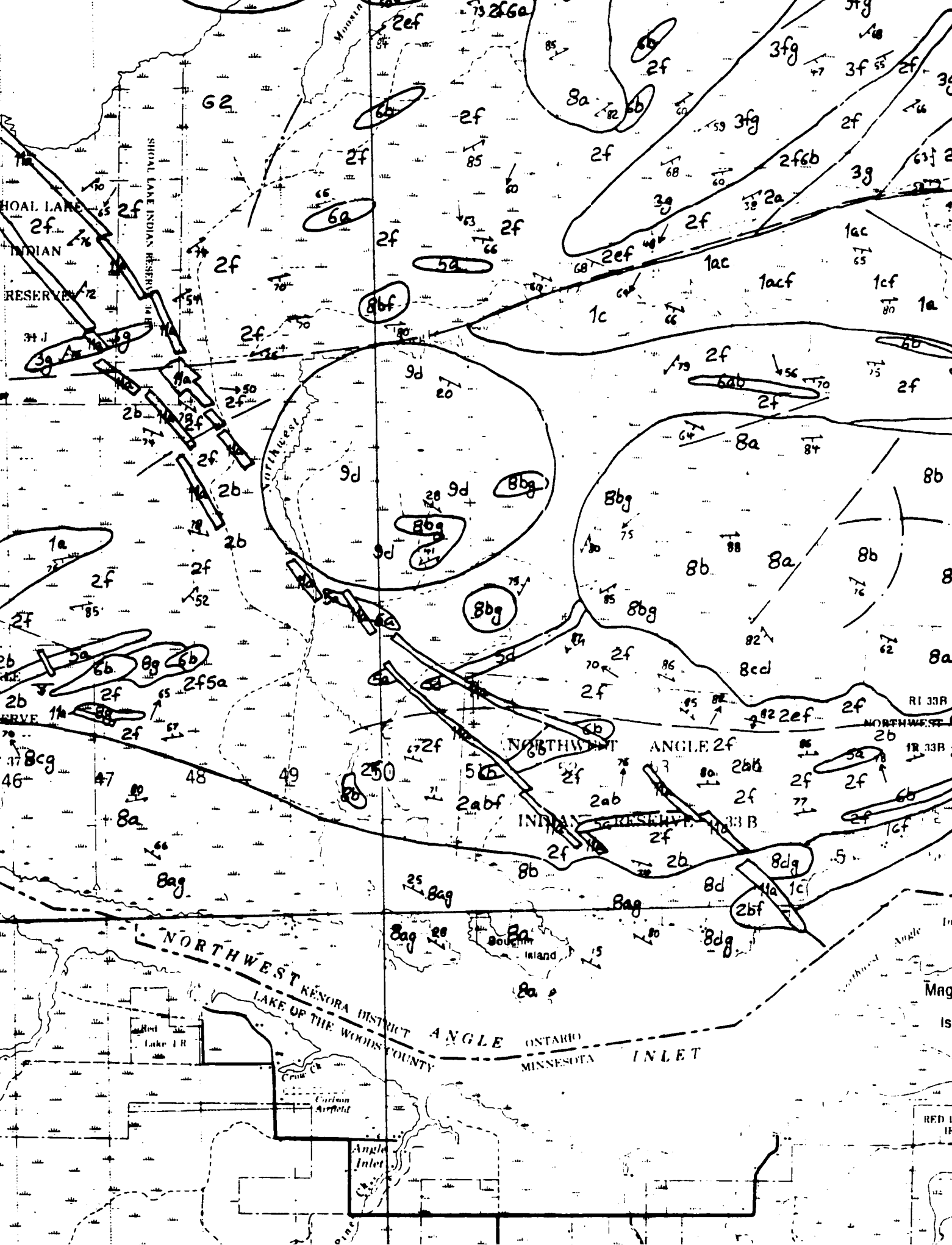


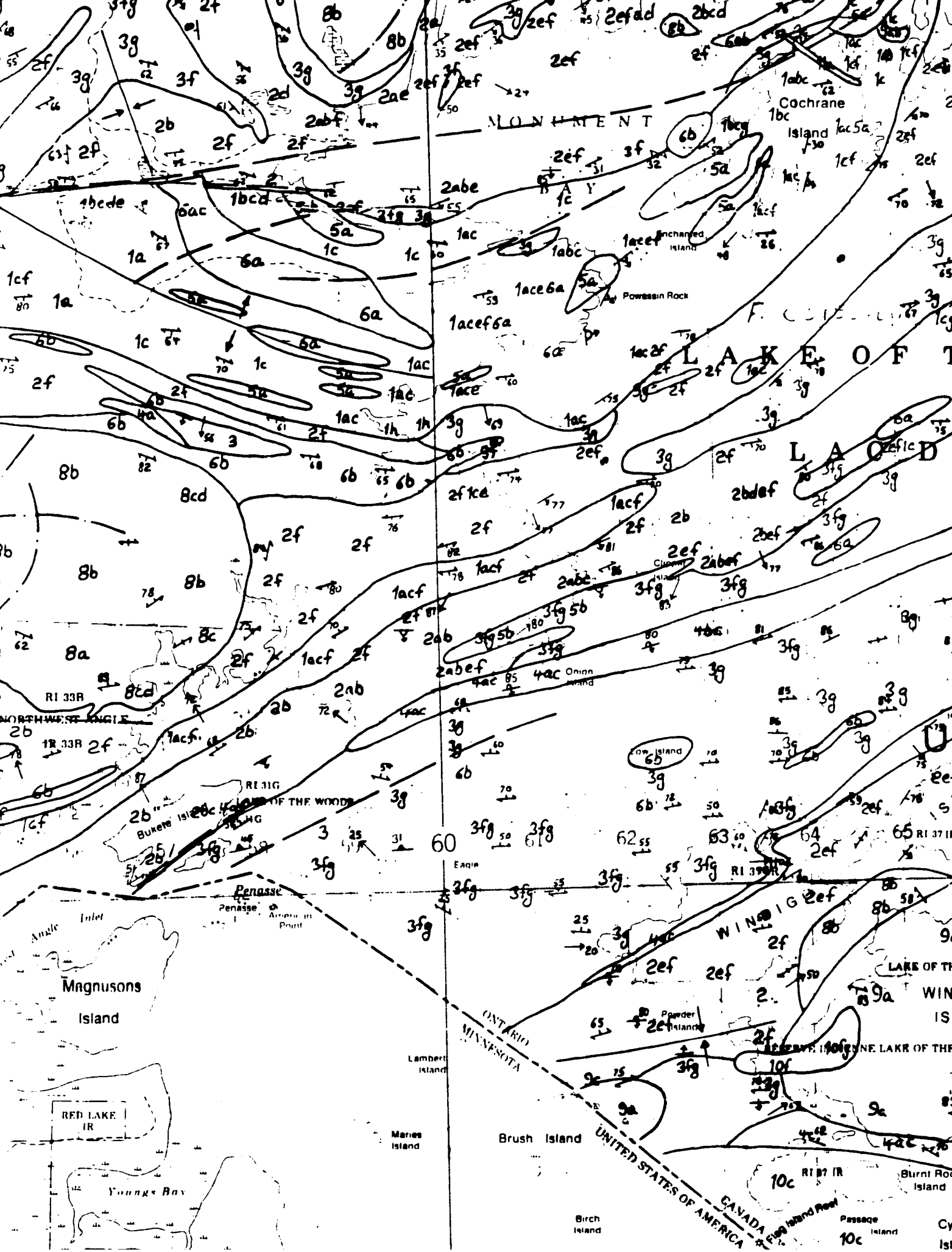


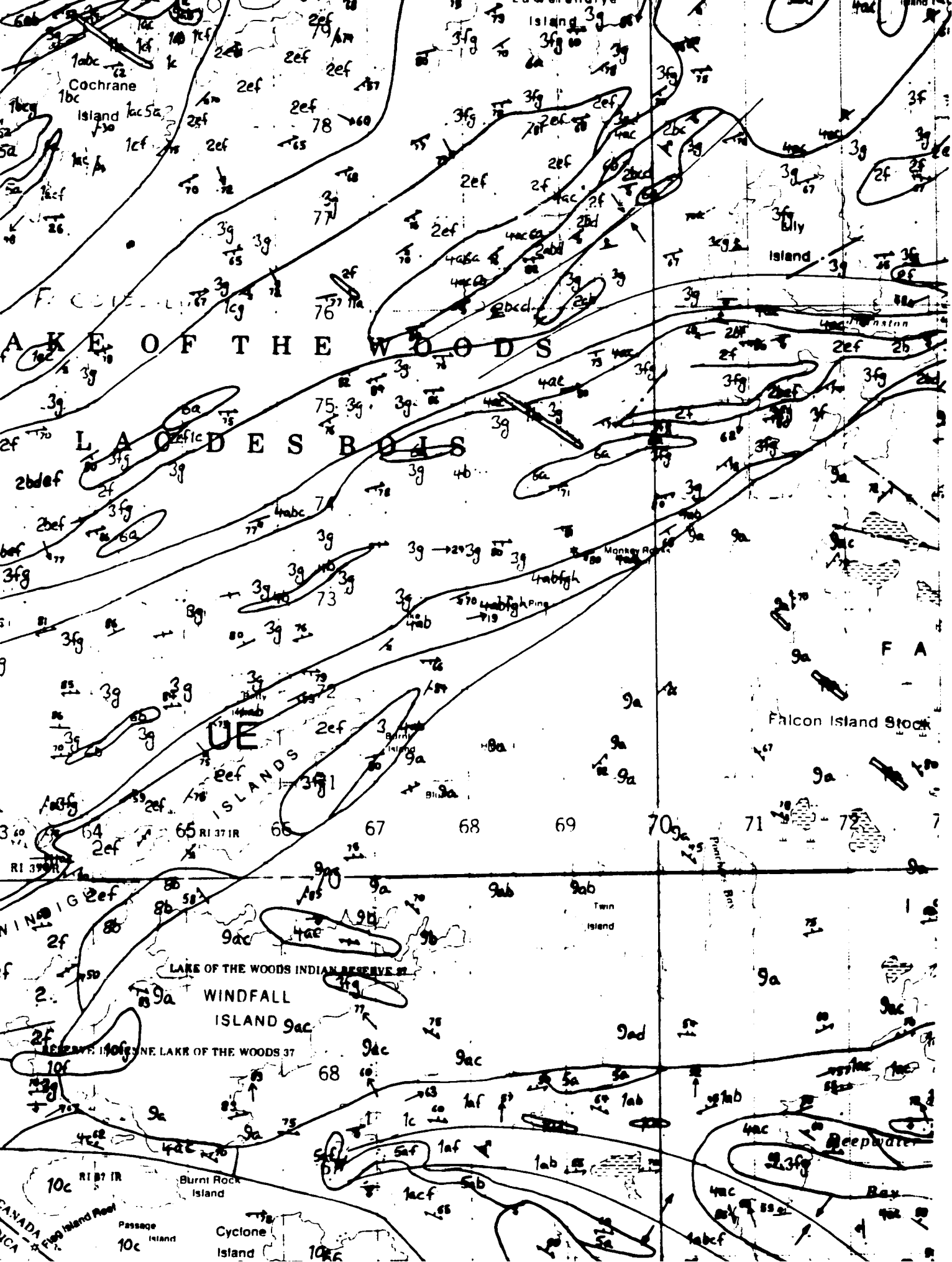


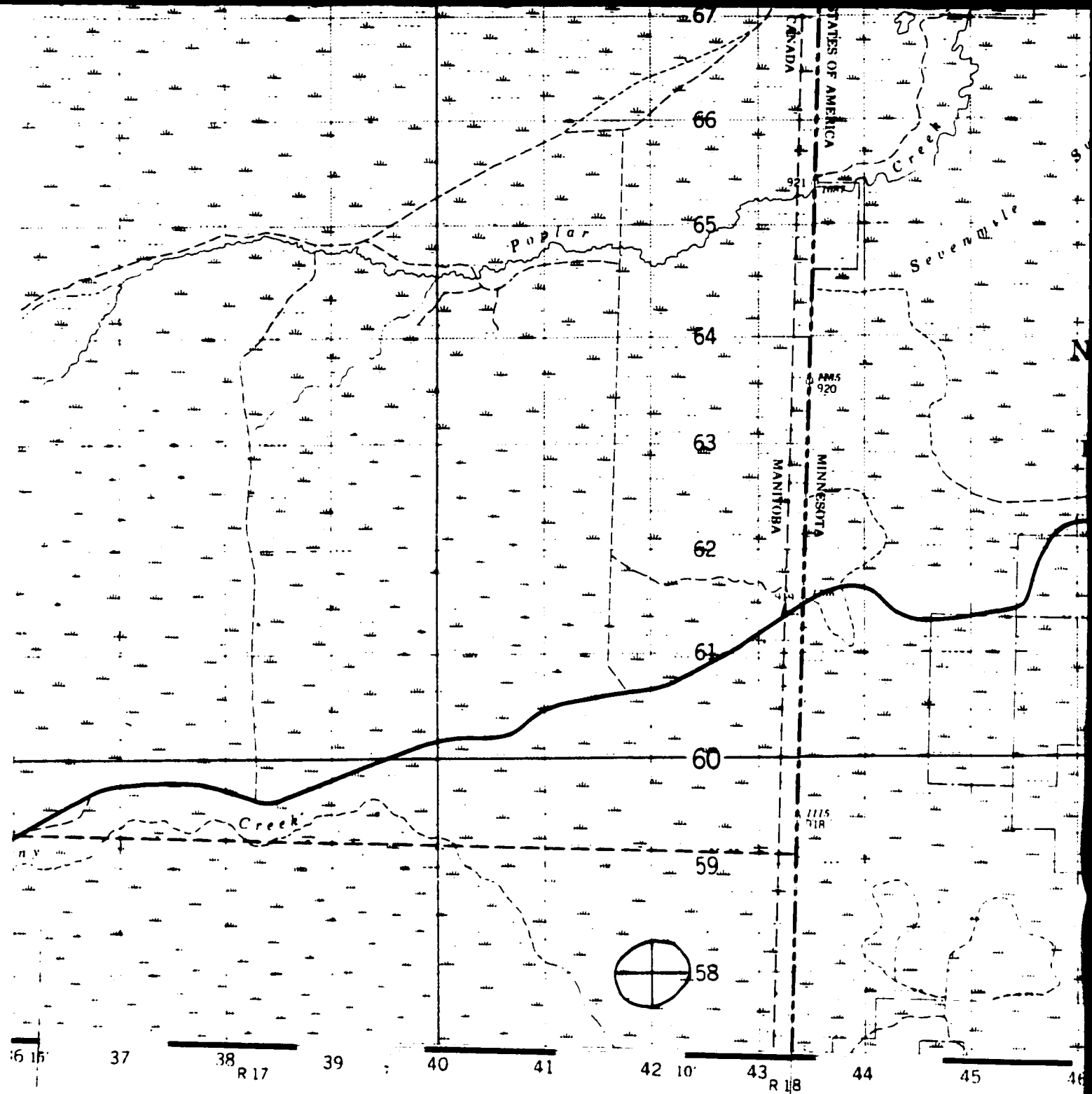


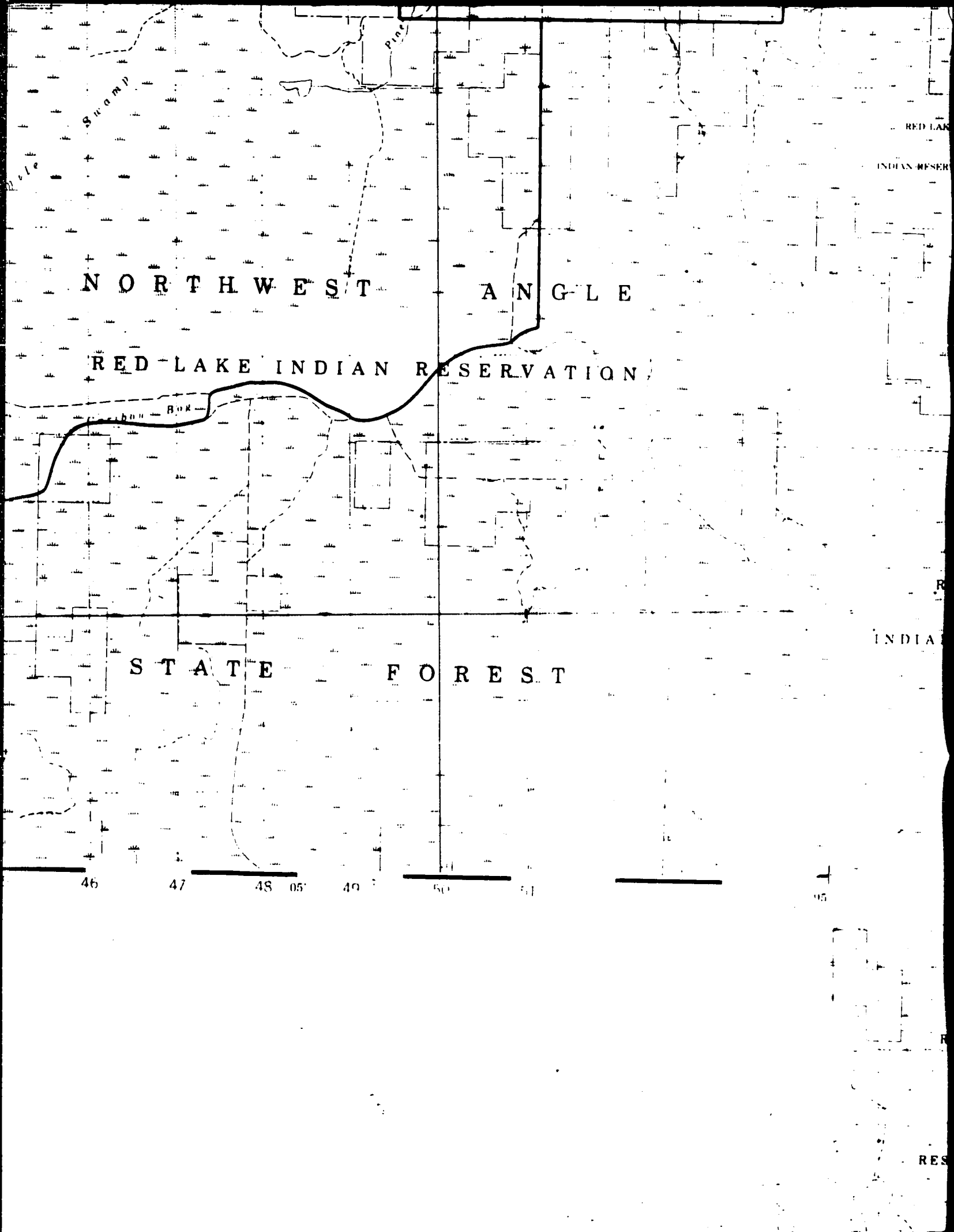


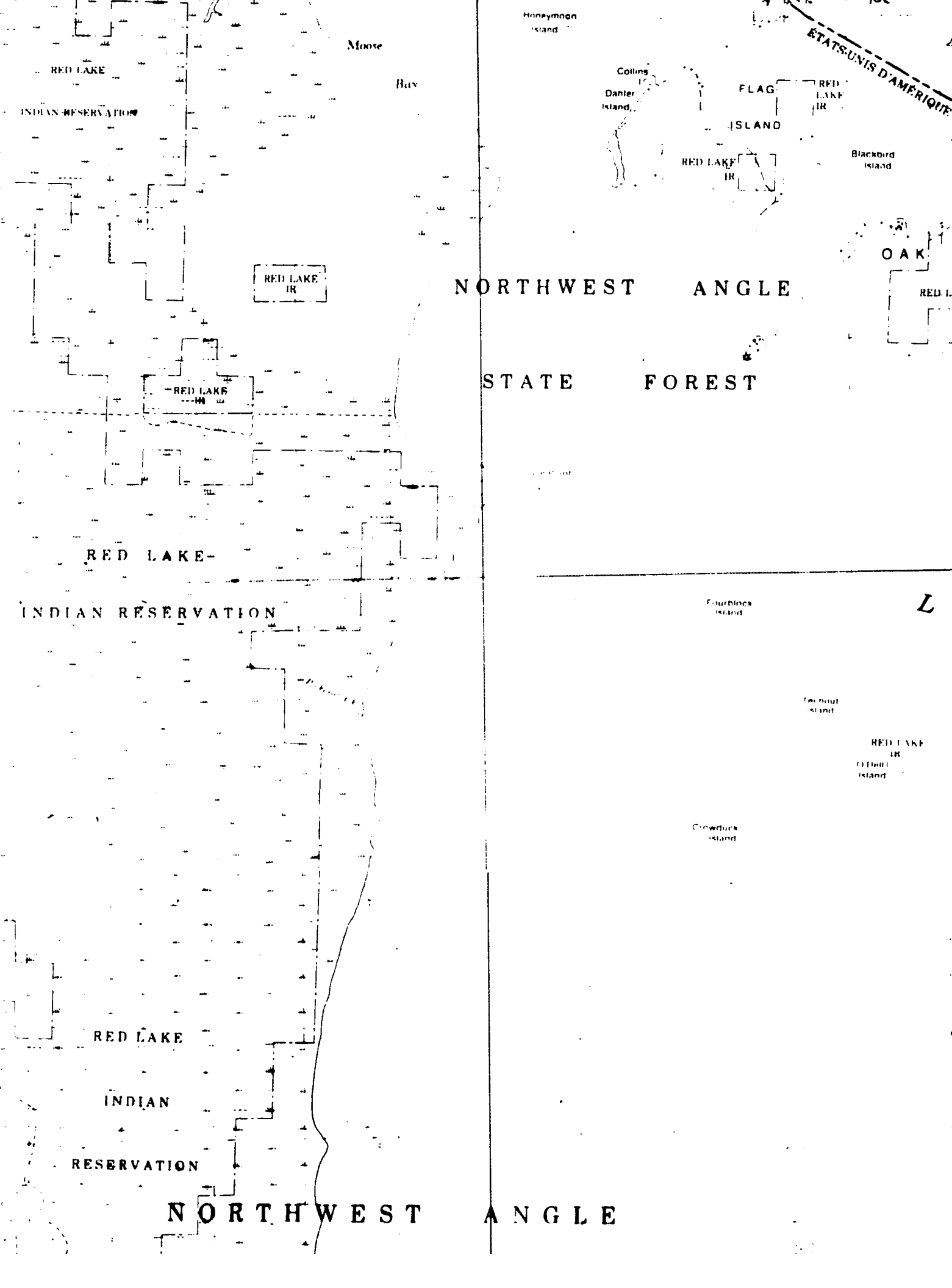












Moose

Bay

Honeymoon
Island

Collins
Island

FLAG
ISLAND

RED LAKE
IR

Blackbird
Island

OAK

RED L.

NORTHWEST ANGLE

STATE FOREST

RED LAKE-

INDIAN RESERVATION

Fourblock
Island

L

RED LAKE
Island

RED LAKE
IR
O'Dell
Island

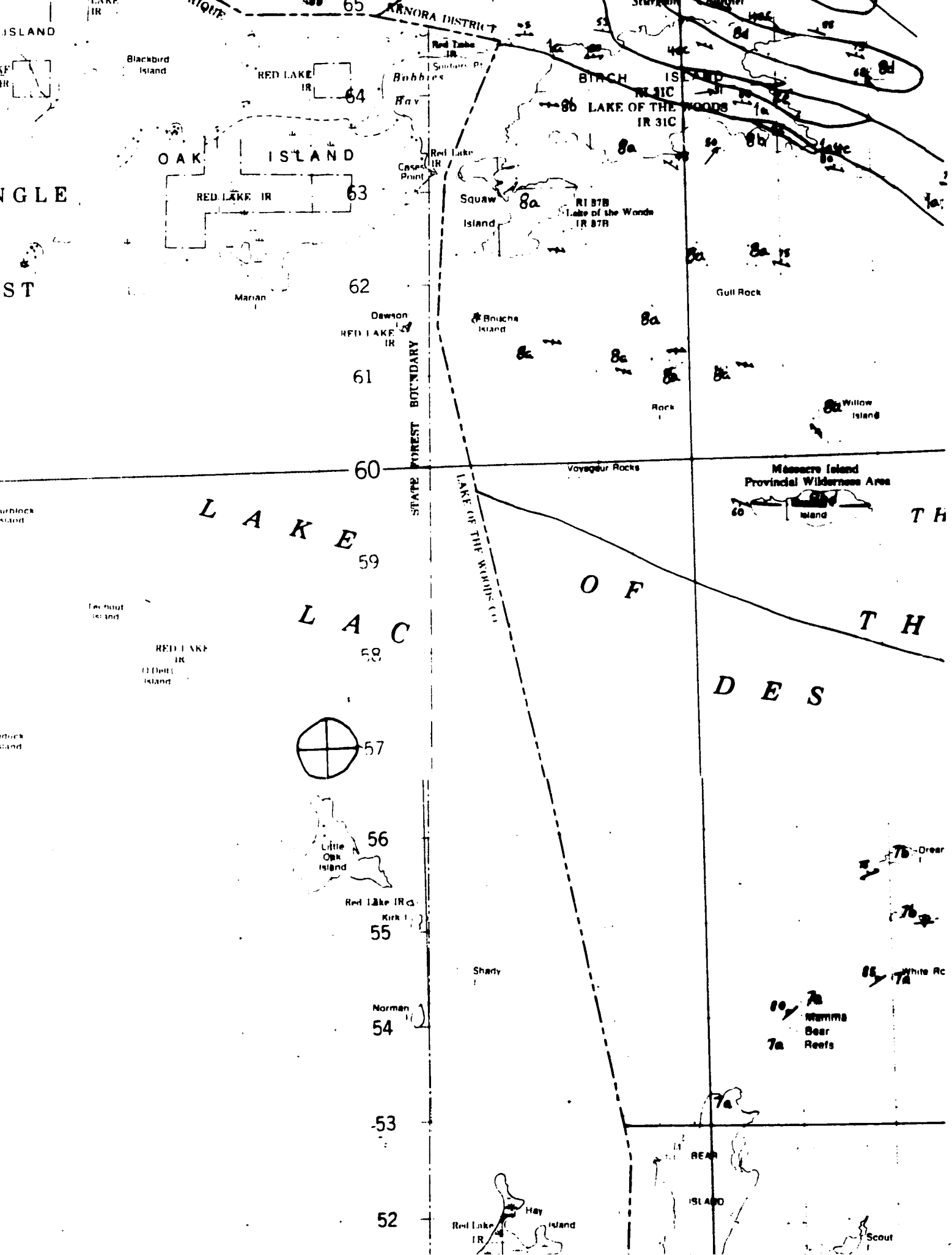
Crowduck
Island

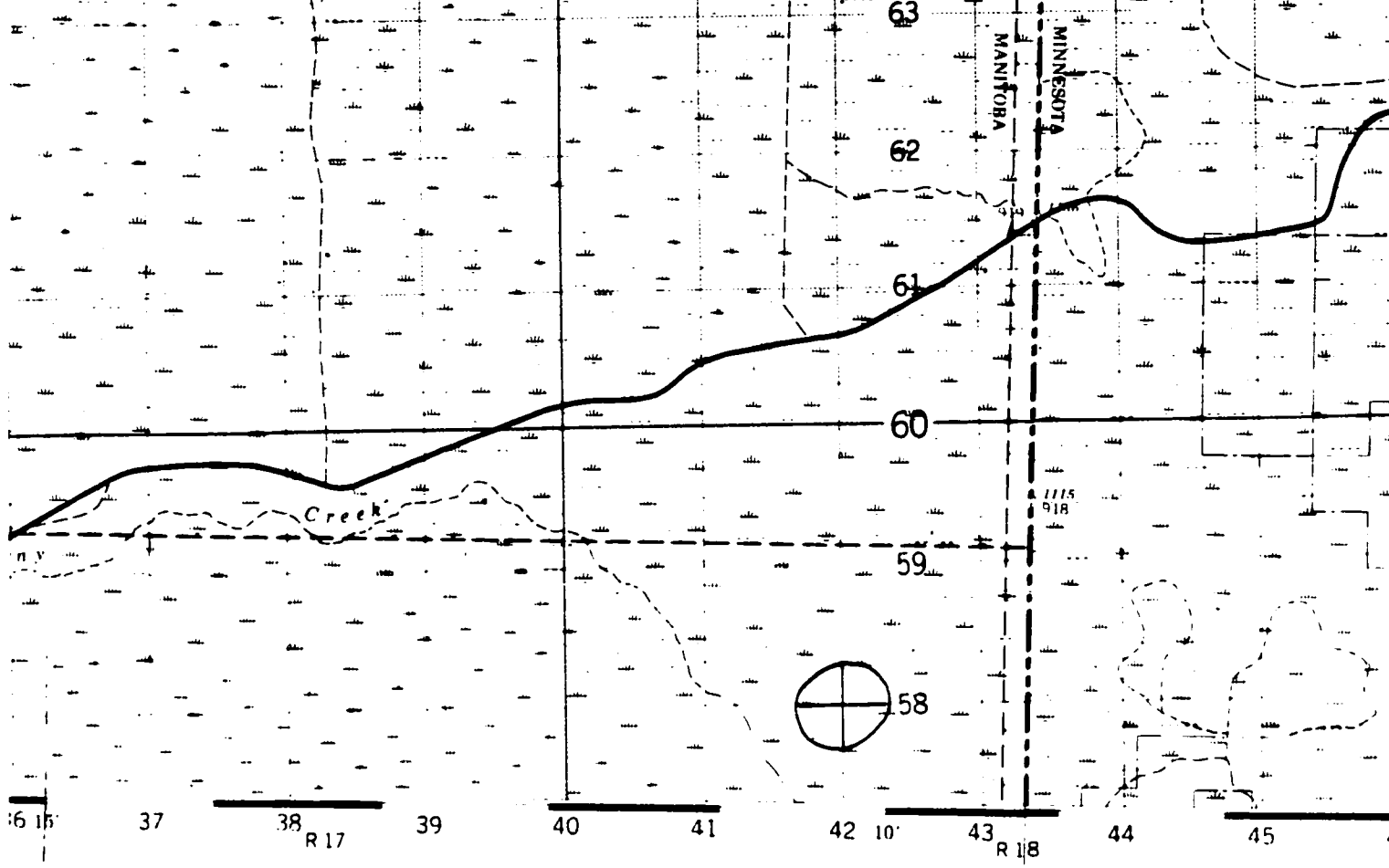
RED LAKE

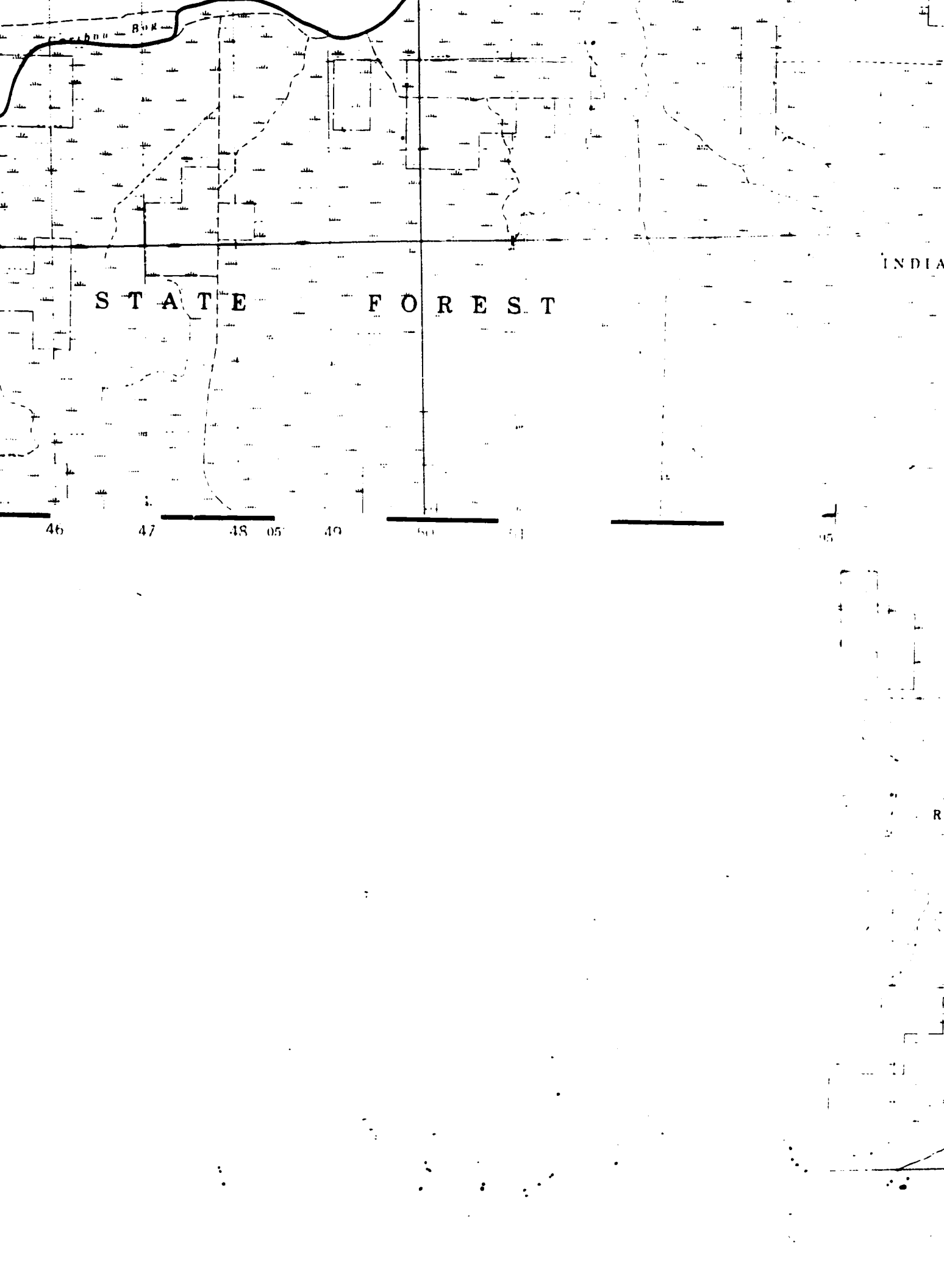
INDIAN

RESERVATION

NORTHWEST ANGLE







STATE FOREST

INDIA

46

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R

STATE

FOREST

RED LAKE

RED LAKE

INDIAN RESERVATION

Fourblock
Island

Two hour
Island

RED I
IN
O. Duff
Island

Crowdick
Island

RED LAKE

INDIAN

RESERVATION

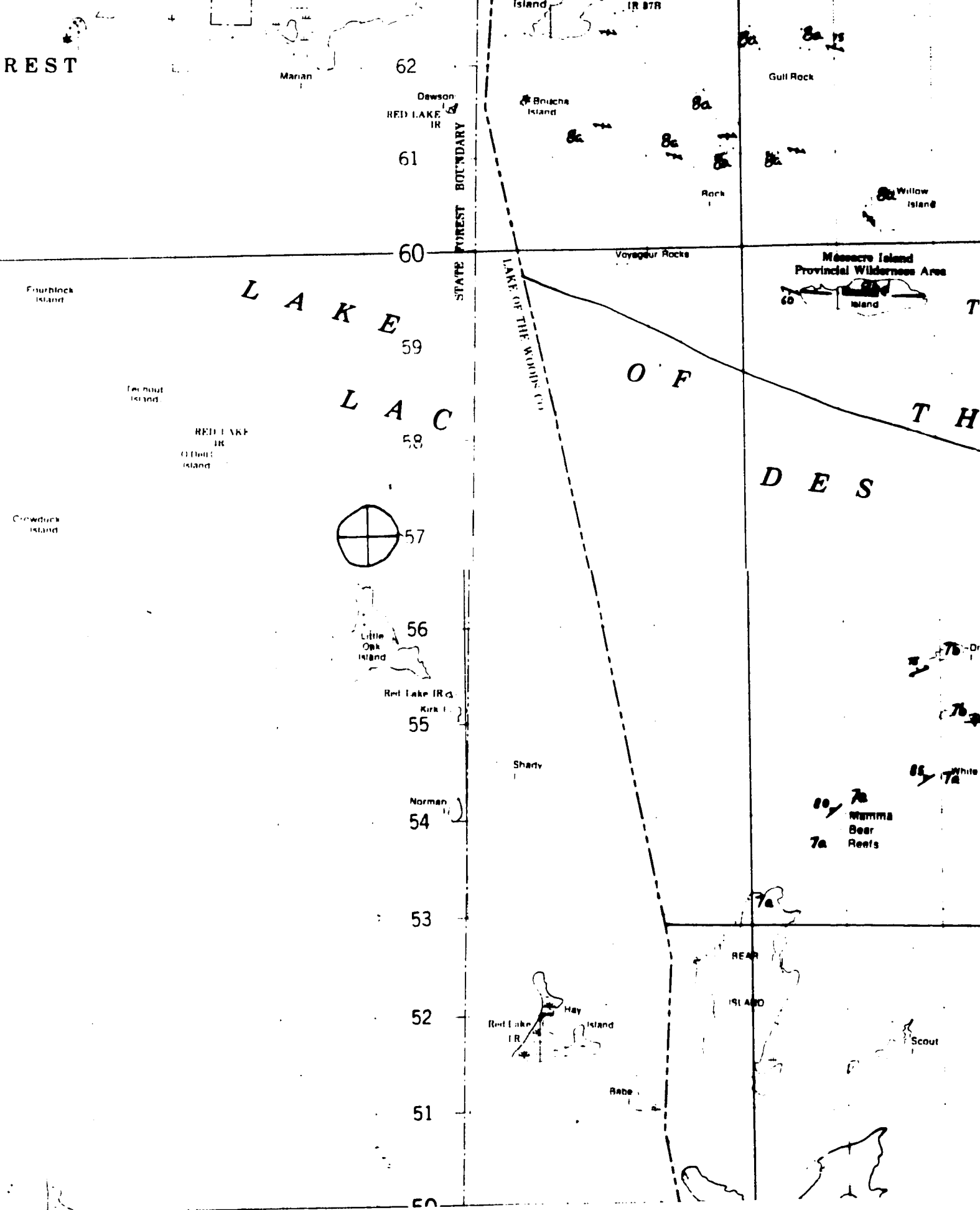
NORTHWEST ANGLE

De Ward
Pt

STATE FOREST

Rusty Pt

REST

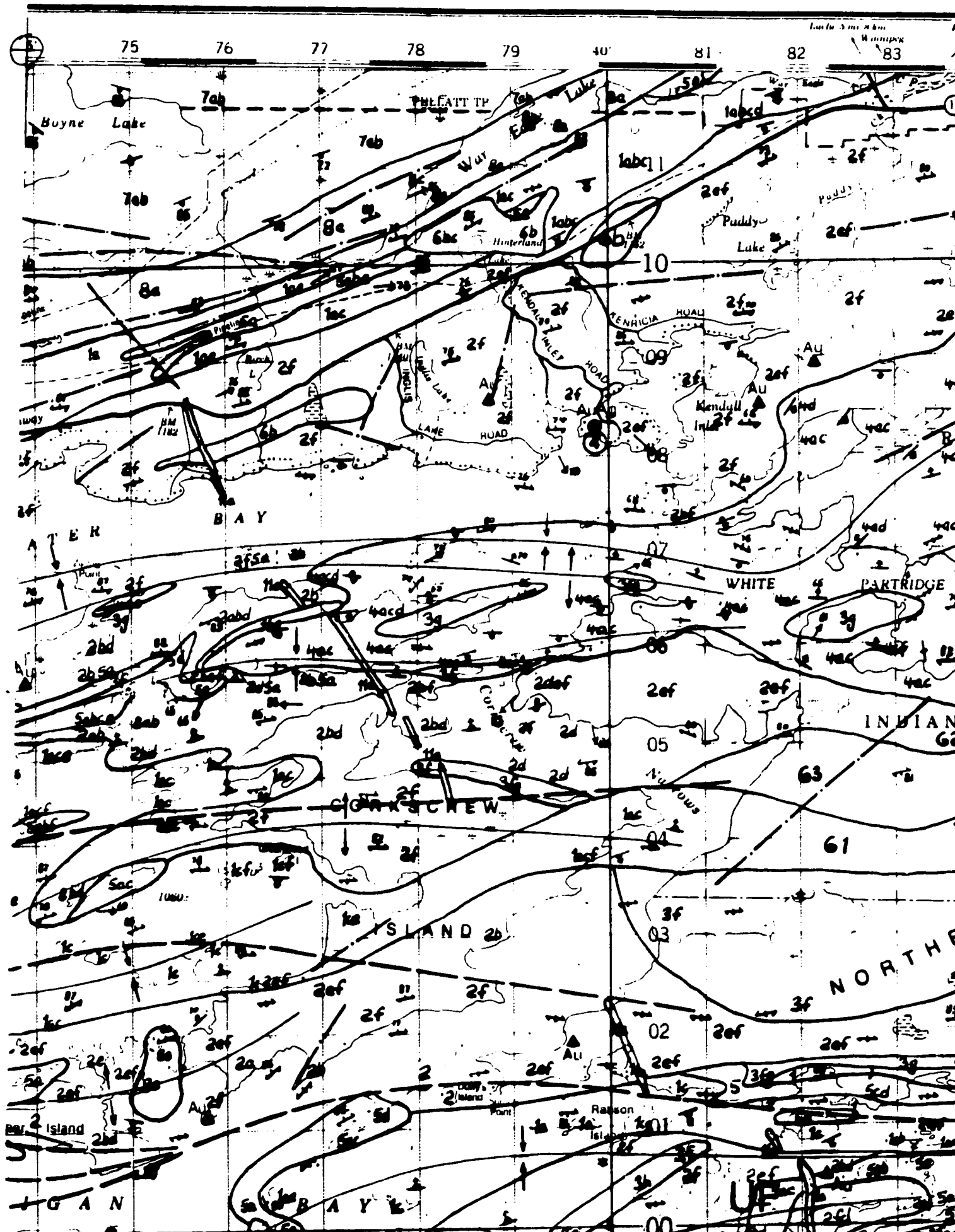


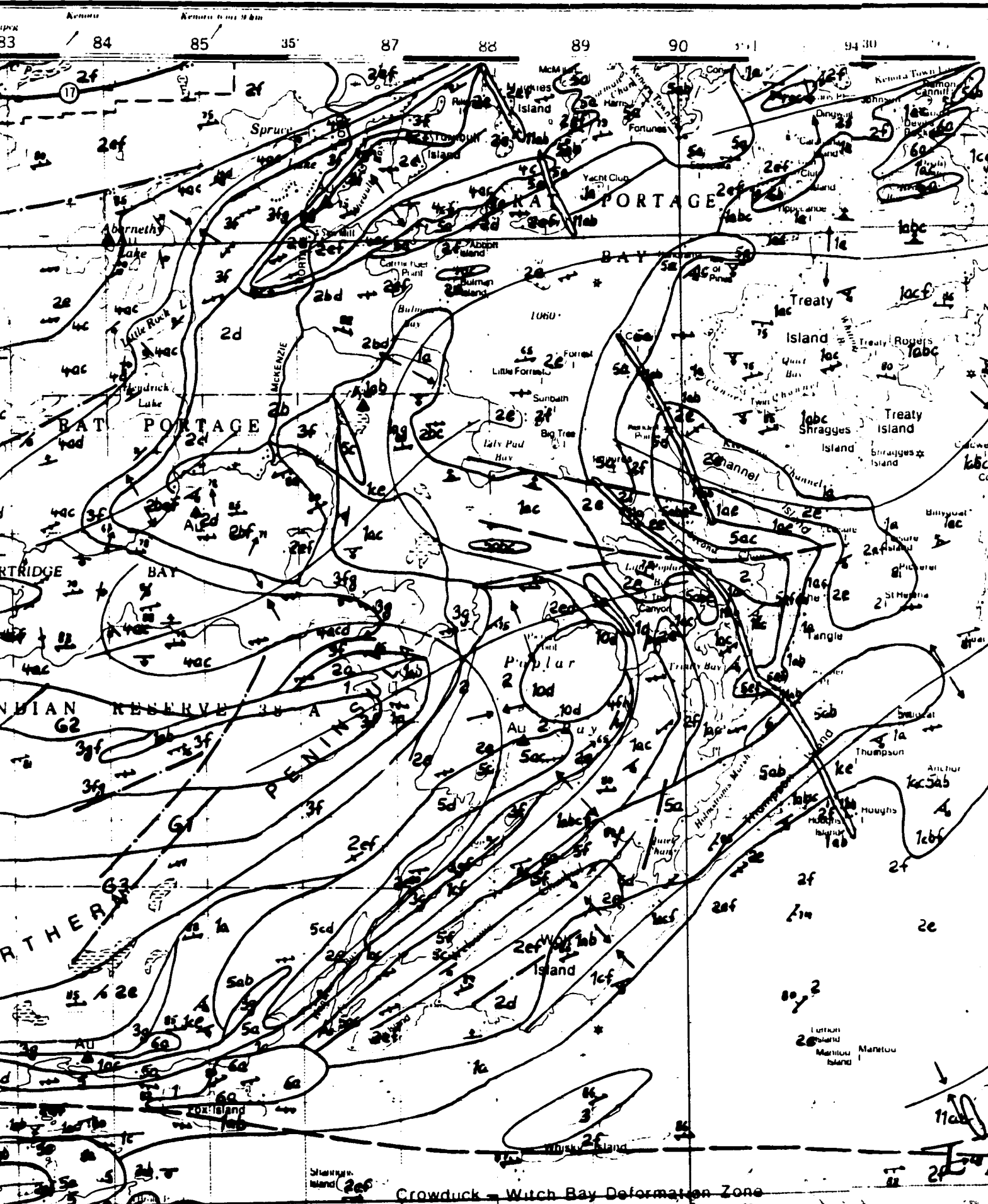
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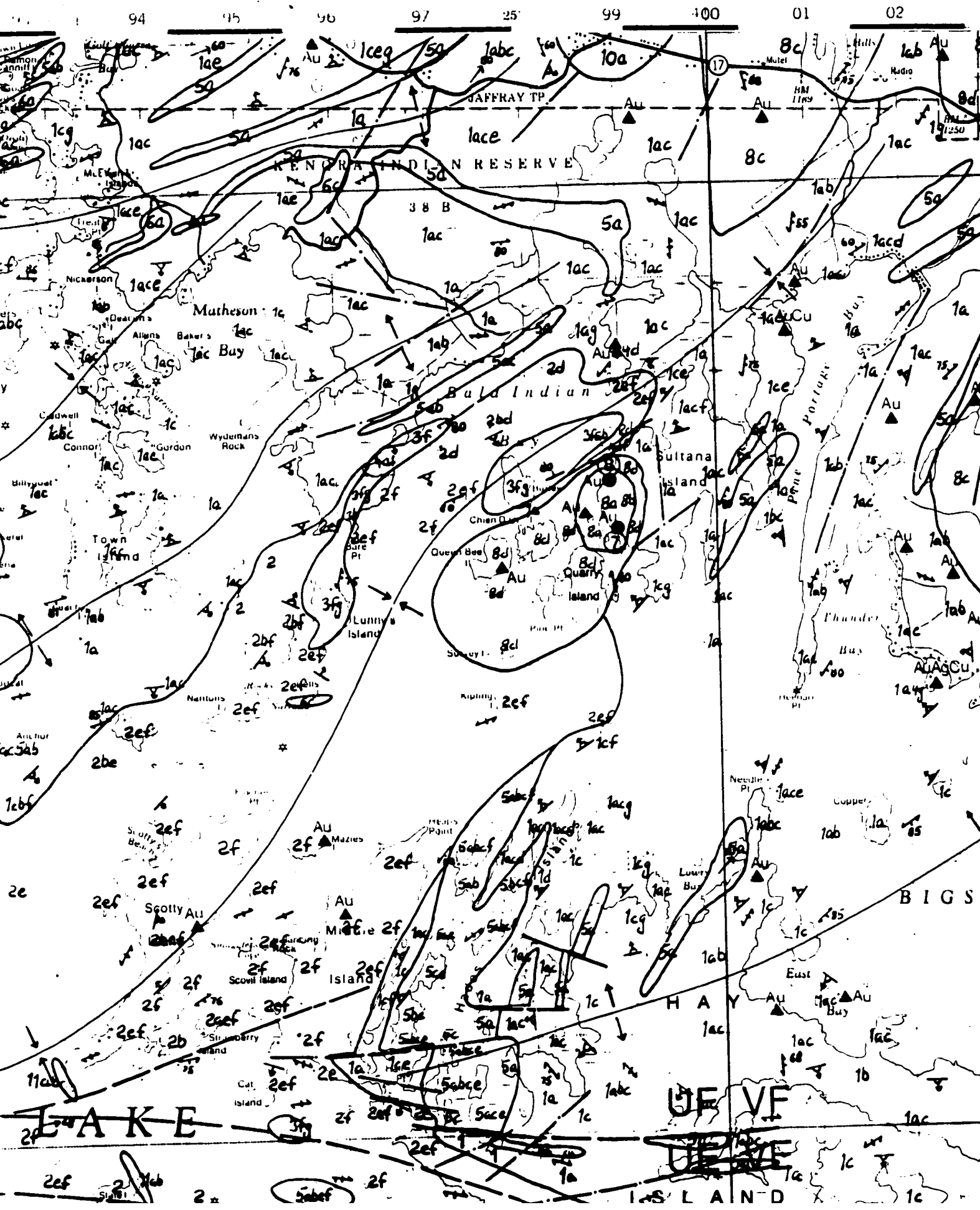
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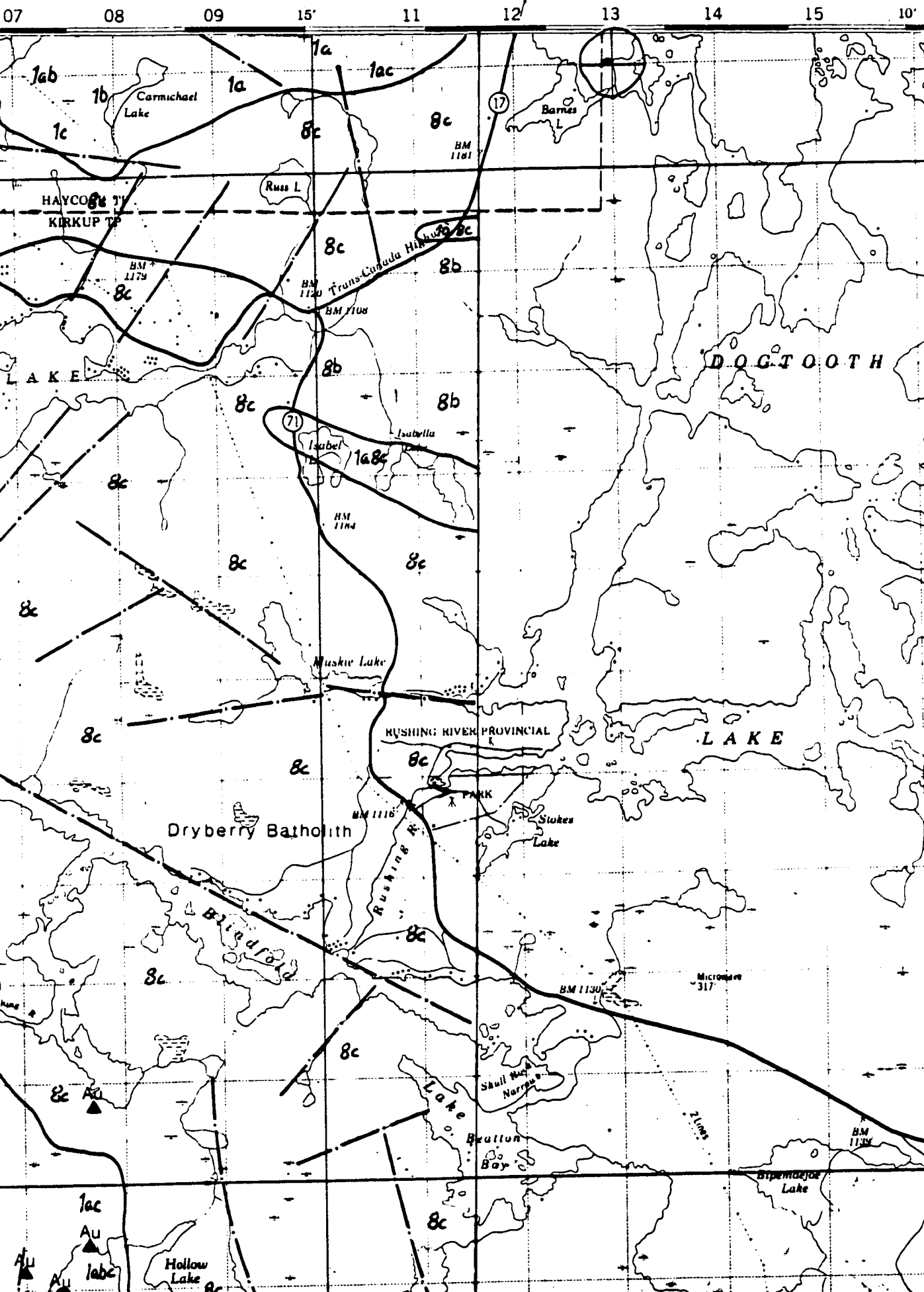
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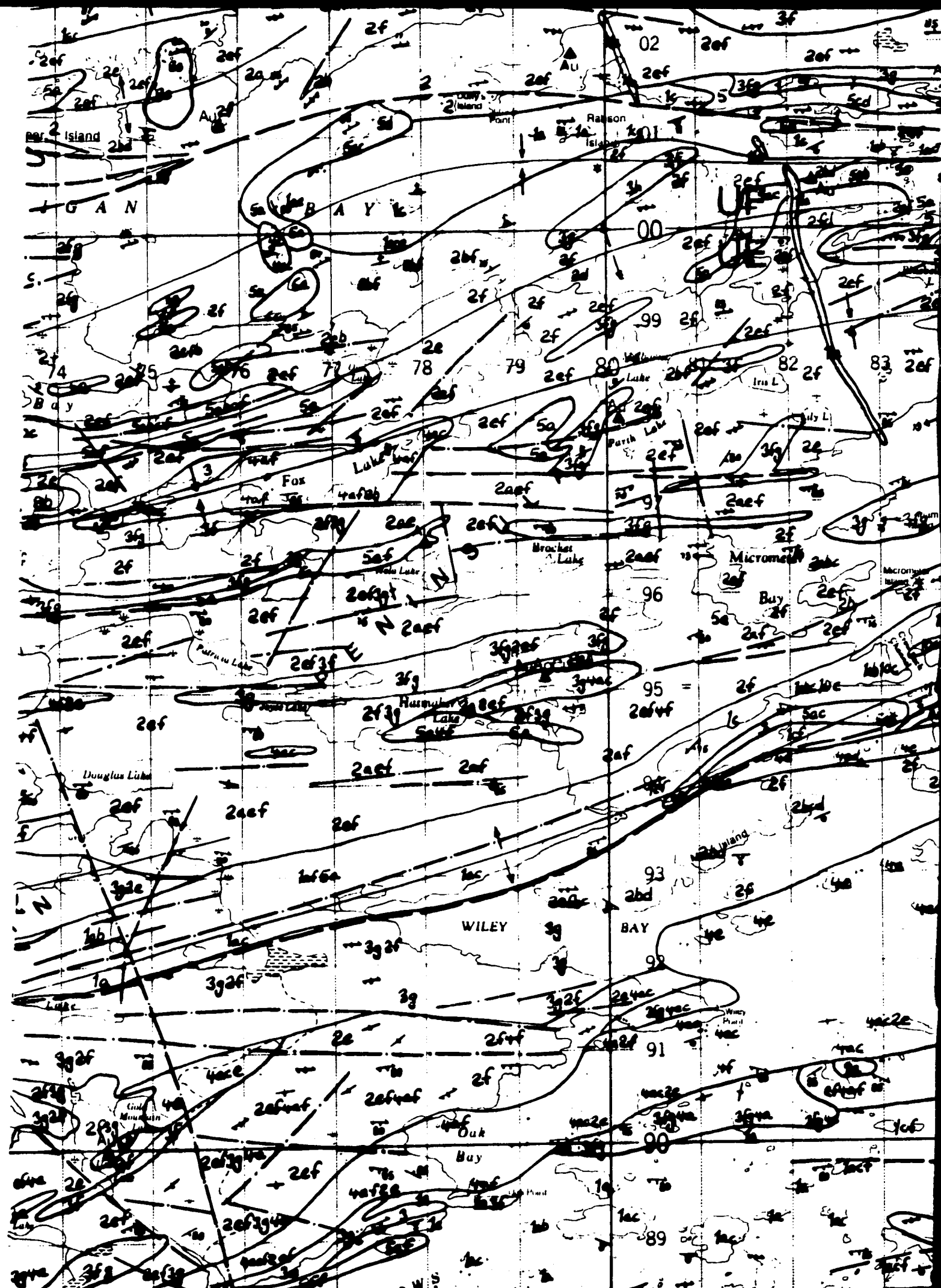
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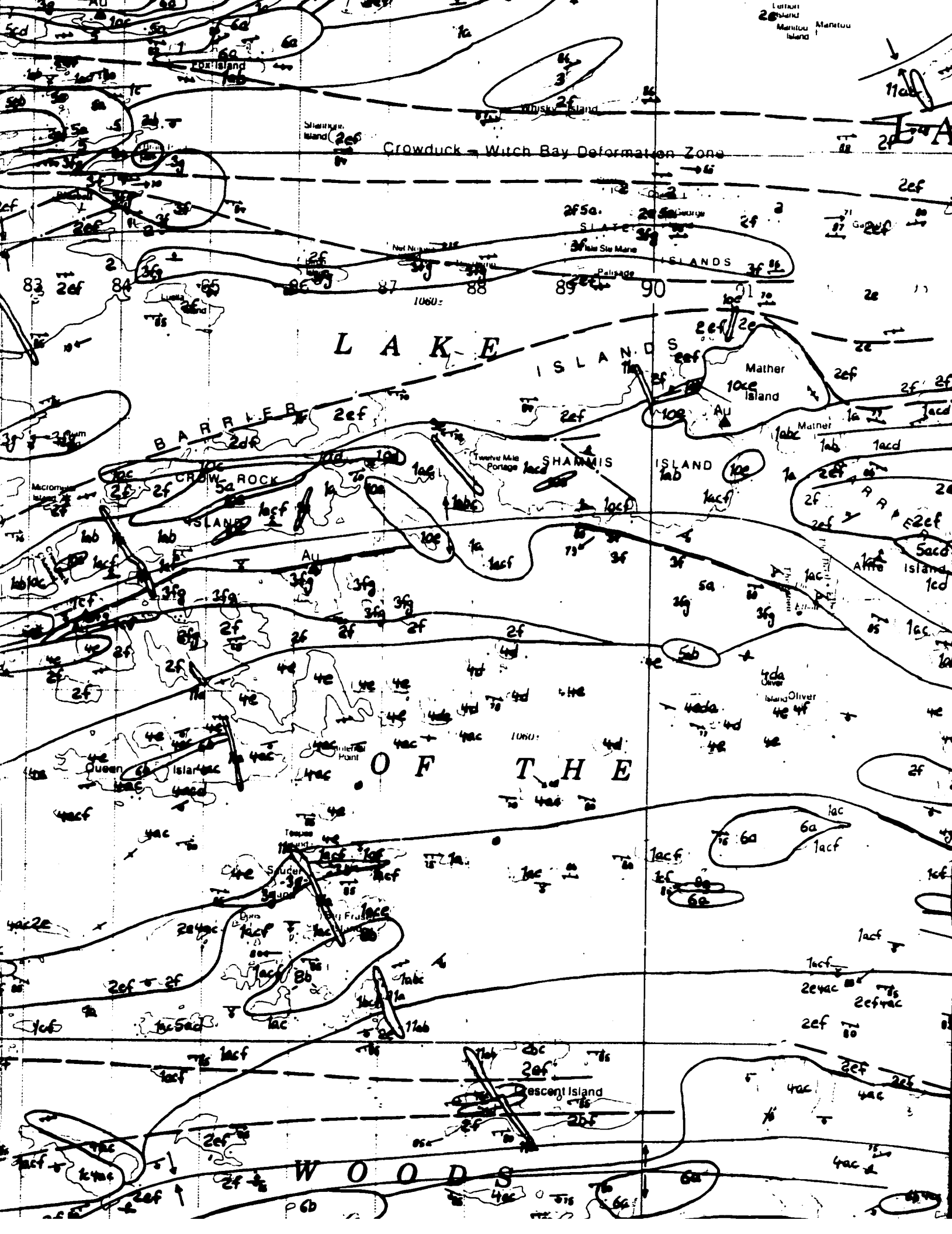


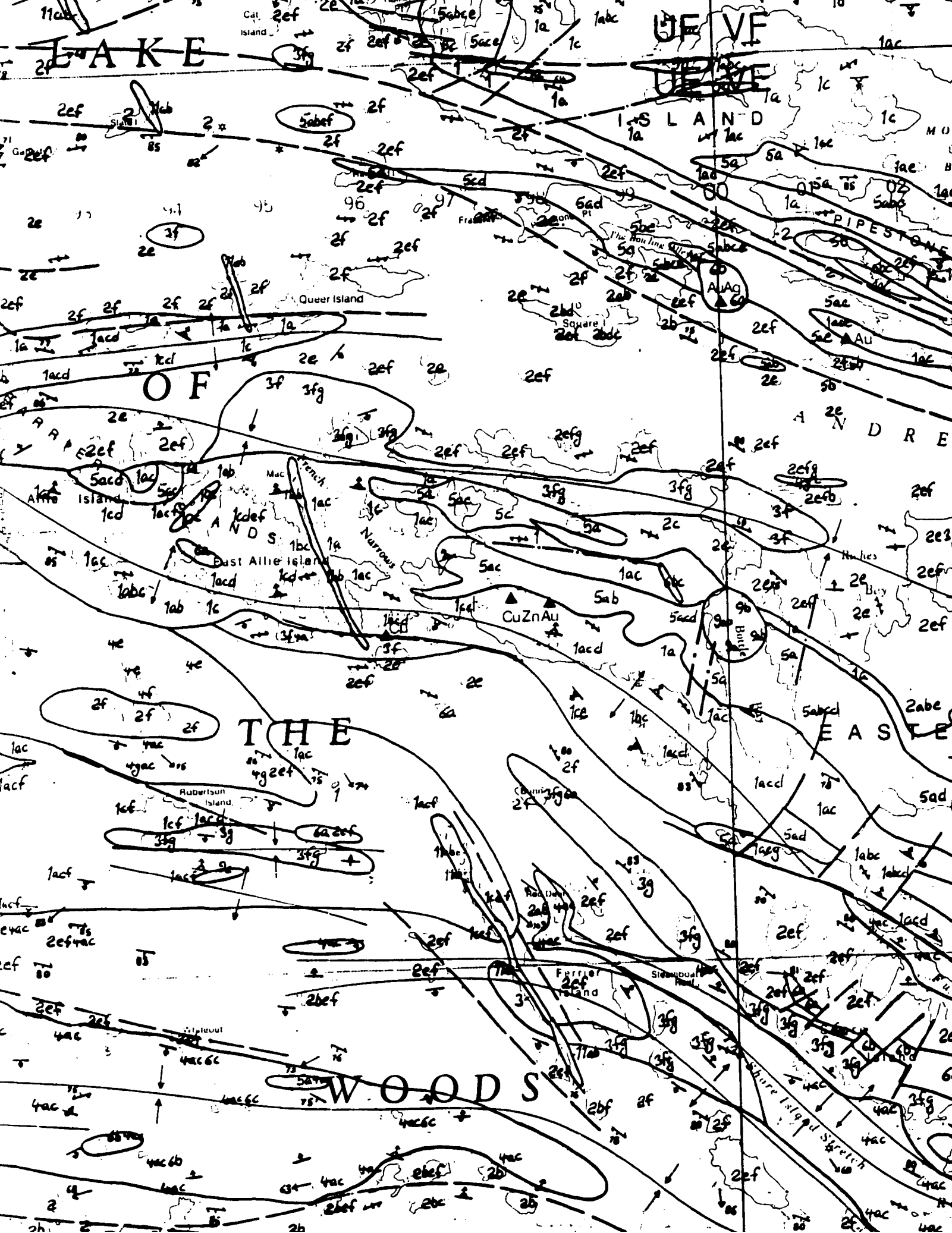


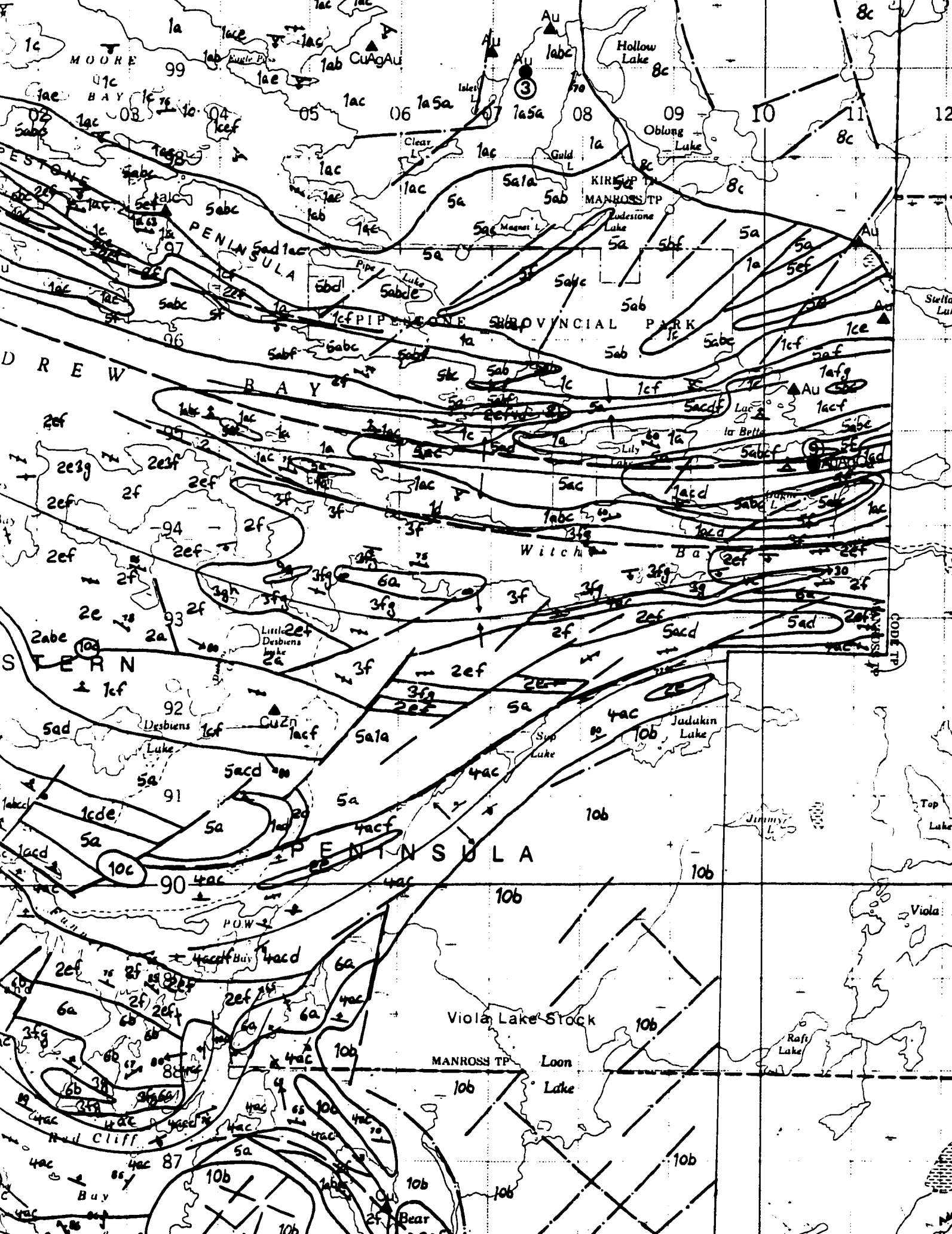


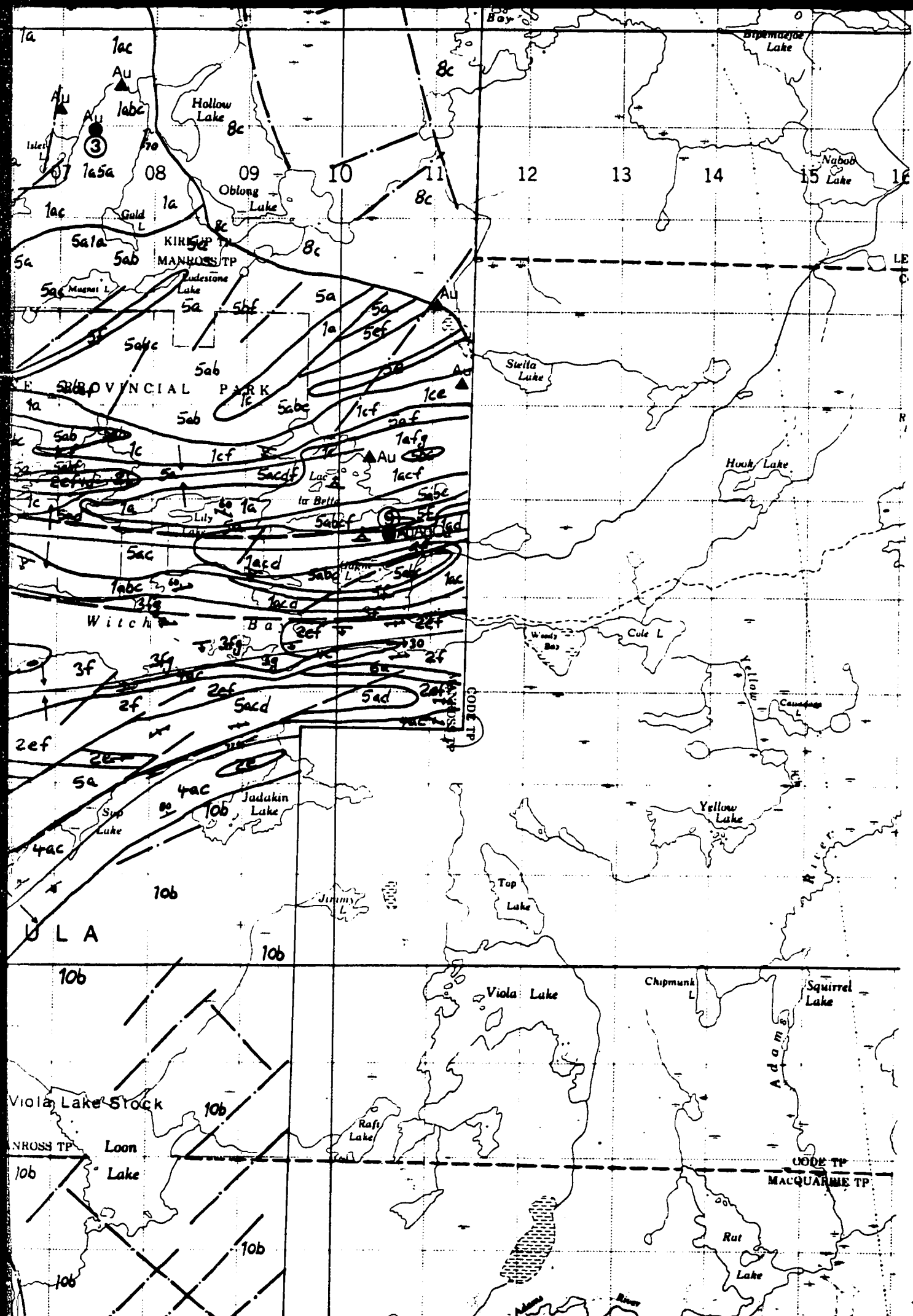


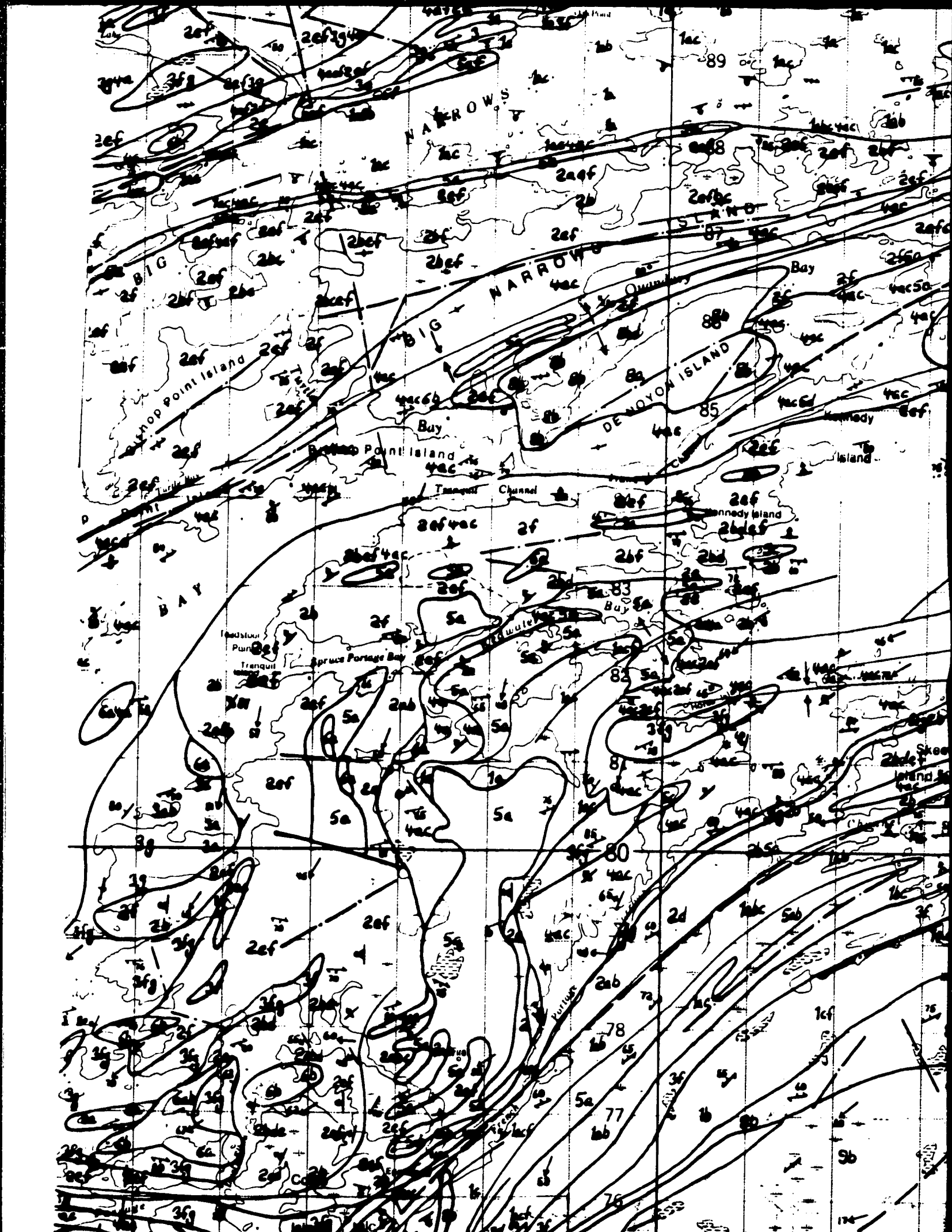


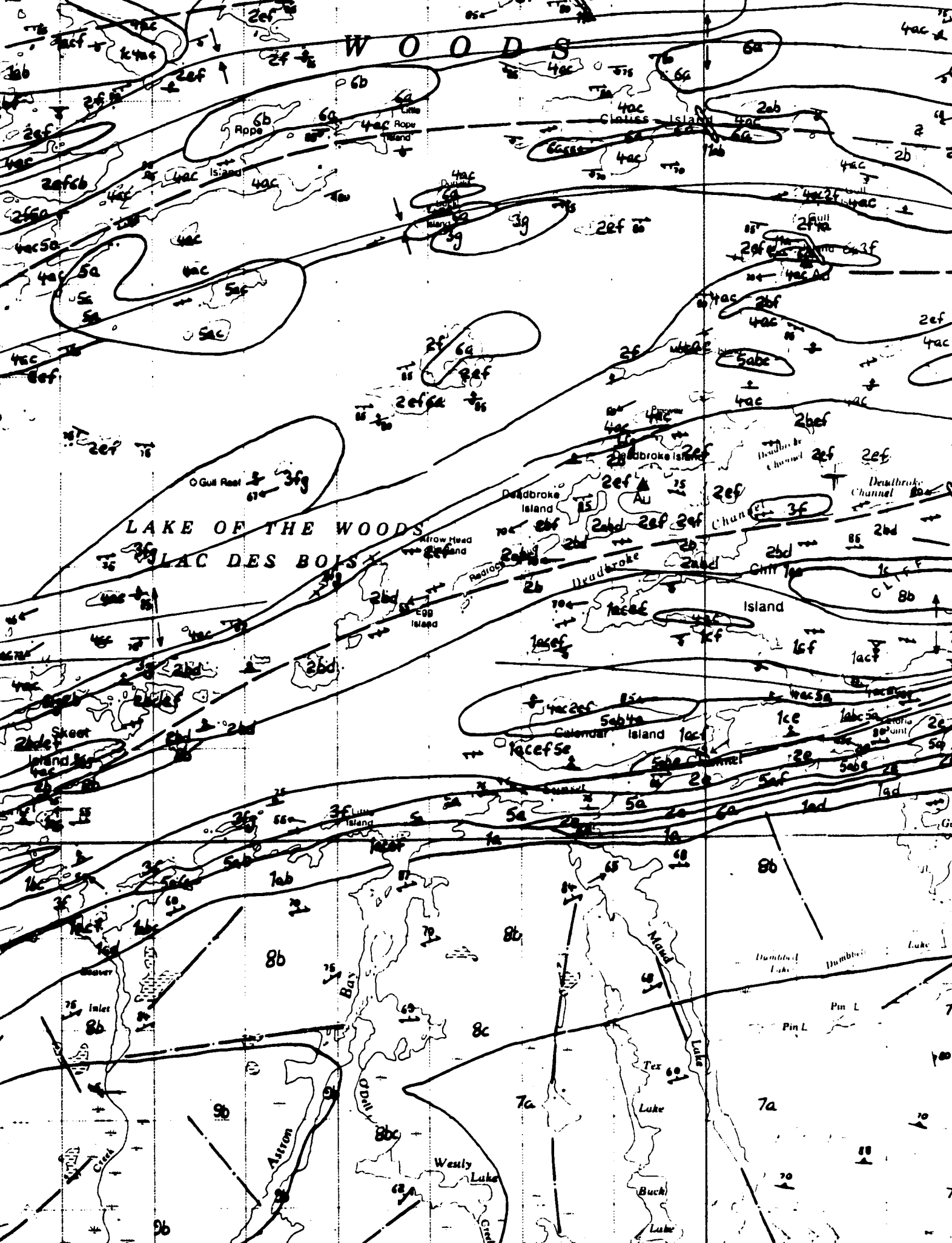




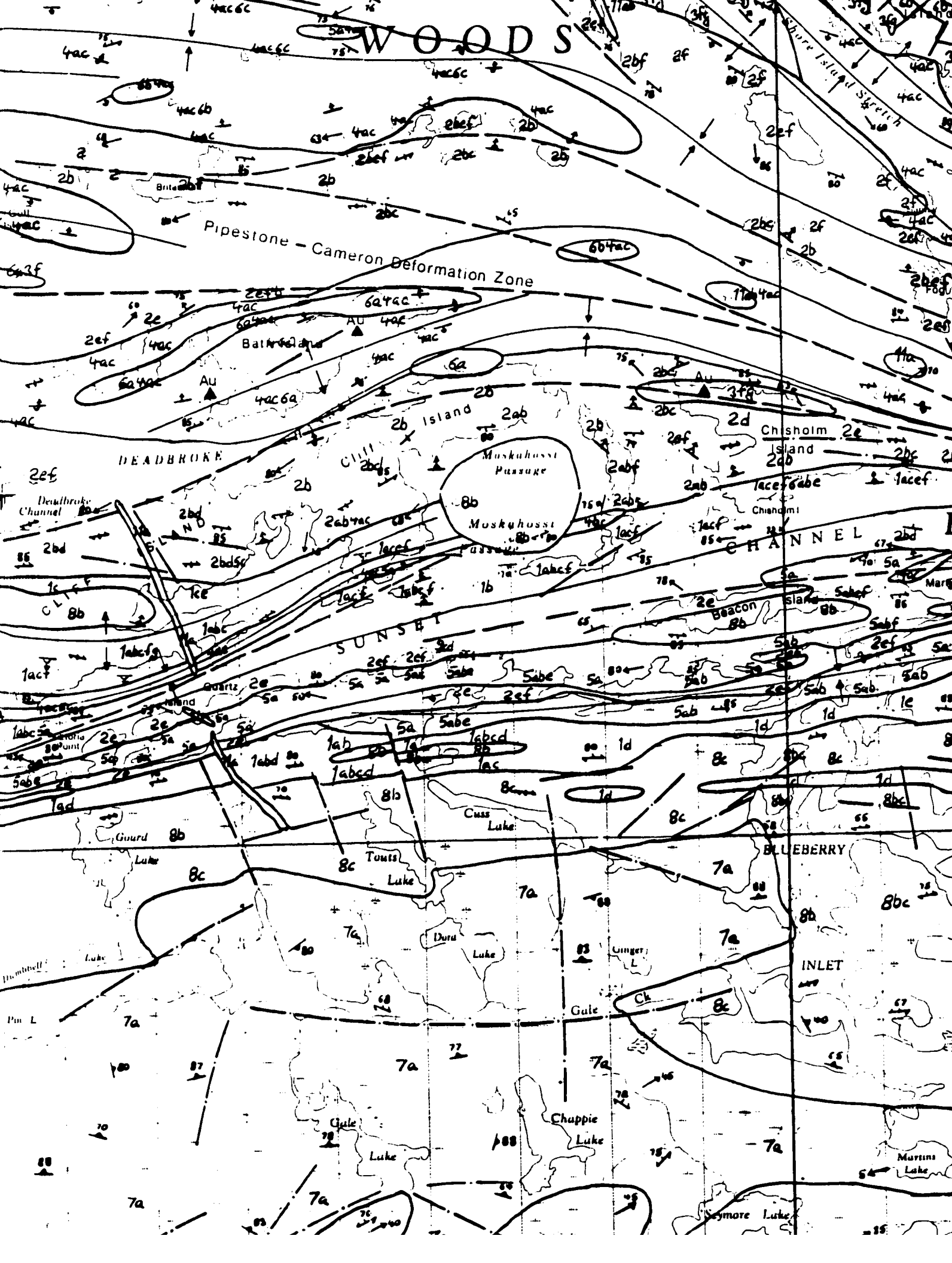


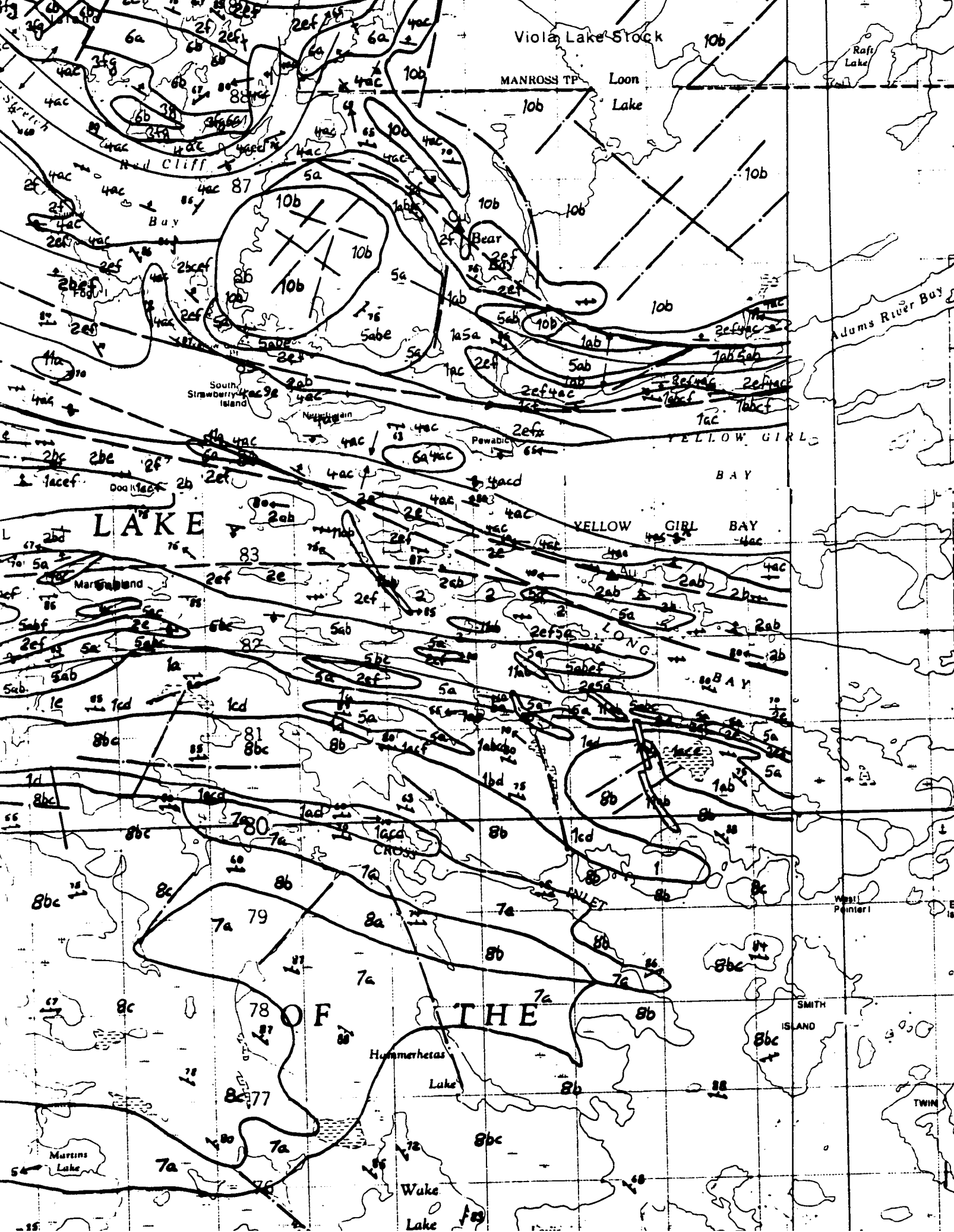


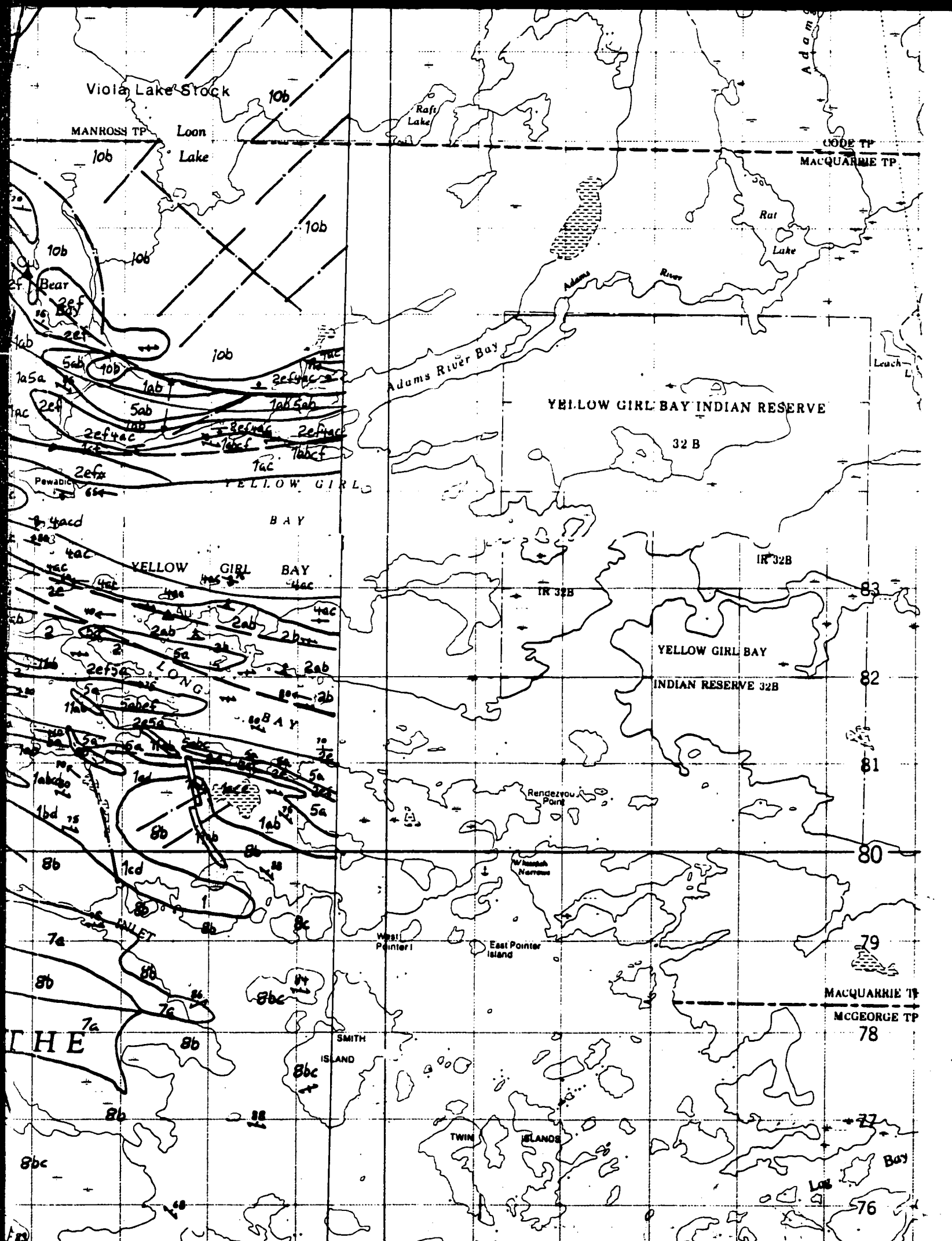


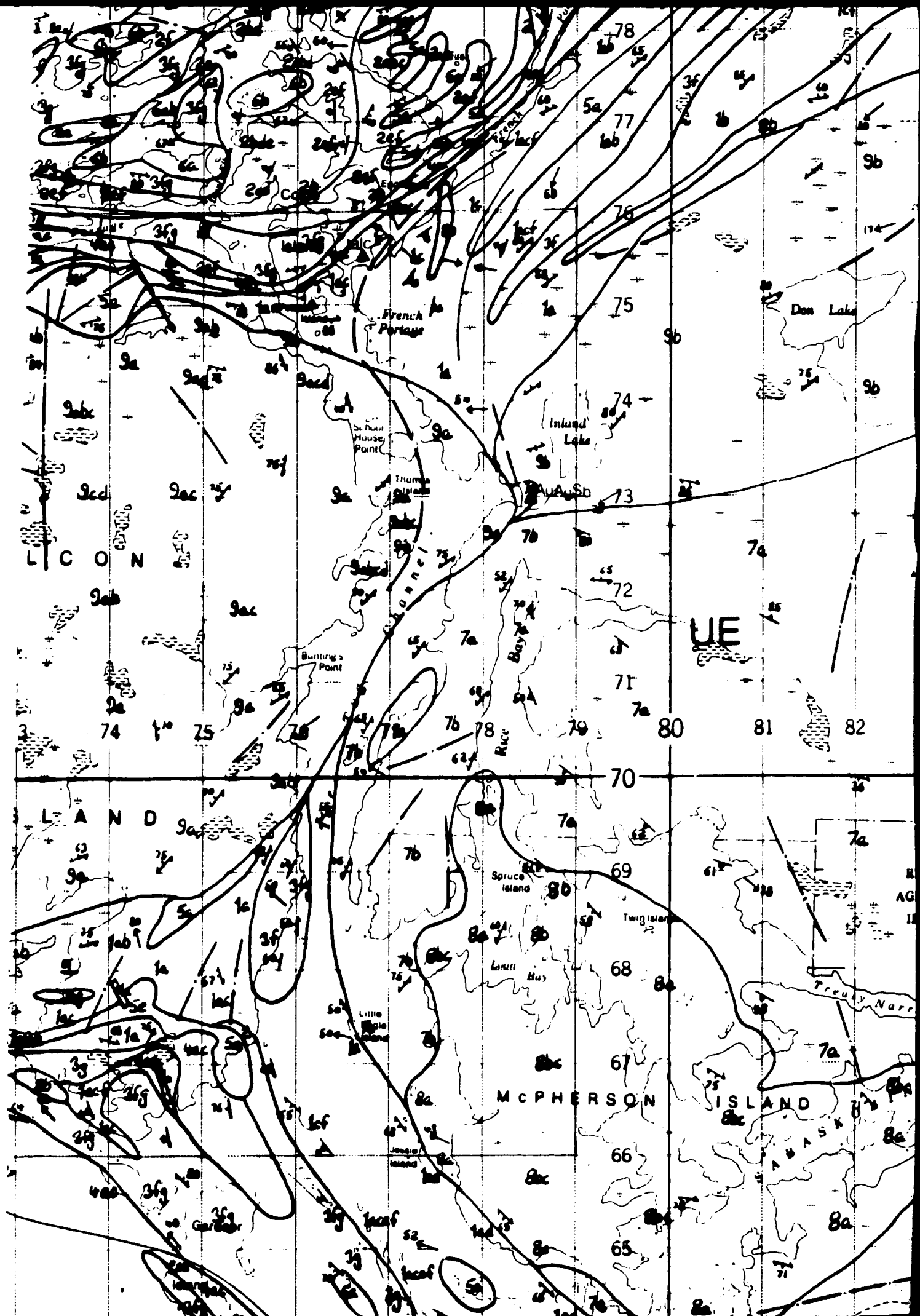


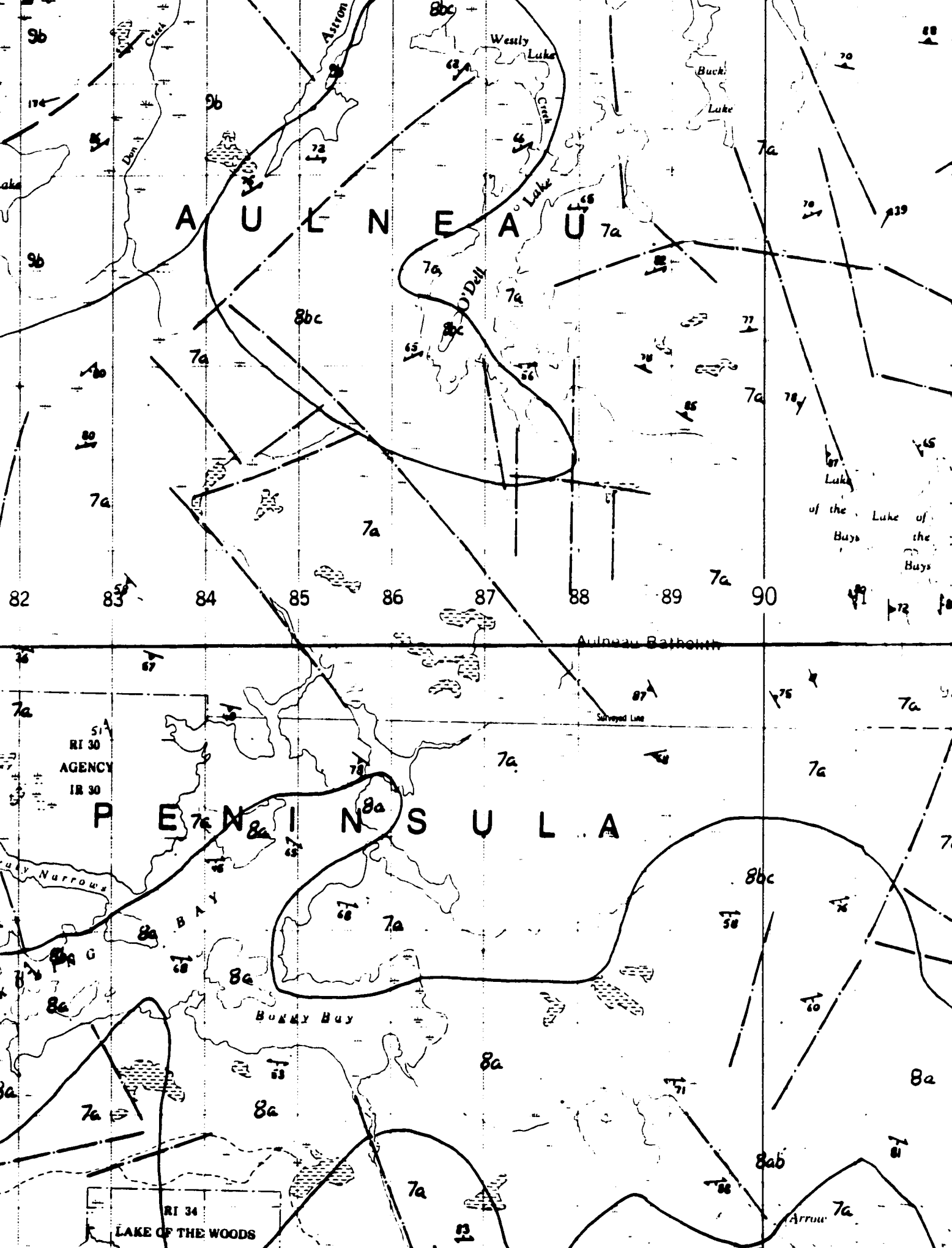
WOODS

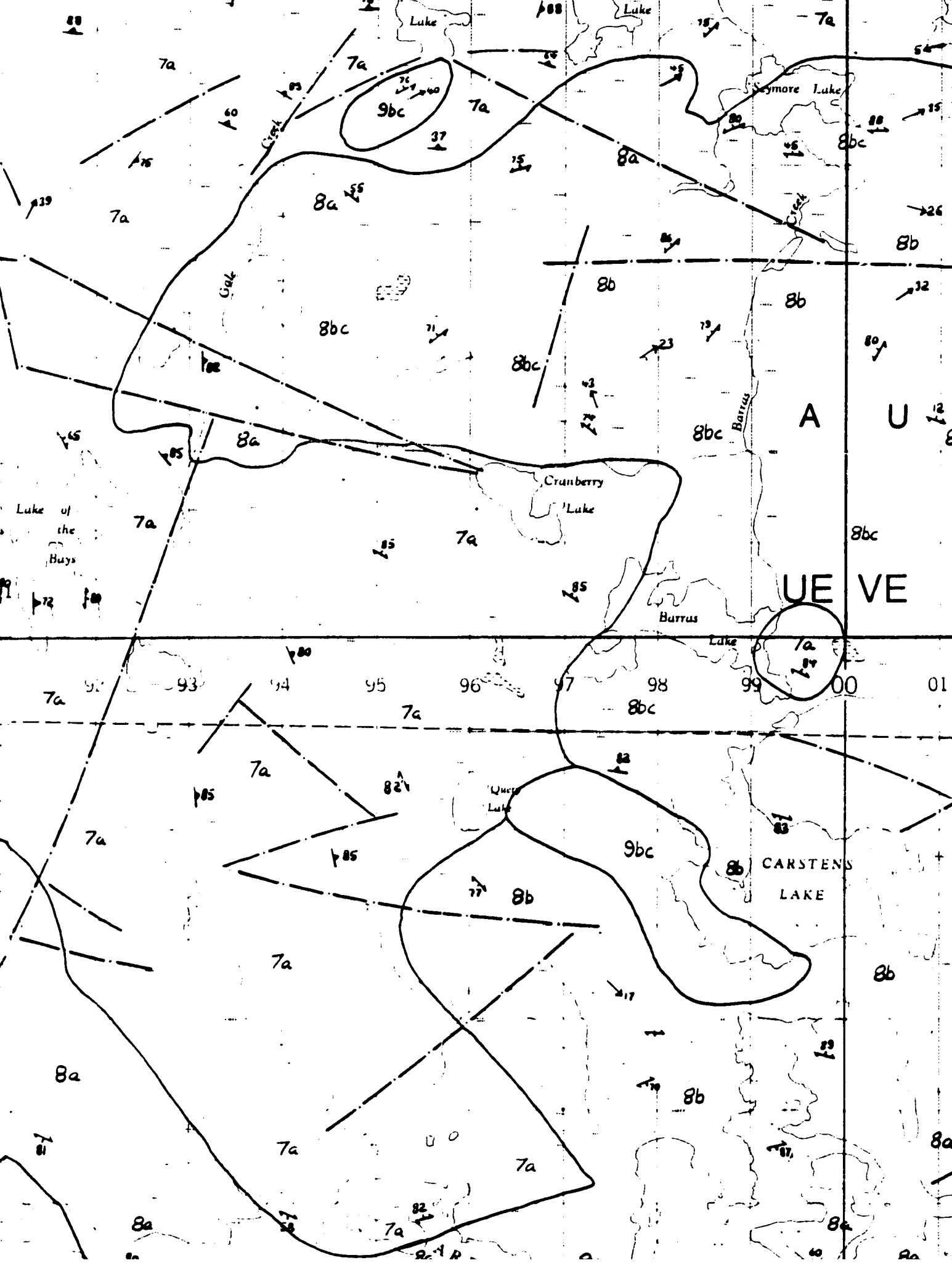


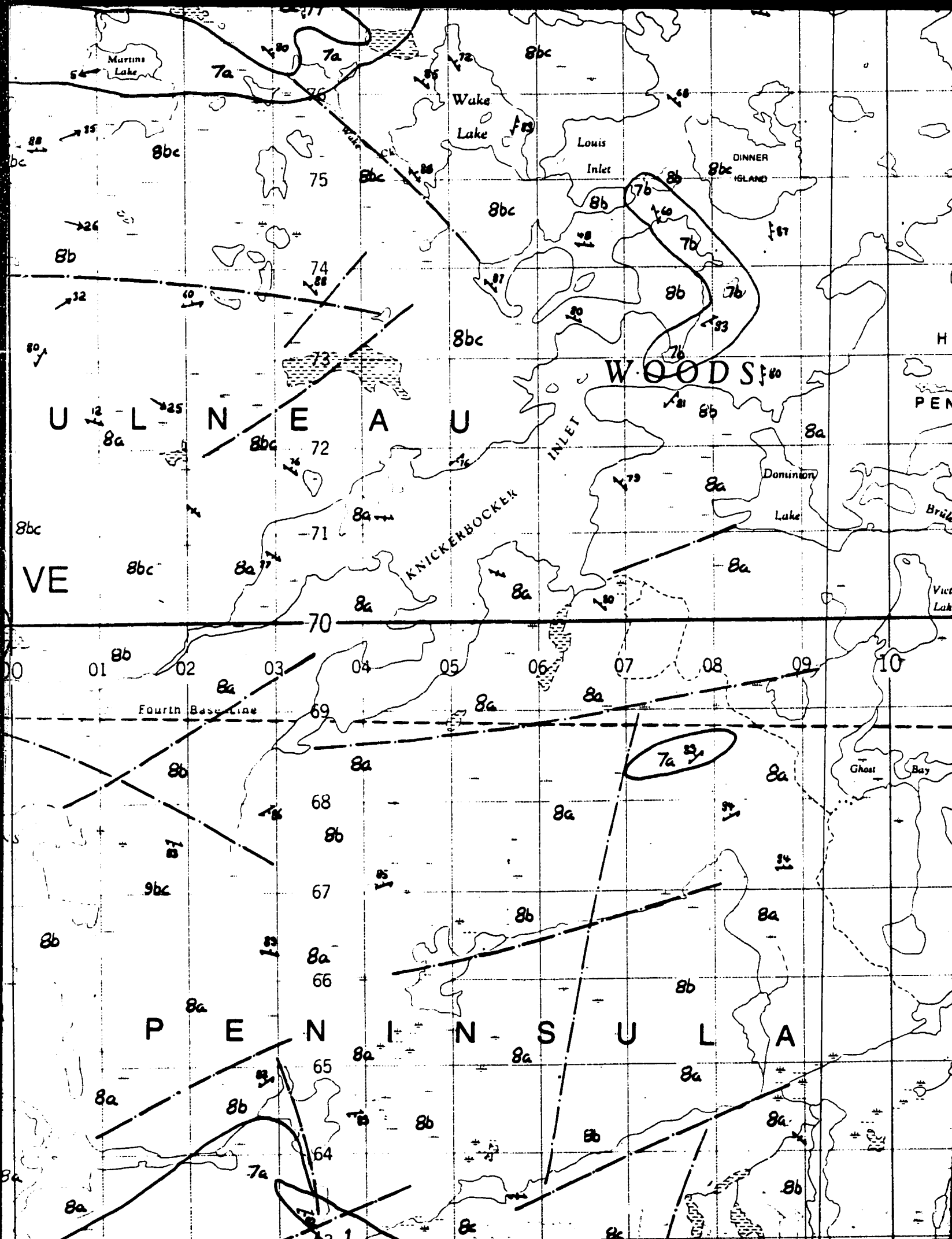


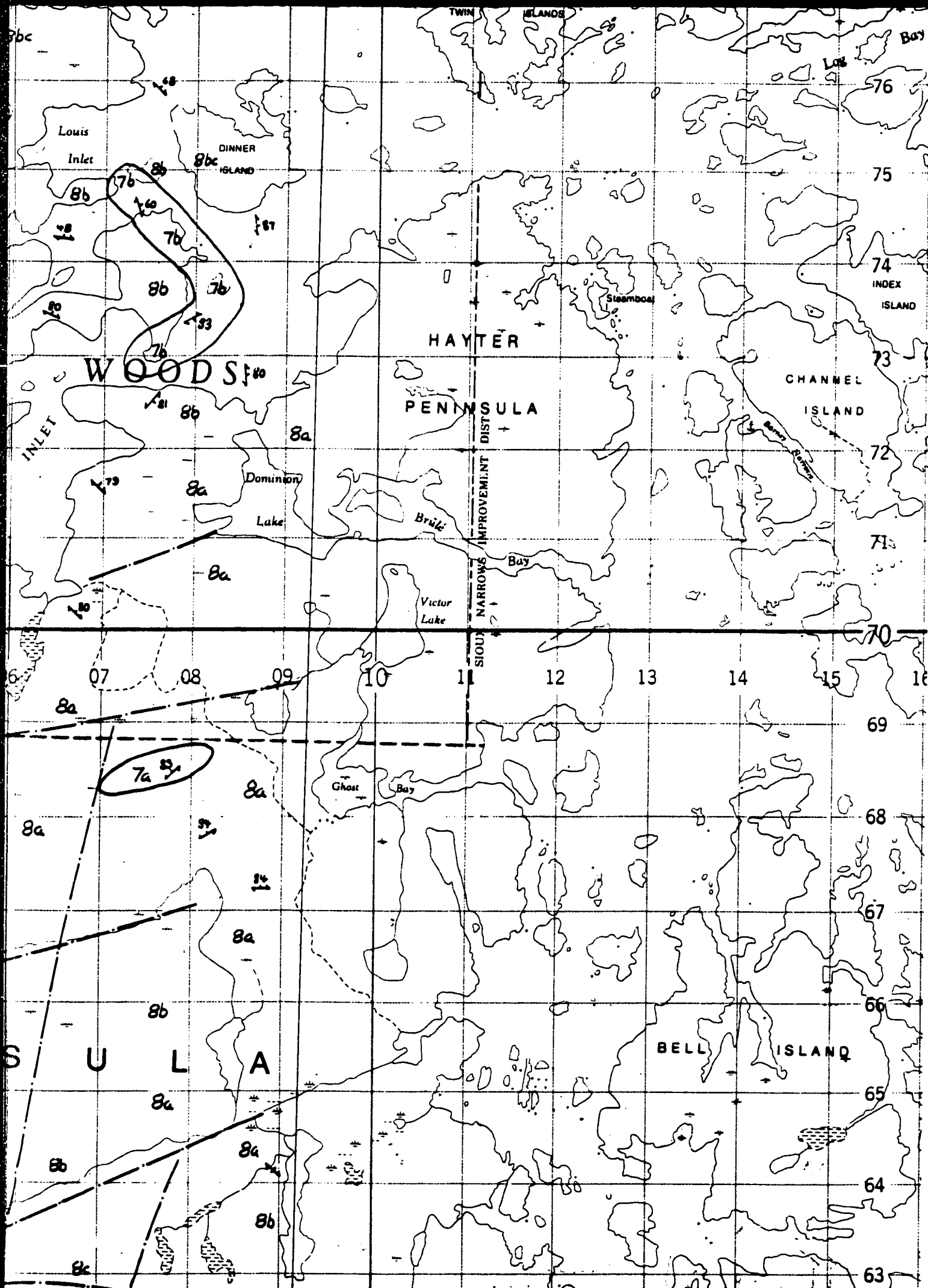


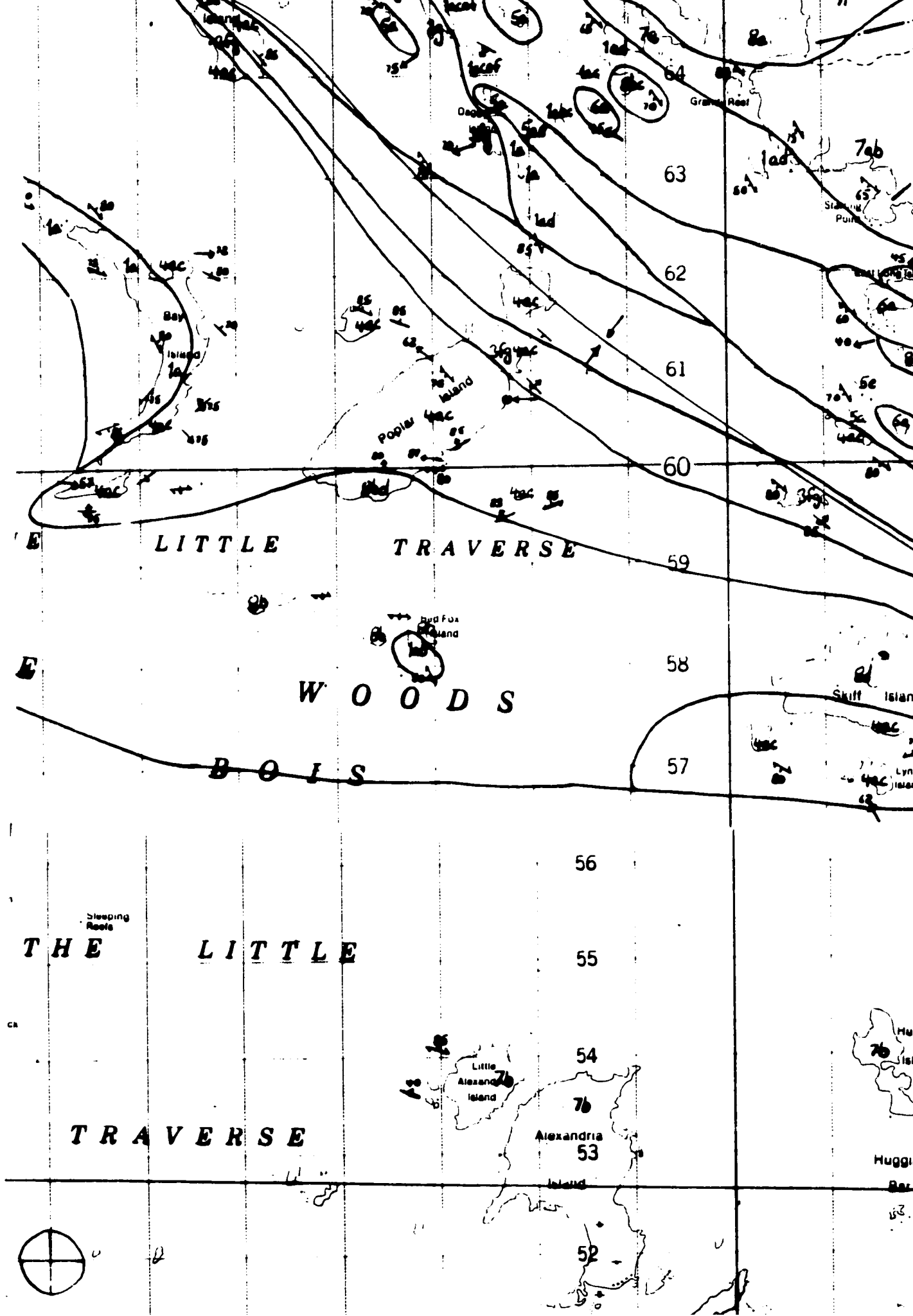


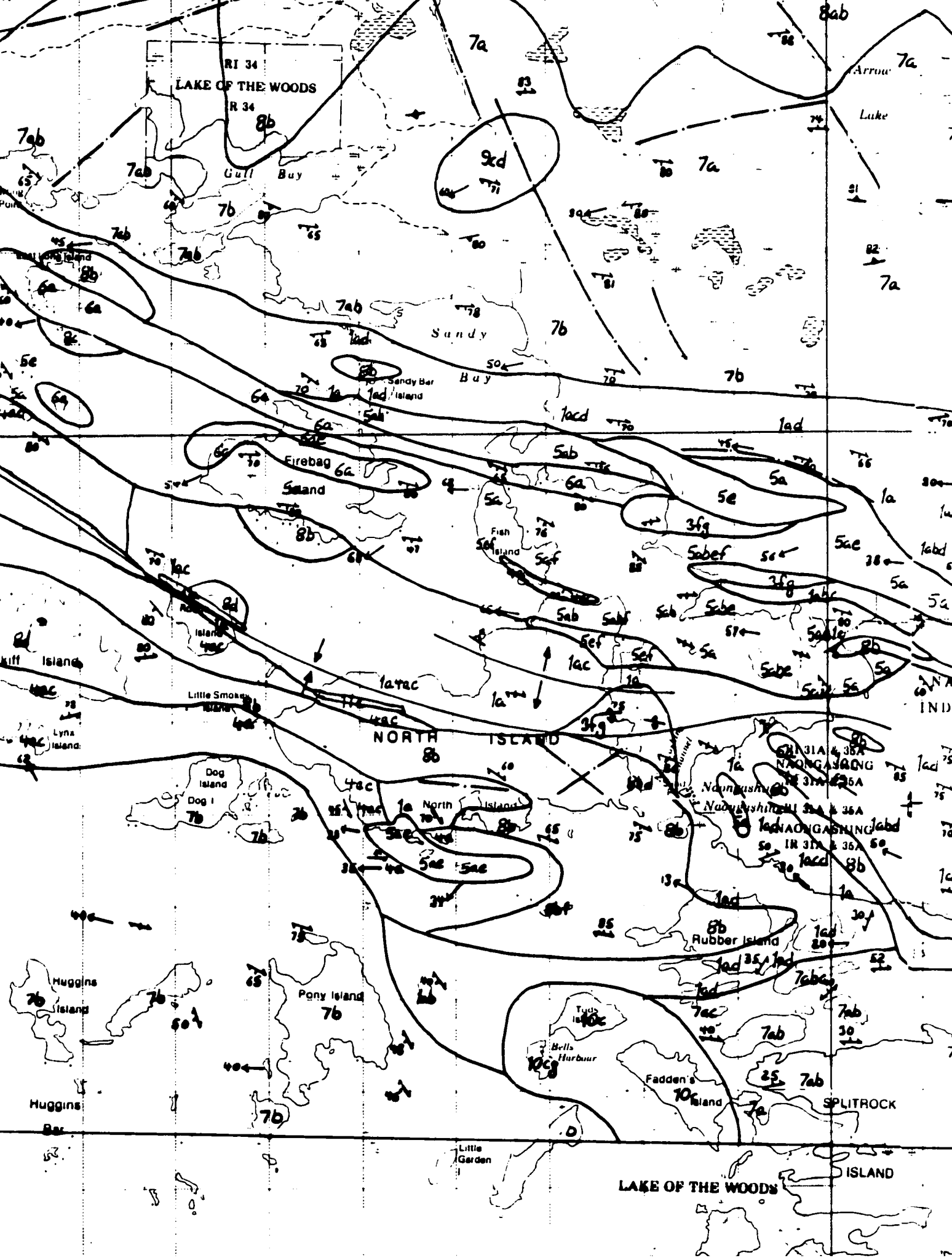


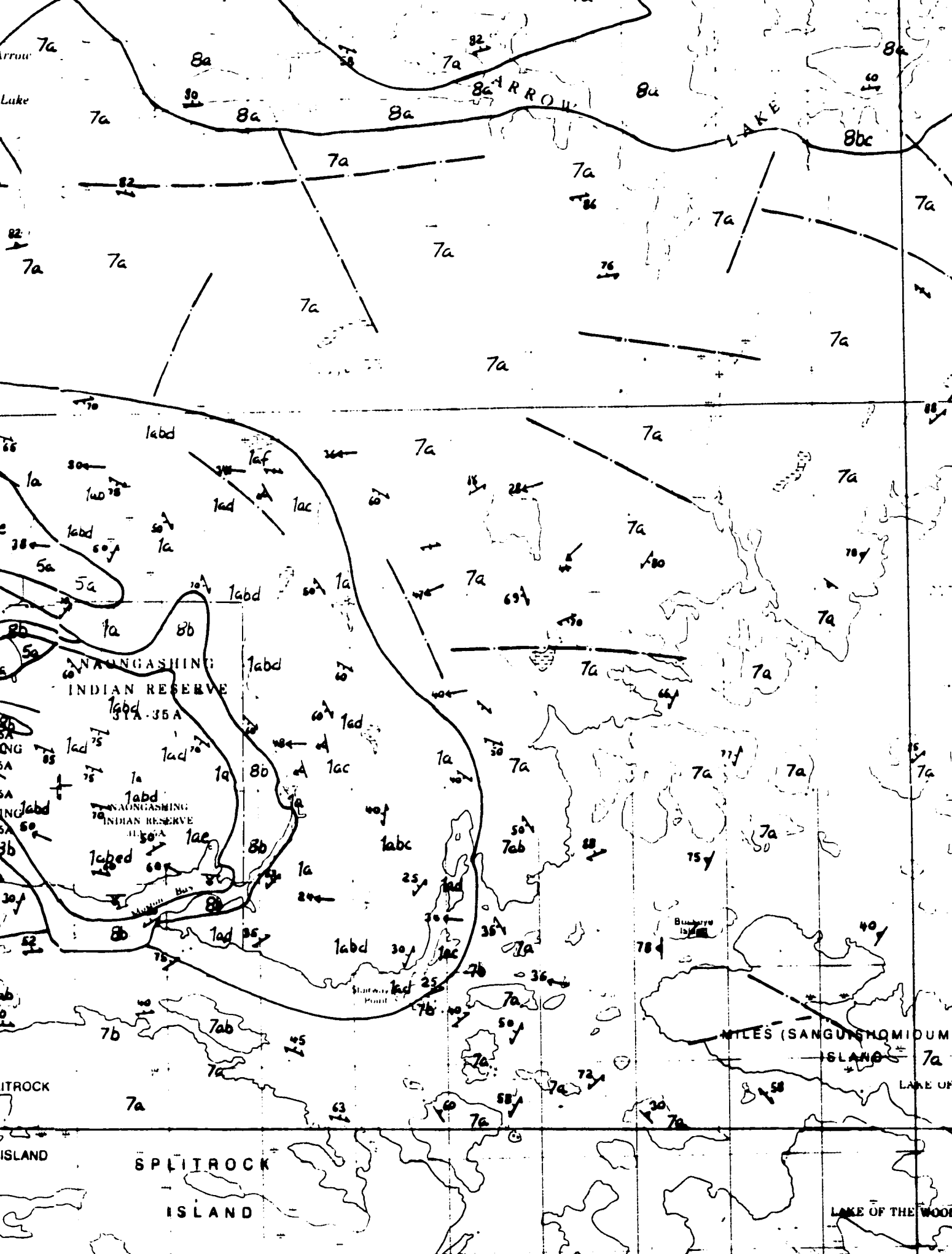


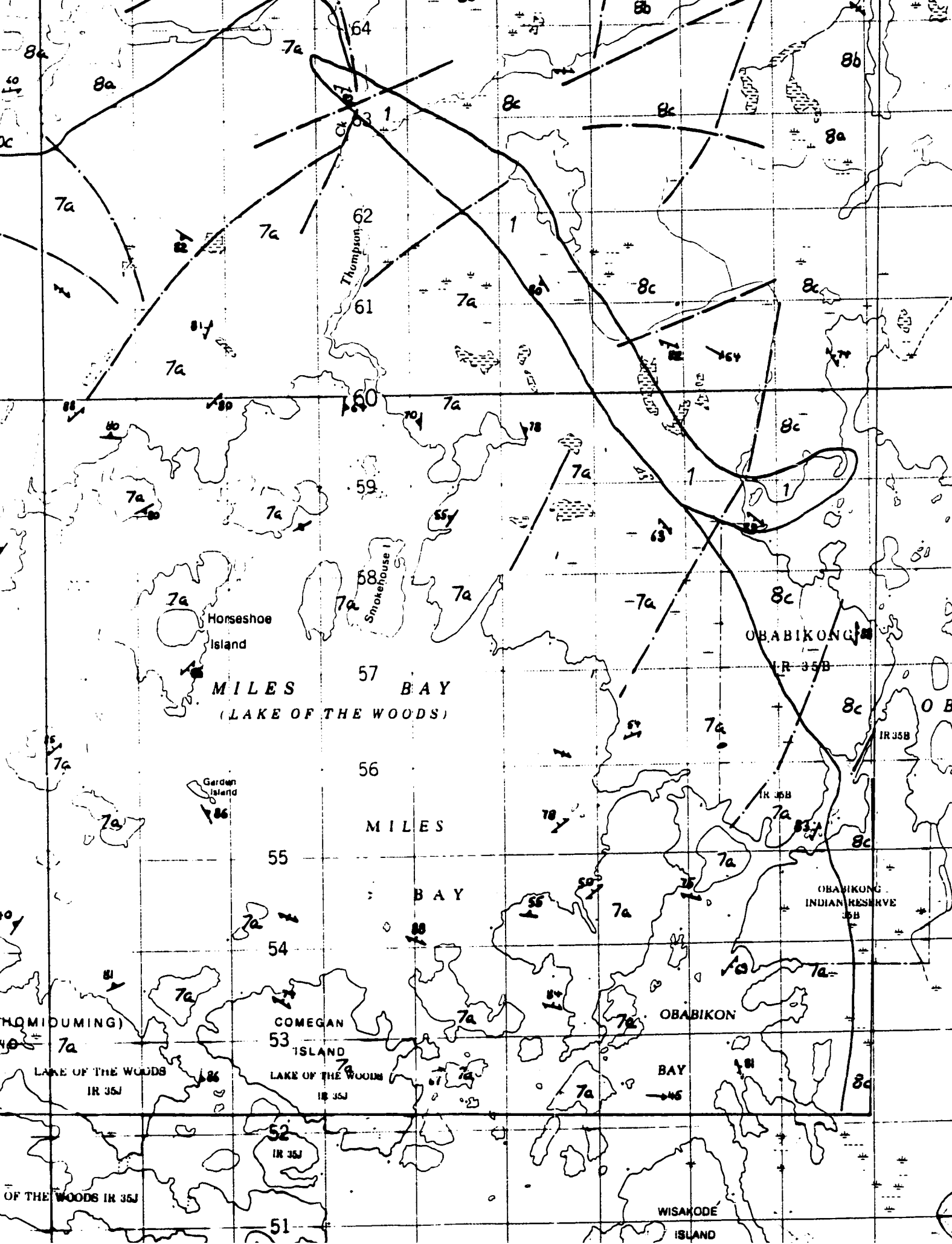


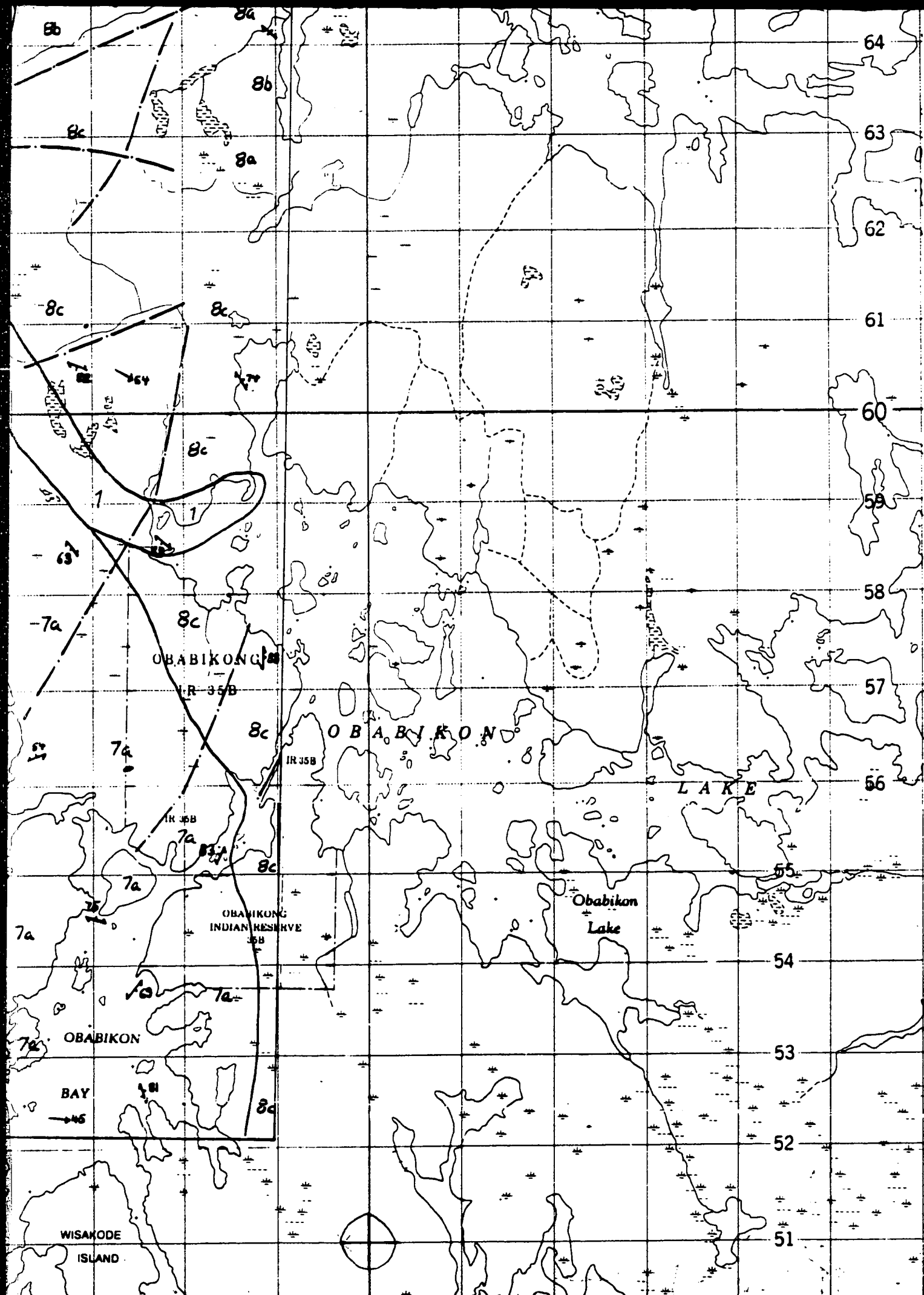












WOODS

BOIS

57

Sleeping
Reefs

THE

LITTLE

56

55

TRAVERSE

54

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Alexandria
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49

48

Seamus
Bay

47

11 31F



North
Mass

Mass

South Mass

Husters
Island

Verona
Bay

Shipwreck Bay

BIG ISLAND GEOGRAPHIC TOWNSHIP

