DEPOSITIONAL ENVIRONMENTS

of the

3.0 Ga. FINLAYSON AND LUMBY

LAKE GREENSTONE BELTS,

SUPERIOR PROVINCE,

ONTARIO, CANADA

by

DAVID KING ©

A Thesis Submitted in Partial Fulfilment

of the Requirements for the

Degree of Master of Science

Lakehead University

OCTOBER, 1996

0-612-52060-9

ABSTRACT

The Finlayson and Lumby Lake Greenstone Belts are located approximately 200km west of Thunder Bay, Ontario, north of Atikokan, Ontario.

Within both the Finlayson Lake and Lumby Lake Greenstone belts two distinct sequences of sedimentary rocks are present. Each of the belts contains an upper and a lower sedimentary rock sequence which differ in age and chemical composition.

The lower sequence of the Finlayson Lake Greenstone Belt is represented by the Little Falls Lake metasedimentary rocks and the laterally equivalent lower Finlayson Lake metasedimentary rocks. These rocks consist of coarse-grained sandstones, conglomerates, and lesser interbedded mafic detritus-rich metasedimentary rocks and are laterally continuous with felsic volcanic rocks to the south. Deposition of these sedimentary rocks was by high-density turbidity current processes. Their chemical composition is distinct from that of the upper Finlayson Lake metasedimentary rocks and suggests a single felsic volcanic source with composition similar to that of the Steep Rock Upper Felsic unit and the Old Tonalite unit. U-Pb geochronology again supports a single source rock, with an age of 2996 ± 0.8 Ma.

The upper band of Finlayson Lake sedimentary rocks are distinct from the sedimentary rocks present in the Little Falls Lake area and lower Finlayson Lake areas. Their chemical composition suggests that the upper Finlayson Lake sedimentary rocks are similar and were continuous with, although fault offset from, the upper Lumby Lake sedimentary rocks (Fenwick, 1976; Stone and Pufahl, 1995). U-Pb data from conglomerate in the southern Finlayson Lake area yield zircon ages ranging from 2997 ± 2.5 to 3002 ± 0.9 Ma. Sm-Nd data suggests that the basin received detritus derived from tonalitic intrusions as well as a slightly older mafic volcanic component. These data agree with geochemical data which suggest that the composition of the upper Finlayson Lake sedimentary rocks lies on a mixing line between the Old Tonalite and the Steep Rock Upper Mafic unit or the Finlayson Lake mafic volcanic rocks.

A well developed coarsening upward sequence is preserved within the upper Finlayson Lake sedimentary rocks. The sequence consists of iron formation and chemical sedimentary rocks at the base, overlain by DE turbidites, which coarsen to pebbly sandstones and conglomerates near the top of the sequence. There is also some lateral facies variation with coarsest-grained metasedimentary rocks exposed in the southern part of Finlayson Lake. These rocks are consistent with deposition from both high- and low-density turbidity currents and were likely deposited by a prograding delta system that was centred south of the area.

As in the Finlayson Lake Belt, the Lumby Lake Belt also contains two stratigraphically distinct sedimentary units. The lower sedimentary unit is represented by the sedimentary rocks present within the Hook Lake area, whereas the upper sedimentary unit is represented by the sedimentary rocks near Norway Lake and west to the Keewatin-Hematite Lakes area.

The lower Lumby Lake sediments are laterally continuous with 2999 Ma old felsic volcanic rocks to the east (Jackson, 1985) and are the resedimented equivalent of them. Their chemical composition is similar to that of the Little Falls Lake sedimentary rocks, and a chemically similar source is suggested.

The upper Lumby Lake sedimentary rock sequence is similar to the upper Finlayson Lake sequence and is the fault offset equivalent (Fenwick, 1976; Stone and Pufhal, 1995). The chemical composition of the sedimentary rocks suggests that the upper Lumby Lake sedimentary rocks had source rocks of the same composition as the source of the upper Finlayson Lake sediments. The upper Lumby Lake sequence is dominated by iron formations and chemical precipitates, with lesser fine-grained clastic sedimentary rocks. There are lateral facies variations from a clastic dominance in the west to chemical precipitate dominance in the east. The predominance of chemical precipitates is evidence of widespread hydrothermal activity throughout the area. It is possible that the upper Lumby Lake clastic sedimentary rocks represent the distal equivalent of the turbidite system developed in the Finlayson Lake area. Alternatively, the upper Lumby Lake portion of the basin may have been fed by a localized source centred in the Norway Lake area. Evidence of this includes a dominance of clastic sedimentary rocks and the presence of debris flow conglomerates in this area.

ACKNOWLEDGEMENTS

I would like to thank Dr. Phil Fralick for his support, insight and patience during the completion of this thesis. The project was partially funded by the Lithoprobe Project. Reino Viitala and Anne Hammond prepared thin sections and sample preparation and Elenor Jensen assisted with the preparation of samples for geochemical analysis. Noranda Mining & Exploration Inc. provided computer hardware and software, as well as encouragement to finish. Much credit goes to my wife, Nicole, who was always supportive, and provided unlimited encouragement and patience when needed.

TABLE OF CONTENTS

LIST OF FIGURES

CHAPTER 1

INTRODUCTION

1.1 Purpose

The purpose of this study was to identify the origin of the metasedimentary rocks within the Finlayson and Lumby Lakes Greenstone Belts. Investigation of the sedimentary facies, stratigraphy, and provenance was used to interpret depositional environments and suggest a possible paleogeographic reconstruction of the area.

Very little work of this type had been attempted within this area in the past. Many previous studies in the general region focused on the rocks of the Steep Rock Lake area to the southeast of the Finlayson and Lumby Lakes areas.

1.2 Location and Access

The Finlayson and Lumby Lakes Greenstone Belts are located within the District of Rainy River, northwest of Atikokan and approximately 200 km to the west of Thunder Bay, Ontario (Fig. 1.01).

The sedimentary rocks within the Finlayson Lake Greenstone Belt can be fairly easily accessed and are exposed in outcrops marginal to Finlayson and Little Falls Lakes. Finlayson Lake can be accessed by a public boat launch on

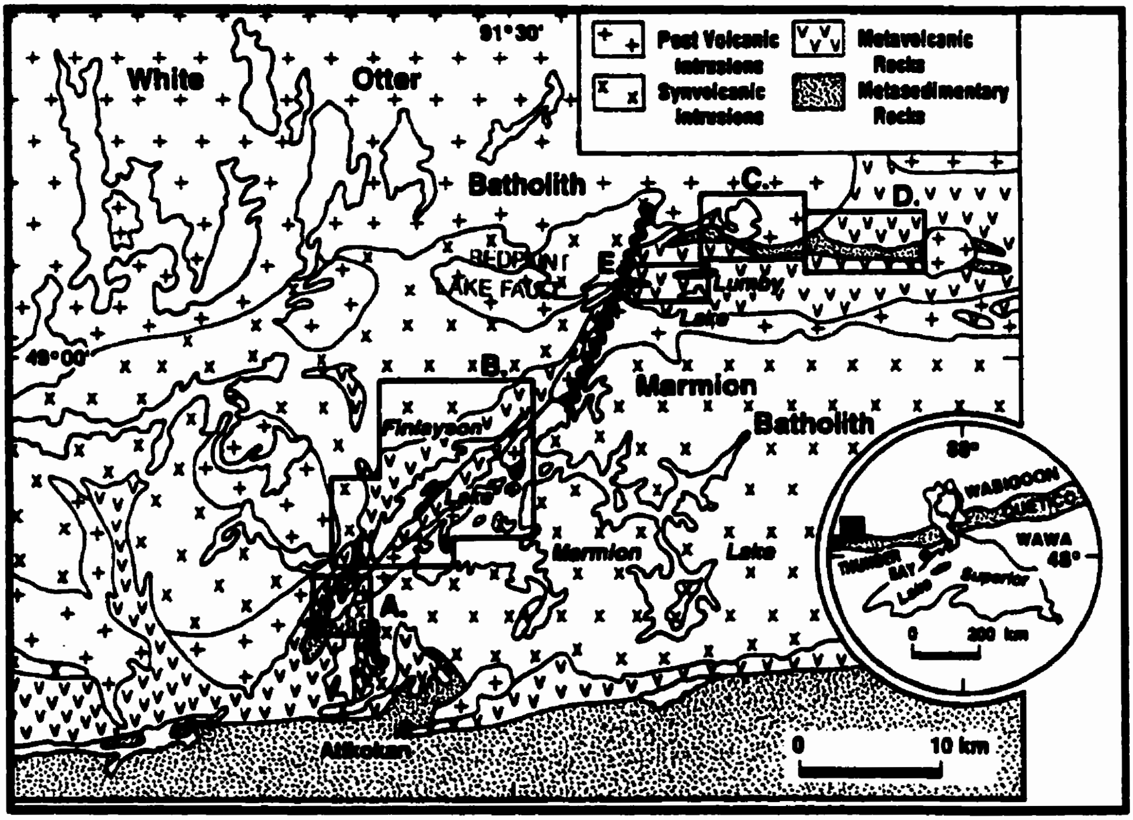


Figure 1.01 Location of the Finlayson and Lumby Lake Greenstone Belts. Letters A-E refer to specific areas studied. A-Little Falls Lake area B-Finlayson Lake area. C-Norway Lake area. D-Keewatin Lake area. E-Hook Lake area.

Highway 622, northwest of Atikokan. Water access to Little Falls Lake is possible via a boat launch on the Seine River Diversion draining Finlayson Lake. The metasedimentary rocks of the Lumby Lake Greenstone Belt are exposed on the shore of lakes within the area. Many of these lakes can be accessed by older gravel roads which branch off of the Sapawe Road. The roads are commonly in poor condition, and require either a four wheel drive truck or an all terrain vehicle. Norway Lake is accessible by float plane or by an old road coming from the north, which meets the Trans Canada Highway approximately 40 kilometres east of Ignace. This road is in very poor condition and an ATV is needed to drive all the way to Norway Lake.

A gravel road provides access to both Pinecone and Cryderman Lakes. The same road passes, within approximately 200 meters, the north shore of Magnesium Lake, which is north of Keewatin Lake. A short portage has been cut between Magnesium Lake and Keewatin Lake, and another short portage joins Keewatin and Hematite Lakes.

1.3 Previous Work

Most of the previous work within the Atikokan area has focused on the Steep Rock Lake area rather than the Finlayson Lake area. Work on the Steep Rock area dates back as early as Smyth (1891), Lawson (1912) and Uglow (1913). Jolliffe (1955, 1966) subdivided the Steep Rock Group into four main formations and suggested the Steep Rock Group unconformably overlies a crystalline basement. Detailed mapping of the Steep Rock Lake area was done for the Ontario Department of Mines by Shklanka (1972). Subsequent studies by Schau and Henderson (1983) and Wilks (1986) supported the theory that the Steep Rock Group unconformably overlies a crystalline basement. Wilks and Nisbet (1988) also suggested that the Steep Rock Group records a rifting event that extended the old gneiss terrain. The Steep Rock and Marmion Complex may have become a tilted block during listric faulting and it is possible that the Lumby Lake succession represents a second tilted block (Wilks and Nisbett, 1988).

A study of the area by Thurston and Chivers (1990) put the Steep Rock and Lumby Lake belts into a group, which they called platform sequences. These consist, from base to top, of quartz arenite ± stromatolite bearing carbonates, oxide facies iron formation and komatiitic or tholeiitic volcanic rocks. They compare these with platform sequences at passive margins, which unconformably overlie approximately 3000 Ma old granitoid rock and older mafic volcanics.

The rocks of the Finlayson Lake area have received less attention and were originally mapped in detail by Fenwick (1976) and again by Stone et al. (1992), whose study included the Steep Rock Lake area as well as the Finlayson Lake Greenstone belts. Stone et al. (1992) stated, referring to the Finlayson Lake Greenstone Belt, "sequences such as on the western side of the Steep Rock Belt, are interpreted as a restricted alluvial fan environment between an elevated volcanic source region and a distant basinal environment characterized by resedimented turbidites".

Early reconnaissance surveys were conducted in the Lumby Lake area by the Geological Survey of Canada (G.S.C.) in the late 1800s, and the area received considerable attention from prospectors in the 1890s. L. F. Kindle traversed the area in 1937 for the G.S.C. The rocks of the Lumby Lake Greenstone Belt were mapped for the Ontario Geological Survey, by Woolverton (1960) and again by Jackson (1985).

Tomlinson et al. (1996) focused a study on the petrogenesis of mafic and ultramafic rocks within the central Wabigoon Subprovince and have made broad conclusions about the origin of the rocks within the central Wabigoon region. Detailed geochemical analysis of the Dismal Ashrock, taken from the nearby, and possibly related, Steeprock Greenstone Belt has identified a deep-mantle plume source for the komatiitic rocks of the Steeprock Group with strong similarities to ocean island basalts (Tomlinson et al., 1996). The authors have suggested that the central Wabigoon region differs significantly from the western Wabigoon region, and is dominated by shallow water platform sequences and rift related volcanics. Furthermore they have postulated that the central Wabigoon represents a fragment of older (>3Ga) continental crust which has separated from the similar North Caribou terrain, also interpreted as an older shallow water platformal sequence. The geochemistry of the volcanic rocks within the Lumby Lake area suggest that a mantle plume source was significant (Tomlinson et al., 1996). However, if as they suggest, the mafic volcanics within the central Wabigoon region are related to oceanic island basalts, derived from a mantle plume source, why would a rift sequence be developed. Modern analogues of oceanic island basalts, such as the Hawaiian Island chain, do not suggest a rifting environment.

Hollings et al (1996) have studied the trace element geochemistry of the volcanic rocks within the Lumby Lake Greenstone Belt. The rare earth element patterns of the basalts within the area signify the absence of crustal contamination and are consistent with generation by a plume from undepleted mantle source. A within plate setting, possibly oceanic plateaux or anomalous ridge segments is suggested (Hollings et al, 1996). Intermediate and felsic volcanic rocks are light rare earth element depleted, consistent with slab melting within the garnet stability field and an intraoceanic arc derived from long term depleted mantle is suggested. Collectively, the volcanic rock suite suggests mafic magmatism in an ocean basin in association with evolved rocks of deep mantle origin (Hollings et al, 1996).

1.4 General Geology

Both the Finlayson and Lumby Lake Greenstone Belts lie within the interior of the Wabigoon Subprovince.The greenstone belts are surrounded on all sides by intrusive rocks of various ages (Fig. 1.02). Tonalite of the Marmion

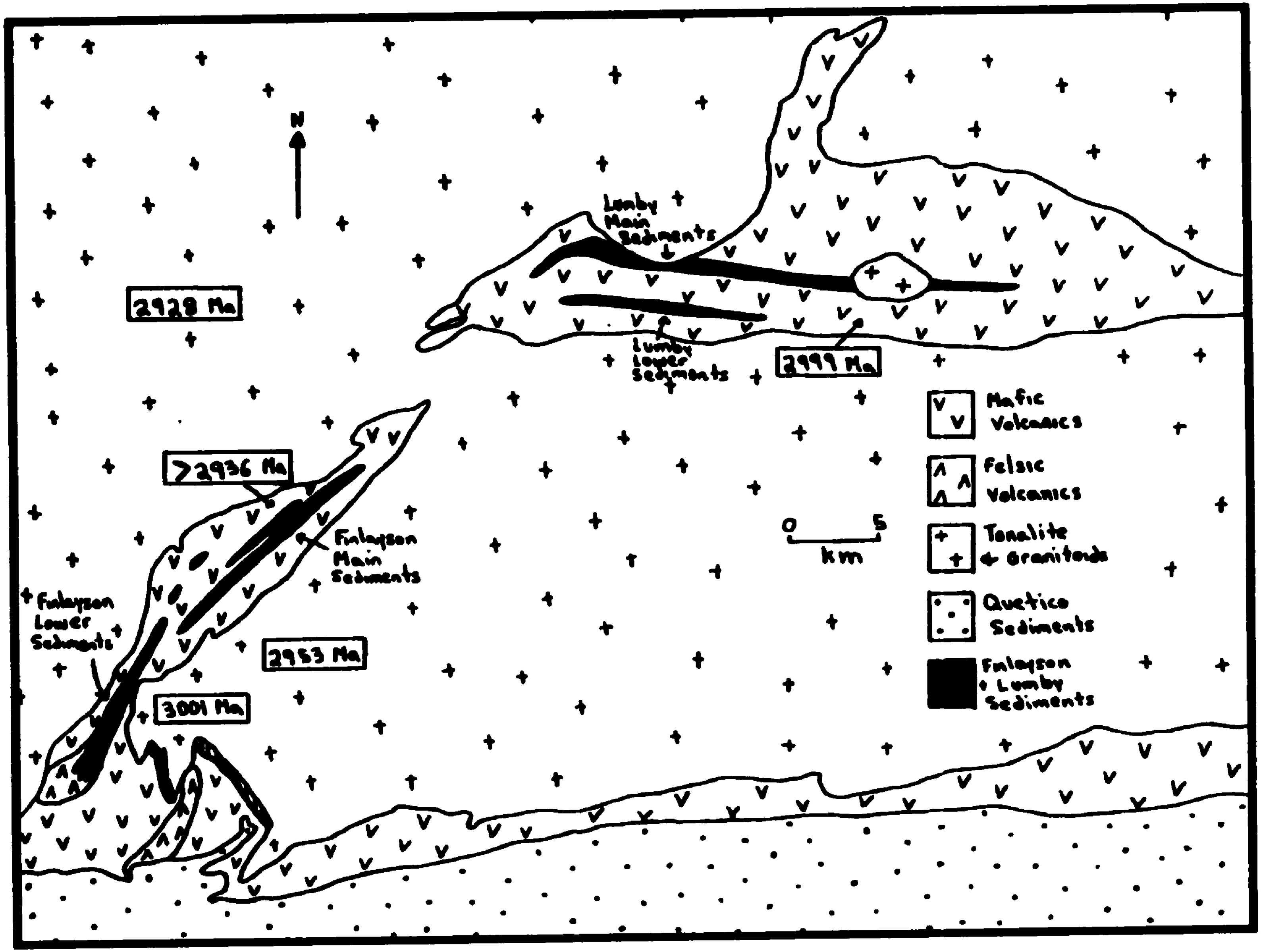


Figure 1.02 General geology of the Finlayson and Lumby Lake Greenstone Belts, showing the U-Pb ages of within the area.

Lake Batholith lies to the east of the Finlayson Lake Greenstone Belt and Tonalite Gneiss of the Dashwa Gneiss Complex occurs to the east (Stone et al., 1992). Smaller intrusive bodies such as the Hardtack Pluton, Eye-Dashwa Pluton, Lefteye Stock, and Righteye Pluton also lie to the west of the belt (Fig. 1.03).

The Finlayson Lake Greenstone Belt forms a narrow belt approximately 25 kilometres in length and 0.8-5.6 kilometres in width, with bedding and foliation striking in a northeast-southwest direction (Fenwick, 1976). The belt consists dominantly of mafic metavolcanic rocks with lesser felsic metavolcanic rocks and metasedimentary strata (Fenwick, 1976; Stone et al., 1992). Metasedimentary rocks of the Little Falls Lake area are laterally gradational into felsic metavolcanic rocks to the south and the Lower Finlayson Lake metasedimentary rocks to the north (Fig. 1.04, 1.05). The Upper Finlayson Lake metasedimentary rocks lie near the centre of the greenstone belt and, although offset by the Redpaint Lake Fault, were likely continuous with the Upper Lumby Lake metasedimentary rocks (Fenwick, 1976; Stone and Pufhal, 1995) (Fig. 1.01).

The Lumby Lake Greenstone Belt is northeast of, and laterally equivalent to, the Finlayson Lake Greenstone Belt (Fenwick, 1976; Stone and Pufhal, 1995). The two belts are offset and separated by the regional northeast-trending Redpaint Lake Fault (Fig. 1.01). The Lumby Lake Greenstone Belt trends in an east-west direction and is bordered on the south by the Marmion Lake Batholith, with the White Otter Lake Batholith to the north. The post tectonic Norway Lake

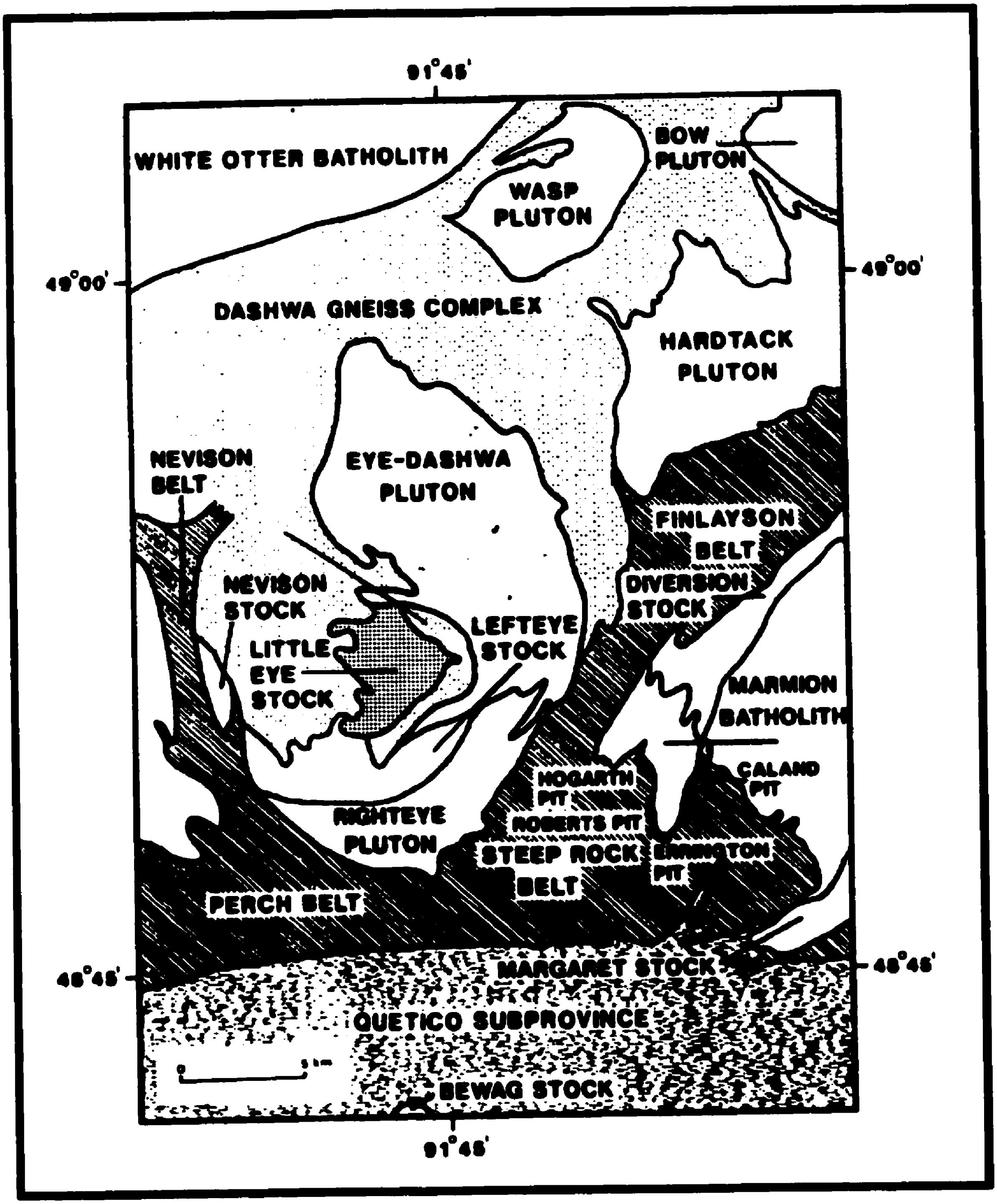


Figure 1.03 Lithotectonic domains and open pit iron mines within the Atikokan area. Reproduced from Stone et al. (1992).

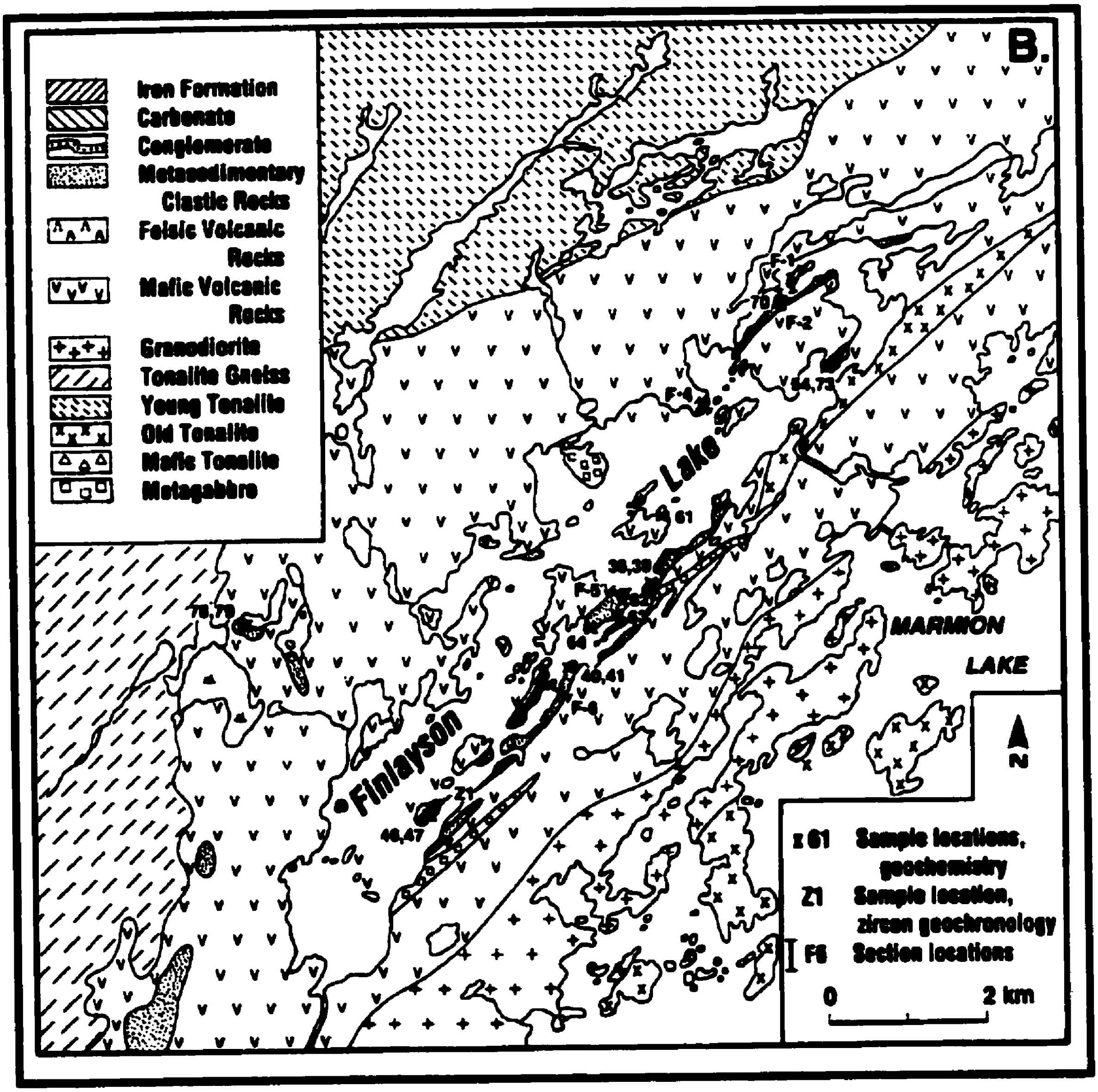


Figure 1.04 General geology of the Finlayson Lake area showing major rock units, sample locations, and stratigraphic section locations.

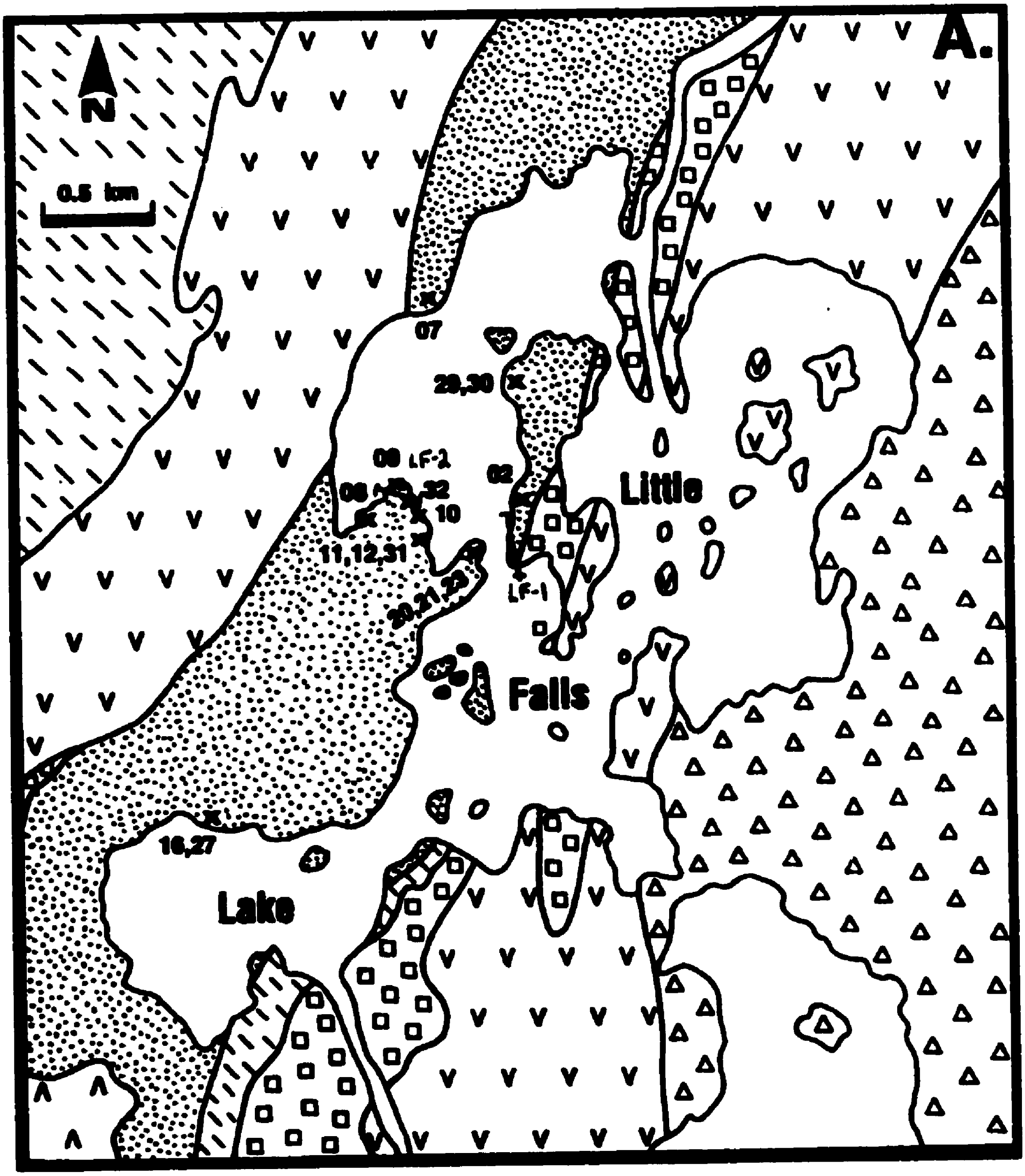


Figure 1.05 General geology of the Little Falls Lake area. Refer to Fig. 1.04 for geological legend.

Pluton and Van Nostrand Lake Stock intrude rocks of the Lumby Lake Greenstone Belt. Mafic metavolcanic rocks dominate the belt, with lesser intercalated felsic metavolcanic and metasedimentary rocks (Jackson, 1985). The Upper Lumby Lake Metasedimentary unit consists of clastic and chemical metasedimentary rocks, and is laterally continuous with the Upper Finlayson Lake Metasedimentary unit (Fenwick, 1976; Stone and Pufhal, 1995). The Lower Lumby Lake Metasedimentary unit is laterally transitional into felsic metavolcanic rocks to the east.

1.5 Metamorphism

The regional metamorphic conditions which the rocks within the Finlayson and Lumby Lakes Greenstone Belts were subjected to should be considered carefully. Metamorphic fabrics within highly metamorphosed rocks can appear similar to many primary sedimentary structures. For this reason care must be used when interpreting structures within the sedimentary rocks found in the area, and only the least deformed outcrops were useful for sedimentary interpretations. The metavolcanic rocks within the Finlayson Lake area are metamorphically zoned (Stone et al., 1992). The metavolcanic belt contains greenschist assemblages within its central portion and amphibolite assemblages near the margins. The metasedimentary rock belts are commonly found within the central portion of the greenstone belt and have been subjected to greenschist facies metamorphism.

The metavolcanic rocks within the Lumby Lake Greenstone Belt have been subjected to similar regional metamorphic conditions as those within the Finlayson Lake Greenstone Belt. The majority of metavolcanic and metasedimentary rocks in the area contain mineral assemblages representative of greenschist facies metamorphism (Jackson, 1985). Local zones of upper greenschist to lower amphibolite facies metamorphism occur adjacent to large intrusive bodies (Jackson, 1985). The contact metamorphic aureoles extend approximately 1 kilometre around the Van Nostrand Lake Stock and up to 2 kilometres surrounding the Norway Lake Pluton.

The rocks within the Finlayson and Lumby Lake Greenstone Belts have generally been subjected to greenschist facies metamorphism, with localized lower amphibolite facies developed due to contact metamorphism during emplacement of intrusive rocks. Although the metavolcanic and metasedimentary rocks within these belts have all been subjected to regional metamorphism, the prefix "meta" will be dropped for the remaining portions of this discussion in order to avoid repetition.

1.6 Structural Geology

The structural interpretation of Archean greenstone belts is often extremely difficult due to a lack of available tools, such as fossil-controlled stratigraphy and the fact that the majority of greenstone terrains have undergone multiple phases of deformation. Although the structural history of individual greenstone terrains is unique, there are several broad structural elements that are common to many greenstone belts.

Greenstone belts are distributed asymmetrically throughout all Archean cratons, however, three different end-member regional outcrop patterns are recognized (De Witt and Ashwall, 1997). Some greenstone terrains consist of broad domal granitoids with interdomal greenstone such as in the Pilbara Craton and the older greenstones of the Zimbabwe Craton. Many greenstone terrains display internally bifurcating lithological domains and irregular granitoid contacts, such as the central Slave Province with numerous mafic volcanic/plutonic belts, which display a bifurcating pattern. Greenstone terrains of this type are interpreted as fold-thrust belts (Kusky, 1989; 1991). A third common pattern consists of long, narrow metavolcanic and metasedimentary terrains alternating with gneissic rocks as is seen in the Superior Province of the Canadian Shield. It has been suggested that the distribution of greenstone belts in the Superior Province reflects the amalgamation of island arcs and interarc basins containing complex early structural fabrics affected by syn- to post-volcanic strike slip motions and pull-apart basins (Card, 1990).

The majority of Archean greenstone belts display very steep to vertical dips. These steep dips are evidence of tectonism within the greenstone terrains and the mechanisms which caused steepening are most likely different within different examples. Some areas appear to have been steepened by imbricate thrust stacking (Swager and Griffin, 1990), while others appear to have been steepened by the granitoid intrusions (Collins, 1989). Tight isoclinal folding or listric fault systems (Marshack et al. 1992) are other mechanisms of producing the steep dips observed.

In many cases structural interpretation of greenstone belts reveals several phases of deformation. Some areas may undergo early extensional or contractional deformation followed by a later phase of contractional deformation (Kusky and Vearncombe, 1997). Some areas show a progression from early recumbent folds associated with thrust/nappe tectonics, through two or more phases of tight to isoclinal upright folds. Detailed mapping of a number of greenstone belts has revealed early thrust faults and associated recumbent folds, which in many cases do not show any associated regional metamorphic fabrics or axial planer cleavage, making recognition of these early structures difficult (Kusky and Vearncombe, 1997). The Favourable Lake Greenstone belt of the Superior Province is an example of early thrust stacking of three different volcanic and sedimentary sequences over one another. This structural stacking was initially interpreted as thick intact stratigraphic sequences comprising several cycles of volcanism and sedimentation (Ayres, 1977). Detailed mapping and precise U-Pb geochronology by Corfu and Ayres (1991) revealed the early thrust stacking.

In the Pilbara Craton several phases of deformation have been interpreted within the greenstone belts and intercalculated gneisses. The complex map patterns can be explained by early thrusting of the gneisses over the greenstones, and at least two episodes of upright tight isoclinal folding which were subsequently modified by a late steep shear zones (Bickle et al., 1980; 1985).

In many cases the folds produced by late horizontal shortening are the most obvious outcrop to map scale structures in greenstone belts (Kusky and Vearncombe, 1997). It is clear from many structural studies of greenstone belts that their structural histories can be extremely complex. Although many greenstone terrains display some structural characteristics which are similar, in general, most greenstone belts have undergone a unique deformation history producing complex structural relationships within each belt.

The structure of the Finlayson Lake Greenstone Belt has been discussed by Jackson (1985), and Stone et al (1992). Younging directions based on pillow tops, graded beds and the lithological continuity of rock units have outlined several folds within the Finlayson Lake Greenstone Belt. Six folds can be identified within the volcanic and sedimentary rocks (Stone et al, 1992). The fold hinge lines trend in a north-easterly direction, with the main band of Finlayson Lake metasedimentary rocks ocuppying the core of a synform, and mafic volcanic rocks outcroping within an antiformal structure to the west (Stone et al, 1992).

Many northeast and east-southeast striking faults cut the rocks of the Finlayson Lake Greenstone Belt (Fenwick, 1976; Stone et al., 1992). The northeast striking faults are parallel to and may be a continuation of the Steep Rock Fault system (Parkinson, 1962; Fenwick, 1976). The Red Paint Lake shear zone (Woolverton, 1960) strikes northeast in the northern portion of the Finlayson Lake Greenstone Belt and exhibits left lateral shear sense (Fenwick, 1976). The fault separates the Finlayson and Lumby Lake Greenstone Belts and suggests that the Lumby Lake Greenstone Belt may be the faulted continuation of the Finlayson Lake Greenstone Belt.

Jackson (1985) suggested that the Lumby Lake Greenstone Belt forms a major synform with an axial trace located centrally in the belt. This study as well as others (Tomlinson et al., 1996) suggest that this may not be the case and that the northern and southern portions of the belt represent two distinct assemblages which differ geologically and possibly in age.

CHAPTER 2

LITHOFACIES DESCRIPTIONS OF THE FINLAYSON LAKE GREENSTONE BELT

Metasedimentary rocks within the Finlayson and Lumby Lake Greenstone Belts cover a large area, and there is extensive lateral facies variation. In order to simplify the lithofacies descriptions, the belts will be divided into several areas, and the sedimentary facies found in each area will be described. The Finlayson Lake Greenstone belt trends in a SSW-NNE direction, and can be divided into four areas, along a transect from south to north. Metasedimentary rocks in the Little Falls Lake area are distinct from those found in the Main Finlayson Lake belt. The metasedimentary rocks in the Finlayson Lake area can be divided into smaller subsections as well. The Finlayson Lake area consists of a lower sequence, likely continuous with the Little Falls Lake sedimentary rocks, and an upper sequence, which includes the main band of sedimentary rocks within the Finlayson Lake area. Due to the lateral extent of the main band of Finlayson Lake metasedimentary rocks, these also can be divided into a southern and northern region.

2.1 Little Falls Lake Sedimentary Rocks

Sedimentary rocks in the Little Falls Lake area are found within the central portion of the main synformal structure striking north-south through the area

(Stone et al., 1992). The sedimentary rock sequence has a maximum width of 2500 m, suggesting an approximate maximum thickness of 1250 m, as a large synform is present. Along the eastern margin, the metasedimentary rocks are in intrusive contact with a metagabbroic unit. On the western margin they are underlain by mafic volcanic rocks, which are laterally continuous with the mafic volcanic rocks in the Finlayson Lake area.

Metasedimentary rocks in the Little Falls lake area are dominantly massive, thickly bedded, poorly sorted, coarse-grained sandstones. These are often interbedded with monomictic conglomerate units, as well as medium-grained sandstone. Bedding contacts and younging directions in these sedimentary rocks are often indistinct, owing to their massive, non-graded nature.

Clastic Sedimentary Rocks

Coarse Granular Sandstone

This facies consist of poorly sorted, medium- to coarse-grained arkosic sandstones (Fig. 2.01), which contain abundant feldspar and quartz granules scattered throughout. Granules are angular to subangular and average approximately 0.2 cm in diameter, but may be up to 0.5 cm. The immaturity of these sands is suggested by their large feldspar content, and angular granules.

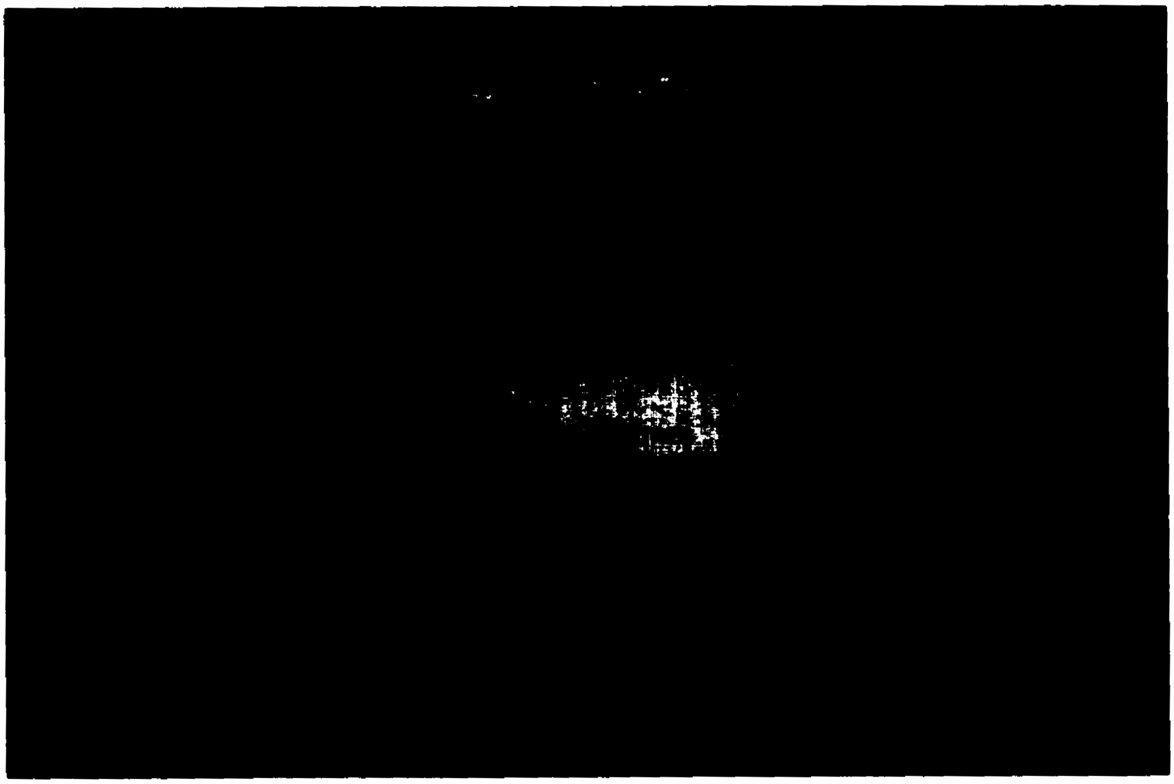


Figure 2.01 Coarse-grained sandstone facies present within the Little Falls Lake area, showing shale rip-ups within the bed.

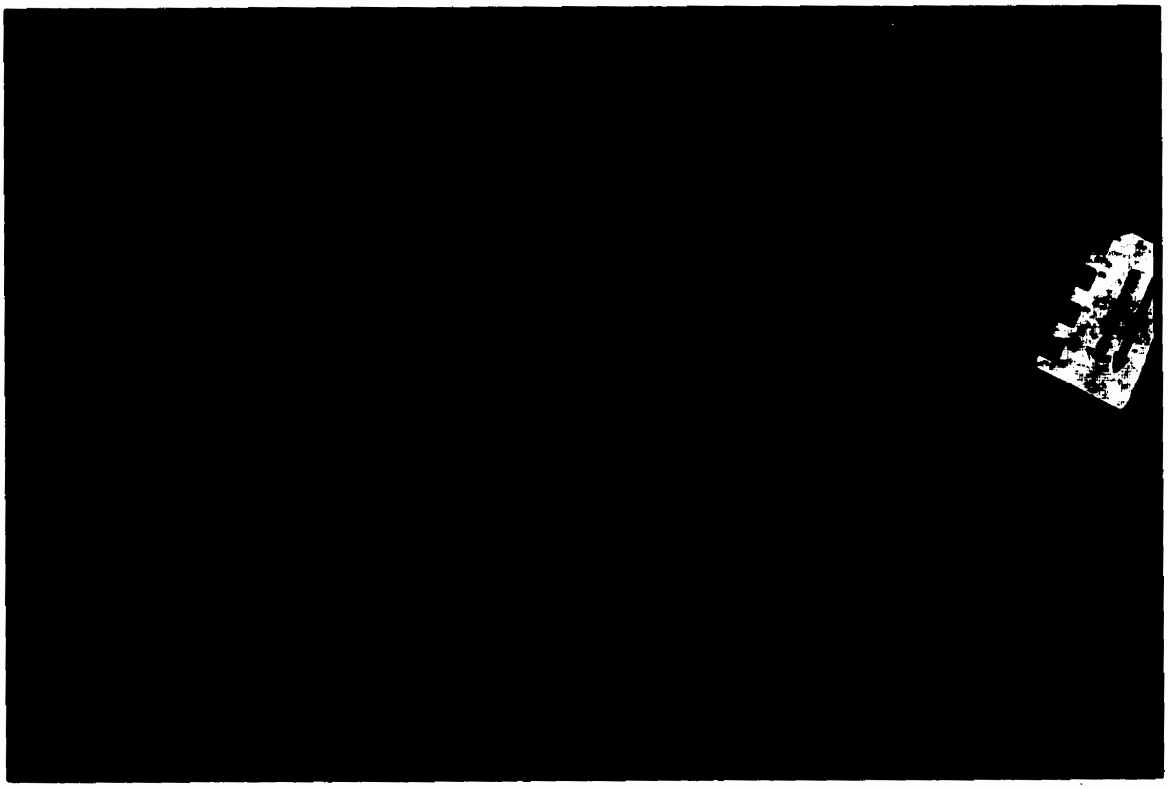


Figure 2.02 Medium-grained sandstone facies present in the Little Falls Lake area. Decimeter scale bedding and thin shale interbeds are visible.

Where discernible, bedding contacts are sharp and recognised either by a thin shaley parting, usually < 1cm thick, but up to 12 cm, or slight grainsize changes between beds, commonly a gradation to a few centimeters of medium-grained sandstone at the top of the bed. In many cases grainsize changes are reflected by differential weathering on the outcrop surface. Bed thicknesses are typically 50 - 100 cm, ranging from 20 cm to approximately 600 cm. The thicker units are commonly fractured along bedding parallel planes spaced 50 - 100 cm apart, which may represent bedding contacts that are indistinct and lack a recognisable shale interbed.

The granular sandstone facies is most commonly massive, lacking any well developed internal structures. Rarely small-scale trough cross stratification is present, recognised by thin, curving clay-rich laminae that define the troughs. Cross laminae are a few millimetres thick and organised into approximately 30 cm thick co-sets. Slightly coarser-grained areas within sandstone beds may reflect grainsize changes within individual troughs, where the cross-stratified laminae are not visible. Minor basal scours, a few centimetres into the sandstone of the underlying beds, are present, although rarely. Sandstones may contain small shale rip-up clasts, up to a few centimetres in length, and thin mud-rich flasers, O.1 cm thick. Although rare, these features can, in some cases, be used to define younging directions in the sandstones.

Bedding planes appear continuous across outcrops a few meters in length, with little evidence of thickening or thinning of the beds.

Coarse-grained granular sandstones are commonly found associated with conglomerate facies and/or medium-grained sandstone facies, and, on a large-scale, are most likely transitional to both.

Medium-Grained Sandstone

Medium-grained sandstones are similar to the coarse-grained granular sandstone facies in character, however they lack the abundant granules (Fig. 2.02). They are also similar to the coarse-grained granular sandstones in having a very low clay content. This facies consists of medium-grained sandstone, rarely containing scattered granules at the base of beds which abruptly disappear upward in the bed. Medium-grained sandstones also show sharp upper and lower bed contacts, apparent through slight grainsize changes, or thin shaley partings between layers. Some degree of grainsize gradation is more common in these sands, often from a medium-grained sandstone base, to a fine-grained sandstone top, which is useful as a younging indicator.

Bedding thicknesses range from 10 cm to approximately 100 cm, with the commonest thickness in the 20 - 30 cm range. Thicker units, greater than 100 cm, in which bed contacts are not readily identified, commonly show slight colour changes, reflecting variation in grainsize, which may indicate bedding in the 20 - 30 cm thickness range.

These sands are most commonly massive, and contain rare shale rip-ups and scoured bases. A coarse-grained sandstone lens, thinning from 60 cm to 5 cm, over a 100 cm lateral distance, is present within one thick, medium-grained sandstone bed.

Monomictic Conglomerate

The Little Falls Lake conglomerates are clast to matrix supported, with an arkosic sandstone matrix, similar to the coarse-grained sandstone facies found in the Little Falls Lake area (Fig. 2.03). Matrix sandstone is medium- to coarse-grained and commonly contains subangular granules of quartz and felsic rock fragments, up to a few millimetres in diameter. Clasts within conglomerates are dominantly pebbles and cobbles, commonly ranging in size from approximately 1 x 0.5 cm to 20 x 5 cm, with rare boulders up to 30 x 20 cm. The clasts are irregularly shaped and are subangular to subrounded. The conglomerates are monomictic with felsic volcanic clasts.

Conglomerates are generally poorly-sorted, although thicker units exhibit some internal organisation into clast-rich and clast-poor zones, or zones with differing clast sizes. These zones, as well as sandstone lenses within conglomerate, suggest that the thick conglomerate units are composed of smaller 50 to 100 cm thick lenses. The boundaries between the lenses are indistinct, and commonly gradational over a few centimetres. Conglomerate units range from 12 to 700 cm in thickness. The conglomerate units within the

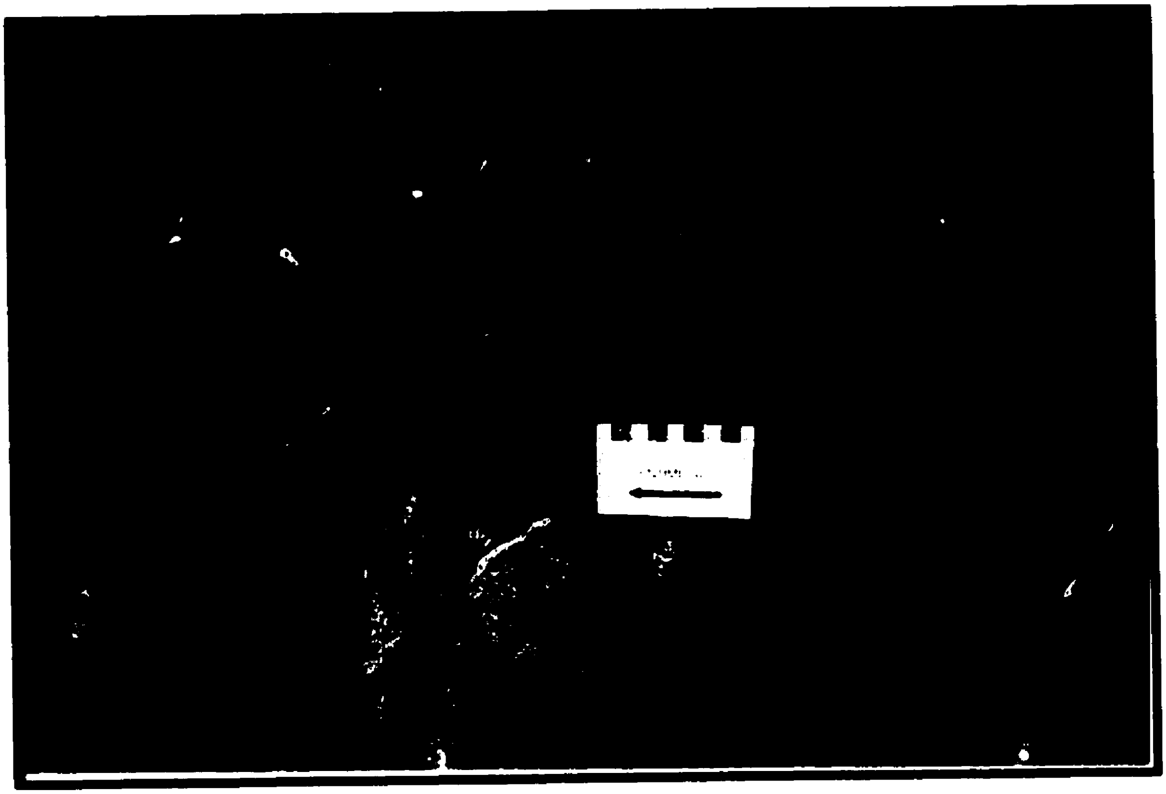


Figure 2.03 Monomict conglomerate facies present in the Little Falls Lake area, showing sub –angular, poorly sorted, felsic volcanic pebbles and cobble.

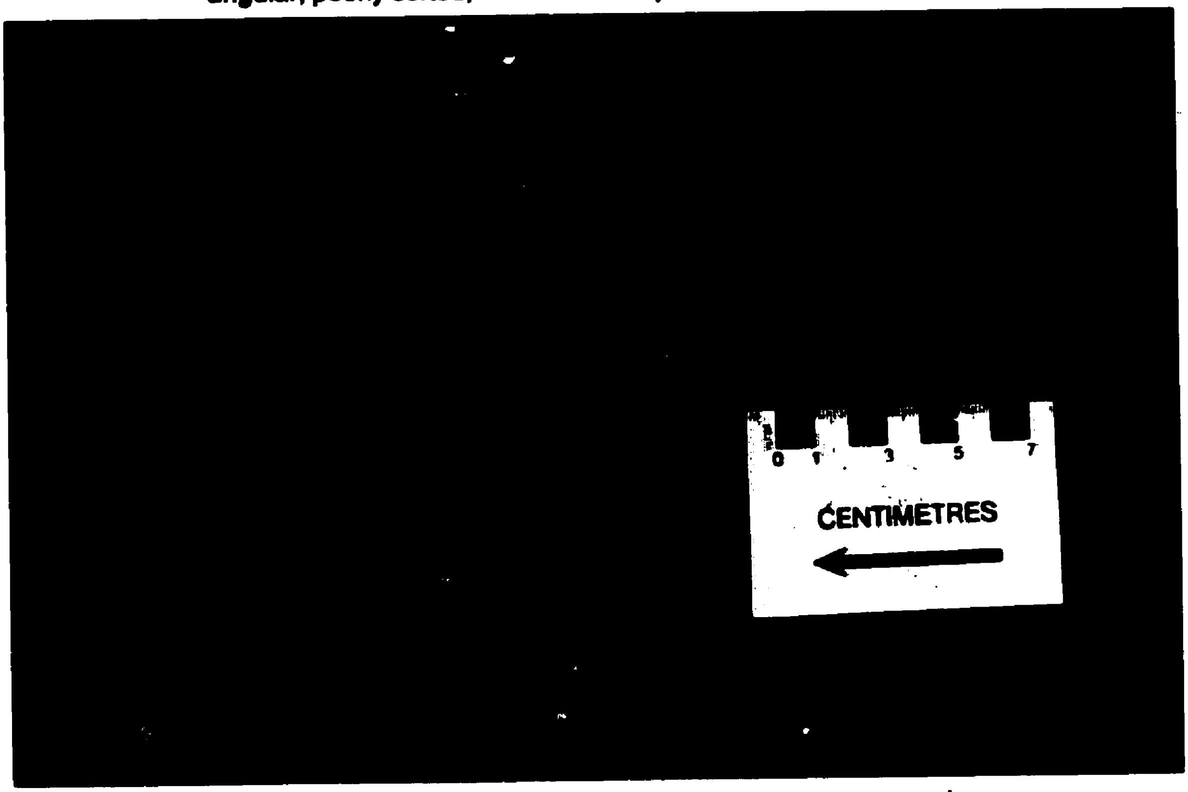


Figure 2.04 Lapilli tuff facies present in the Little Falls Lake area showing fine-grained mafic matrix, and scattered lapilli size fragments.

Little Falls Lake area are always associated with the arkosic sandstone facies. Contacts between beds of these two facies are dominantly sharp. Conglomerate rarely grades upwards into coarse granular sandstone.

Mafic Detritus-Rich Sandstone

Mafic detritus-rich sandstones, which are somewhat darker in colour than the arkosic sandstones, are also present in the Little Falls Lake area. These are fine- to medium-grained and more chlorite-rich than the arkosic sandstone facies. Their higher chlorite content suggests a more mafic source than that of the arkosic sandstones found in the area.

The mafic detritus-rich sandstones are dominantly non-graded. However, they rarely show an abrupt gradation to a finer-grained sandstone top within the upper few centimetres of the bed. Bed thicknesses range from approximately 30 to 250 cm, with an average in the neighbourhood of 100 cm. Both upper and lower bedding contacts are sharp. The mafic detritus-rich sandstones are found interbedded with the coarse-grained, granular and the medium-grained, arkosic sandstones.

Lapilli Tuff

The lapilli tuff facies consists of a fine-grained, mafic ash matrix containing lapilli size fragments of felsic volcanic material in matrix support (Fig. 2.04). Lapilli are subrounded, irregularly shaped and range in diameter from a few millimetres to approximately 2 cm. Lapilli are poorly-sorted and are scattered throughout the beds, or they occur in lenses within individual tuff beds. Contacts with adjacent units are always sharp. The strained thickness of lapilli tuff beds ranges from 10 to 100 cm.

Shale

Thin beds of silty shale separate some of the thicker, coarser-grained beds in the Little Falls Lake area. Shale units are usually very thin, <1 cm, but may be up to 12 cm in thickness. Silty shales are non-graded and no internal structures are visible. Contacts with the coarse-grained facies are commonly quite sharp, and rarely gradational. The gradation does not contain a complete series of intervening grainsizes but rather represents fine silt grains infiltrating between the coarse sand grains within the top few millimetres of the underlying bed. The shale facies is rarely present as small rip-up clasts contained near the base of some of the coarse-grained sandstone beds.

2.2 Lower Finlayson Lake Sedimentary Rocks

Isolated pockets of fine- to medium-grained sandstone are found within the mafic volcanic rocks just west of the western shore of Finlayson Lake. These sedimentary rocks are similar to the medium-grained sandstones found in the Little Falls Lake area. Poor exposures within the area prohibit a thorough description of these sandstones.

Fine- to Medium-grained Sandstone

The sedimentary rocks in the lower Finlayson Lake band consist of thin units of fine- to medium-grained sandstone. Bedding contacts are difficult to recognise and are in some cases visible through slight grainsize changes between individual beds. Where discernible the sandstones appear to be bedded on a medium scale, ranging from approximately 10-25 centimetres in thickness. Sandstones are poorly sorted, non-graded, and no internal structures were visible.

2.3 Southern Finlayson Lake Sedimentary Rocks

The main band of sedimentary rocks within the Finlayson Lake area is laterally extensive. To simplify descriptions it is divided into southern and northern portions. The sedimentary rocks occupy the centre of the main synclinal axis that trends parallel to the length of the Finlayson Lake greenstone belt (Stone et al., 1992). The sedimentary rocks are in conformable contact with, and are underlain by mafic volcanic rocks to the east of the sedimentary package.

The sedimentary sequence has a maximum width of approximately 500 m. The sedimentary package conformably overlies the mafic volcanic rocks on the eastern contact and displays a well developed coarsening upward trend. Iron formation and fine-grained sandstones directly overlie the volcanic rocks, and the sedimentary sequence coarsens upward into coarse-grained sandstones and conglomerates to the west.

Clastic Sedimentary Rocks

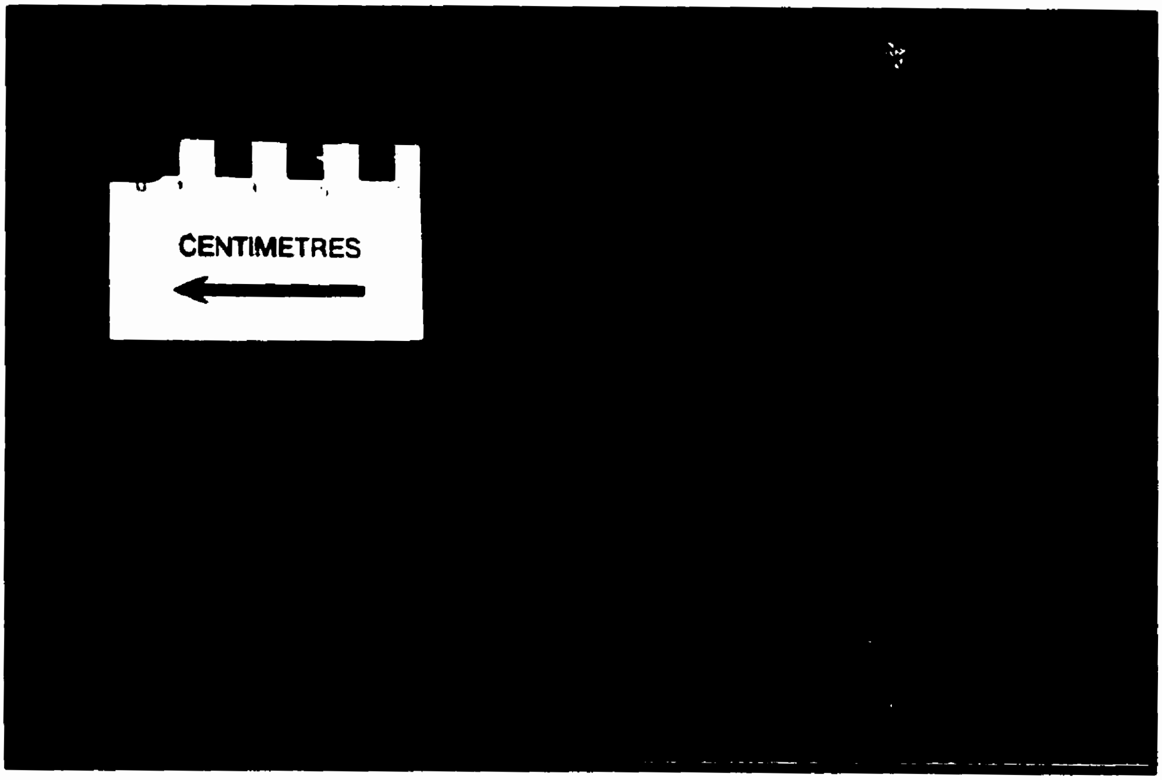
Fine-grained Sandstone and Siltstone

Fine-grained sandstone and siltstone couplets are common in the Finlayson Lake sedimentary belt. These are thinly bedded, ranging in thickness from <1 cm to approximately 5 cm, with an average of 2-3 cm. Beds are composed of a fine-grained sandstone base and a silty, clay-rich top. The coarser-grained base makes up between 50 and 75 % of the total bed thickness. Each unit begins with a sharp lower contact, overlain by the fine-grained sandstone base. There is usually a rapid gradation upwards over a few millimetres in the upper portion of the layer, from the fine-grained sandstone base to the clay-rich top of the bed. In other cases the gradation may be so rapid that the sandstone and siltstone components appear to be two distinct layers. In many beds well developed, millimetre scale, parallel laminations are visible in siltstone and claystone layers forming the upper portion of the beds.

Medium- to Coarse-Grained Sandstone

There is a continuous gradation between the fine-grained sandstone and siltstone beds to the medium-grained sandstone. Individual beds are composed of a medium- to coarse-grained basal portion, which commonly shows a well developed gradation to a fine- or medium-grained top (Fig. 2.05). The upper section of these beds may also show millimetre-scale, parallel laminations composed of fine-grained sandstone and siltstone. The coarse-grained base usually comprises 50 - 75 % of the total bed thickness. In some cases the uppermost portion of the bed may be composed of a few centimetres, or less, of shale. The shaley top of these beds is commonly very thin, and rarely well developed.

Bed thicknesses are between 5 cm and 40 cm with an average thickness of 15 cm - 20 cm. The grainsize of individual beds is usually proportional to the bed thickness, with the thicker beds having a coarser-grained base than the thinner beds. The coarsest portion of the bed may contain small granules that are usually 1 - 5 mm in diameter, but may be up to small pebble size. Granules are



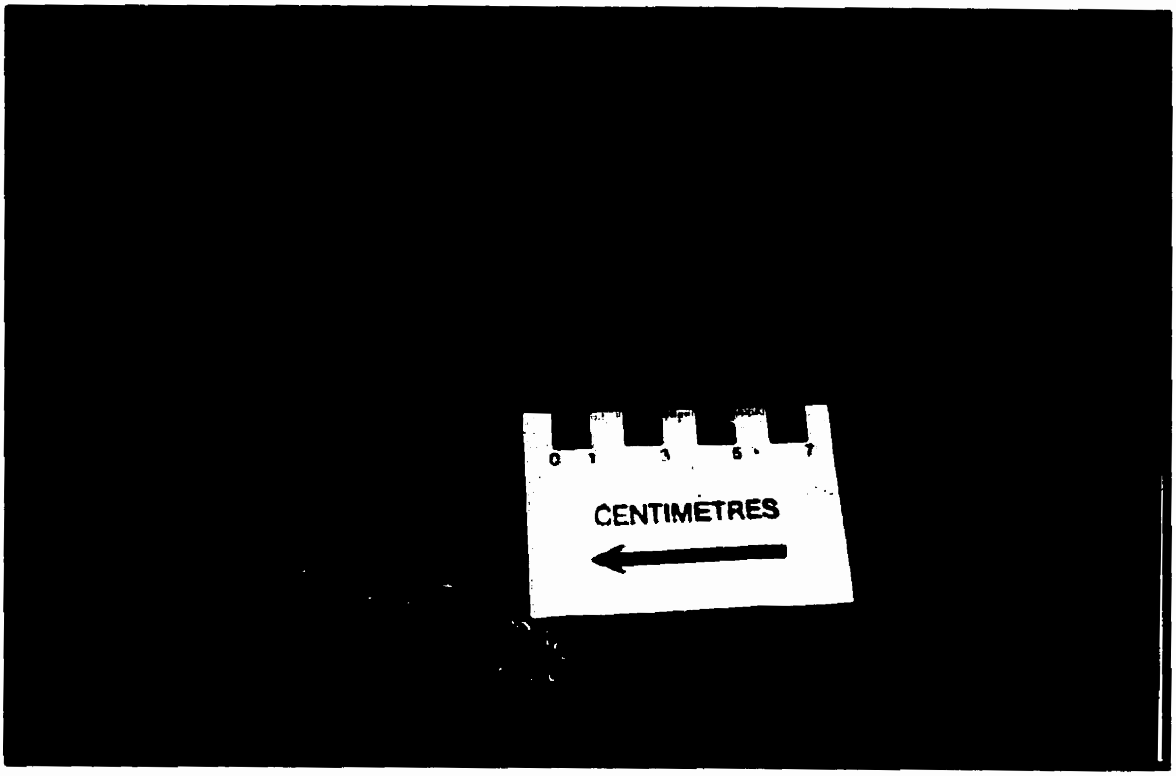


Figure 2.05 Medium- to coarse-grained sandstone facies present in the Finlayson Lake area. Showing well developed grading within beds and sharp upper and lower contacts.

composed of quartz, and granitic plus felsic-volcanic rock fragments. The coarse-grained base may also show well developed coarse tail grading of the granules throughout, making younging direction very apparent.

Beds commonly have very sharp upper and lower bed contacts with rare scouring into the underlying bed. Coarse-grained bases are occasionally loaded into the silty-clay top of the underlying unit. Beds also contain rare shale rip-up clasts, approximately 2 x 20 cm, within their basal portion.

Medium- to Coarse-Grained Scoured Sandstone

This facies is similar to the medium- to coarse-grained sandstones, but contains scallop shaped scours within the coarse-grained basal portion of the beds (Fig. 2.06). These have a medium- to coarse-grained base which often shows a quick gradation to a fine-grained sandstone top. The base is composed of a number of lenses, which are separated by clay-rich finer-grained sandstone streaks. The lenses range in size from approximately 2 x 12 cm, to 5 x 50 cm. Lenses are filled with medium-grained or coarse-grained sandstone often containing granules up to a few millimetres in diameter. Granules are composed of quartz and granitic plus felsic-volcanic rock fragments. Clay-rich streaks that separate the lenses are usually less than 0.5 cm in thickness. Bed thicknesses of the sandstones ranges from approximately 10 - 30 cm.

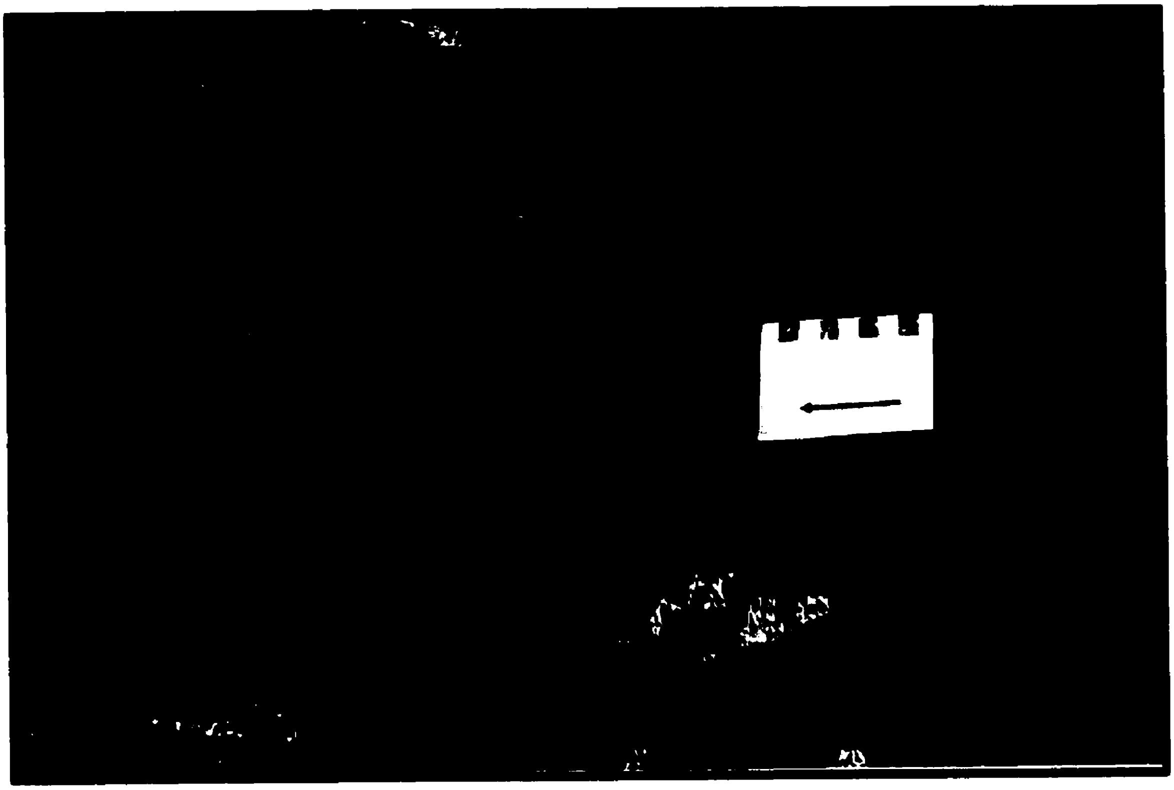


Figure 2.06 Coarse-grained scoured sandstone facies present within the Finlayson Lake area. Scallop shaped scour is visible to the left of the scale card.

Figure 2.07 Magnetite-chert iron formation present in the Finlayson Lake area. Brecciated chert layers within a dark black magnetite rich matrix.

Conglomerate

Conglomerates within the Finlayson Lake area are polymictic with pebble to cobble sized clasts. Diameters of the clasts ranges from less than 1 cm to 20 cm, with an average diameter of approximately 2 - 3 cm. Clasts are dominantly rounded to subrounded and composed of granitoid rock fragments, having quartz crystals up to a few millimetres within a felsic matrix. Lesser felsic volcanic and mafic volcanic rock fragments were also present in the Finlayson Lake conglomerates. Conglomerates are non-graded and the sorting of clasts is moderate to poor. The majority of conglomeratic units are clast supported, with a medium-grained sandstone matrix, however, there is a complete gradation to matrix supported pebbly sandstone beds.

Conglomerate units are up to approximately 4 m in strained thickness. In most cases the thicker units are composed of smaller wedge shaped lenses, that average approximately 20 cm in thickness and are up to 50 cm thick. The lenses are defined by clast-rich and clast-poor areas. In addition, coarse-grained sandstone lenses occur within conglomerate units. Conglomerate beds appear to be of random thickness, and are not organised into any pattern of beds of similar thickness. Sandstone lenses are approximately 3 - 10 cm thick and often pinch out laterally over a few 10's of centimetres. These sandstone lenses are not gradational to the conglomerate lenses and are visible separating conglomerate lenses, as well as within conglomerate lenses. Clay-rich, fine-grained sand may be present near the top of some conglomerate lenses.

Chemical Sedimentary rocks

Chert

Chert in the southern Finlayson Lake area is white to grey or greyish-black in colour. Purer chert tends to be white, while the darker coloration is due to variable amounts of chlorite, magnetite, or carbon mixed with the chert.

Chert layers are thin, but are laterally continuous over outcrops with lengths on the scale of a few meters. Layer thicknesses range from approximately 0.1 cm to a 4cm. However, some of the thicker layers may be composed of several thin layers. In some cases this is suggested by thin laminations within the thicker cherts, which may be apparent through slight colour changes or micro-laminations of pyrite. Bedding contacts are dominantly sharp from one bed to the next, but occasionally a gradation from chert to magnetite over a few millimetres in thickness is present.

Magnetite

Magnetite is commonly intimately associated with chert in the southern Finlayson Lake area (Fig. 2.07). A gradation exists between chert-rich layers and magnetite-rich layers. The chert-rich magnetite layers appear grey to black, with the most magnetite-rich layers being black in colour.

Magnetite layers range in thickness from < 0.1 cm to approximately 1 cm. Thin laminations, 0.1 cm or less in thickness, of nearly pure magnetite and cherty magnetite are often visible in the thicker magnetite layers. These layers are thin, but usually laterally continuous within an outcrop. Both upper and lower contacts are commonly sharp. However, occasionally a gradation from chert to magnetite occurs over a thickness of a few millimetres.

Sulphides

Sulphide facies iron formation is not very common in the southern Finlayson Lake area. Thin laminations, 0.2 cm or less, of pyrite or pyrrhotite are occasionally found within chert units. These layers are commonly continuous throughout the chert exposure.

Sulphides are also present within the slates of the area. These layers range in thickness from a few millimetres to a few centimetres and may be continuous or thin out laterally over 1-2 meters along strike. The sulphides are usually associated with graphite-rich slates and may be interbedded with them or disseminated throughout the graphitic slate layers.

2.4 Northern Finlayson Lake Sedimentary rocks

Sedimentary lithofacies in the Northern Finlayson Lake area are similar to those found in the Southern Finlayson Lake area. However sedimentary rocks here tend to have a finer-grained nature. The sedimentary facies found in Northern Finlayson Lake consist of fine-grained sandstone and siltstone couplets, medium- to coarse-grained sandstones, conglomerate, as well as chemical precipitates. As in the Southern Finlayson Lake area, sedimentary rock sequences conformably overlie mafic flows, or alternatively, in some areas directly overlie a thick unit of intermediate agglomerate

Clastic Sedimentary Rocks

Intermediate Agglomerate

This facies consists of thick beds of intermediate tuff breccia. Clasts of intermediate volcanic composition are found within a medium-grained, more chloritic, ash matrix, which is dark green in colour. Agglomerate is monomictic and clast supported and is composed of subrounded clasts. In the lower parts of the section cobble sized clasts are dominant, up to 30 x 8 cm in size, with an average of approximately 15 x 4 cm. Higher in the section the clasts are in the pebble size range, averaging 2-3 centimetres in diameter. Clasts are highly strained, and are stretched more in the vertical direction than the horizontal.

The agglomerate occurs in beds that are up to 12 meters in thickness, with the thickest beds present lower in the section. Beds in the lower section are usually between 400 cm and 1200 cm thick, while higher up section the beds are thinner, ranging between 100 and 400 centimetres, with an average of approximately 200 cm.

Some of the agglomerate beds show some grading, from tightly clast-supported cobbles at the base to loosely clast-supported pebbles near the top of the bed. Although there is some overall grading of the clasts, they are quite poorly sorted. The matrix of the agglomerate tends to change near the tops of the beds as well. Moving up within an agglomerate bed the matrix commonly becomes less chloritic and finer-grained, containing increasing amounts of black, graphitic slate, which may or may not contain sulphides. This change takes place within the uppermost 1 - 2 m of the agglomerate bed. Most agglomerate beds terminate with a few centimetres, to a few tens of centimetres, of thinly laminated sulphide facies iron formation and black graphitic slate. The graphitic slate and sulphide content of the agglomerates is also noticeable in the weathering pattern of the beds. Near the tops of beds, the weathered surface becomes increasingly more pitted, taking on a "rotted" texture.

In thin section, the more mafic matrix and felsic nature of the clasts is visible. The clasts are fine-grained, and are composed dominantly of quartz, feldspar, and carbonate, with lesser amounts of epidote, chlorite, and muscovite. The matrix is also fine grained, but is dominated by chlorite and muscovite, with meagre quartz and less epidote. The matrix contains a much higher proportion of opaque minerals than do the clasts.

Conglomerate

Polymictic conglomerate is found in the northern part of Finlayson Lake at outcrop #18. Conglomerate is clast supported and contains mainly felsic volcanic and a lesser amount of mafic volcanic clasts. Rare banded chert-magnetite iron formation and chert clasts are also present. Clasts are subrounded to subangular and poorly sorted. The matrix is composed of fine- to medium-grained sandstone.

Conglomerates appear massive, with a maximum thickness of approximately 12 m. However, within the massive units are zones up to approximately 100 cm thick, which contain more BIF clasts than the majority of the conglomerate. This may be suggestive of bedding or lensing which is not apparent through distinct bedding contacts, due to the homogeneous nature of the conglomerate.

Medium- to Coarse-Grained Pebbly Sandstone

This facies consist of sandstone beds with a medium- to coarse-grained base, and a finer-grained sandstone or siltstone top. The gradation to the fine-grained top is commonly quite sharp and takes place over a few millimetres to a couple of centimetres. The medium- to coarse-grained base of this facies usually contains a variable amount of granules, a few millimetres in diameter, to small pebbles, up to approximately 1 centimetre in diameter (Fig. 2.08). The granules and pebbles, which are dominantly quartz and granitic rock fragments, are generally subangular to subrounded . Coarse-tail grading is commonly visible, as the granules and pebbles decrease in size upward and die out within the upper part of the coarse-grained base of the bed. The coarse-grained base may also contain rare shale rip-up clasts up to a few centimetres in length. The finer-grained tops of the beds often contain well developed parallel laminations between fine-grained sandstone and siltstone. A well developed shale cap is rare in the medium- to coarse-grained pebbly sandstone facies.

Bed thicknesses range from approximately 10 - 30 cm, with an average between 15 - 20 cm. The medium- to coarse-grained basal division usually comprises approximately 60 - 80% of the total bed thickness. The thinnest beds have a lower coarse-grained to fine-grained ratio in the range of 30/70. Both upper and lower contacts of beds in this facies are sharp.

Fine-grained sandstone and siltstone couplets

This facies consists of thin units of paired fine-grained sandstone and siltstone layers (Fig. 2.09). The base is typically graded from a fine-grained

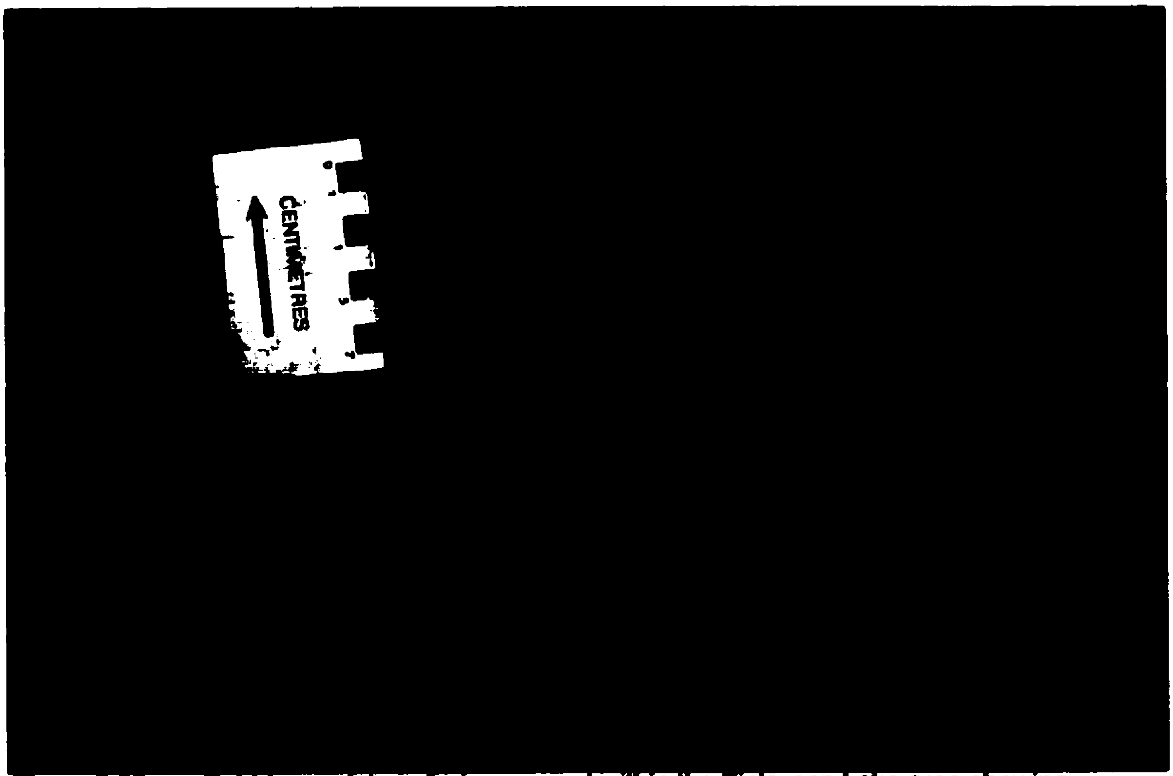


Figure 2.08 Pebbly-sandstone facies present within the Finlayson Lake area showing decimeter scale bedding and sub-centimeter scale pebbles.

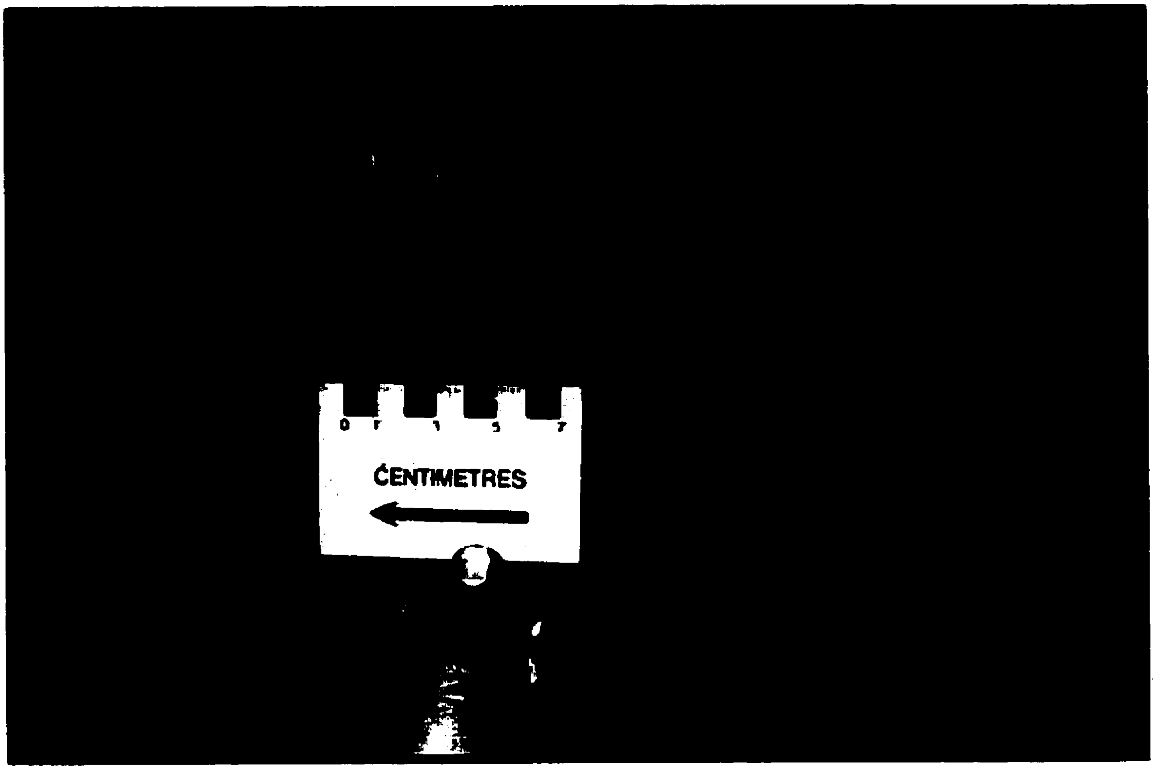


Figure 2.09 Fine-grained sandstone and shale facies present within the Finlayson Lake area showing contact between fine-grained sandstone and shale bed..

sandstone to a very fine-grained sandstone or siltstone. This commonly undergoes a subtle or fairly abrupt gradation to a silty, clay-rich, top. The basal section contains well developed normal grading, while the upper section of the bed is usually parallel laminated coarse silt and clay. Alternatively there may be parallel laminations throughout the entire bed, although these are commonly better developed near the top of the beds. The uppermost portion of the bed usually consists of a thin clay drape, up to 1 cm in thickness. Bed thickness ranges from approximately 1 cm to 10 cm, but most commonly is between 2 cm and 5 cm. This facies is also gradational with thicker beds up to approximately 20 cm thick. The thicker beds commonly have a medium-grained sandstone base, graded to a silty or silty-mud top. The lower medium-grained sandstone division commonly comprises 50-70% of the total bed thickness.

Bedding contacts are dominantly sharp between beds and are also fairly continuous over outcrops of a few meters in length. The lower contact may occasionally show loading of the coarser-grained base of the upper bed, a few millimetres to a centimetre, into the silty clay top of the underlying bed. Rarely, thin shale rip-up clasts, up to a few cm x a few mm in size, are observed within the coarser-grained basal section of beds.

Chemical Sedimentary Rocks

The chemical sedimentary rocks in the Finlayson Lake area are commonly intimately related to one another making it difficult to discuss one facies without involving others in the description. Chert, magnetite, sulphide, and graphitic slate are all present in the area and are often interlaminated on a mm scale.

Chert

Chert in the Northern Finlayson Lake area varies in colour from white to a dark grey-black. The variation from white to black is due to an increase in the amount of clastic material and graphite incorporated into the chert. The colour variation may be gradual, showing a gradation from white to black within a single layer, or abrupt, with sharp contacts defining the graphite-rich and more pure chert layers.

Units that are dominated by chert range in thickness from a few centimetres up to approximately 200 cm. These chert units may show millimetre-scale laminations to centimetre-scale layers internally. Layering may be defined in several ways, such as a slight colour variation, or by thin interbeds of another facies of chemical sediment. Commonly, 5 -15 cm thick layers are visible within the chert units, from a pure white chert base to a dark grey graphite-rich top. Sulphide-rich zones may be found disseminated throughout the chert or as thin lenses and millimetre-scale laminations within a chert dominated unit. Contacts are commonly sharp and laterally persistent. Rarely, some degree of soft sediment deformation is expressed by a wavy, irregular basal contact with the underlying unit.

Sulphides

Sulphides in the northern Finlayson Lake area consist dominantly of pyrite and lesser pyrrhotite. Sulphides may be found in units ranging from <1.0 centimetre to approximately 100 cm in thickness. Thicker units are in some cases continuous over outcrop distances of a few meters, or will thin laterally giving a lens shape to the overall unit.

As with the other chemical sedimentary rocks, sulphides may dominate a unit or be found as thin laminations within units dominated by another facies. Millimetre-scale laminations of pyrite are found within chert or graphitic slate dominated units and can be useful in determining the layering within these units. The thicker sulphide dominated units often have thin millimetre- to centimetre-scale interbeds and lenses of chert or graphitic slate. Sulphides may appear massive where recrystallization has destroyed original layering. When not obliterated by recrystallization, thin laminations are often visible within the thicker sulphide units. The laminations range from <0.1 cm to 0.3 cm in thickness, and appear laterally continuous, with sharp upper and lower contacts.

Sulphide may also be found as round to oval masses within graphitic slate. These spheroids range in diameter from <1 cm to approximately 4 cm. Internally they consist of concentric, millimetre-scale laminations.

Graphitic Slate

Graphitic slate is common in the northern Finlayson Lake area. It is present as thin interbeds between mafic volcanic flows and pyroclastic rocks, or within the main sedimentary band. There appears to be a gradation from graphitic argillite to chert dominated layers. Units may be thin, only a few centimetres in thickness, or up to tens of centimetres thick. Graphitic slate is dark grey to black in colour, and often has a "greasy" appearance where it is particularly rich in graphite. Slate is commonly very schistose, and may have a "rotted" appearance when weathered.

Graphitic slate is often found associated with sulphide facies iron formation, as thin laminations within the sulphides or chert, or it can be the dominant member of the unit. Graphitic argillite dominated units appear massive or show thin millimetre-scale laminations. Laminations are expressed by colour variation, due to variation in clastic, graphite or chert content. Layering is commonly continuous, with sharp upper and lower contacts.

Carbonate

Carbonate is present in the northern Finlayson Lake area within a possible hydrothermal stockwork system on the point near the location of sample 54 (Fig. 1.4). The carbonate forms a vein system within the volcanic rocks at the base of the sedimentary sequence. An intricate system of anatomising veins, which are now filled with iron carbonate may be evidence of a hydrothermal stockwork system in the area (Fig. 2.10). The carbonate veins are up to 2-3 cm in thickness and branch upward through the volcanic rocks were they are capped by a chert unit.

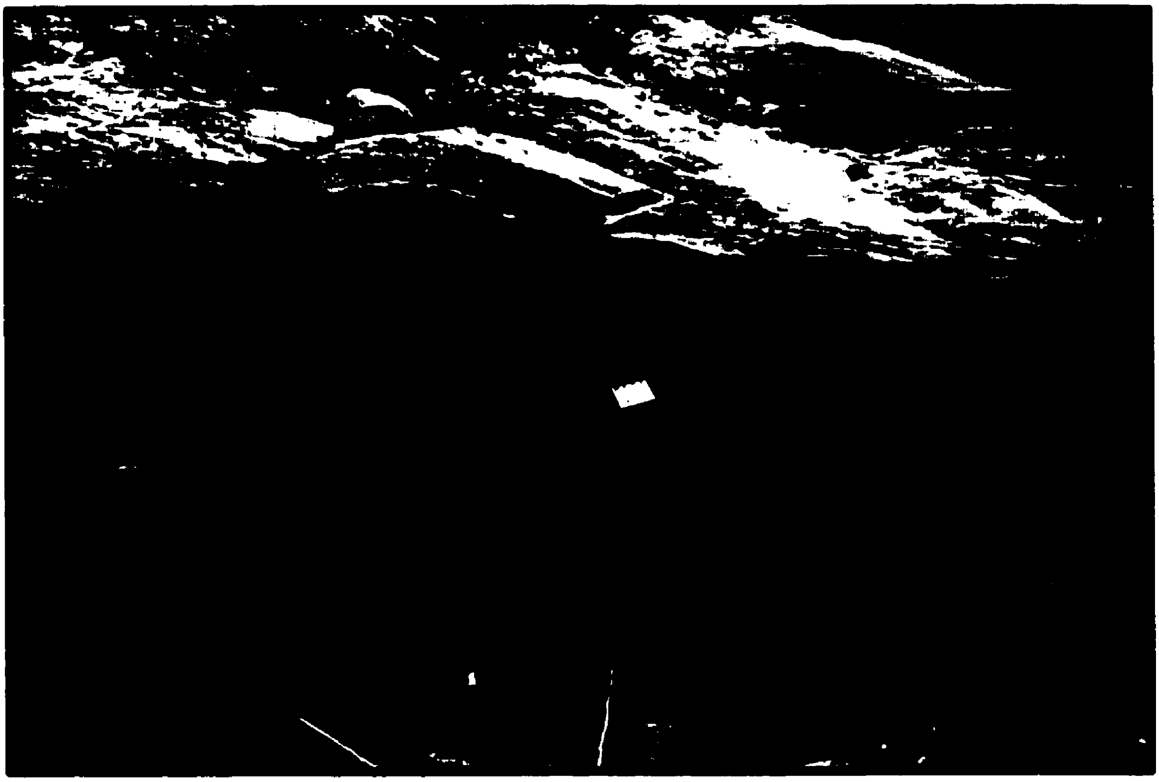


Figure 2.10 Hydrothermal stockwork system present within the Finlayson Lake area. Photo shows anastomosing carbonate viens at the base of the photo capped by chert. Scale card is 9cm wide.

CHAPTER 3

LITHOFACIES DESCRIPTIONS OF THE LUMBY LAKE GREENSTONE BELT

Many of the clastic sedimentary lithofacies in the Lumby Lake area are similar to those which form the main sequence of sedimentary rocks in the Finlayson Lake area, although the rocks tend to be finer-grained. Chemical precipitates are more common, and form thicker sequences in the Lumby Lake area, becoming the dominant lithofacies along a west to east transect through the belt. For the purposes of this thesis the Lumby Lake Sedimentary Belt will be divided into 4 separate areas and the sedimentary lithofacies found in each will be described. Three of the areas, 1) Norway Lake, 2) Pinecone and Cryderman Lakes area, and 4) Keewatin and Hematite Lakes area, contain a continuous band of sedimentary rocks which form the Upper Sedimentary Belt. These also define a west to east transect through the belt. The Hook Lake area contains sedimentary rocks which are continuous with felsic volcanic rocks to the east, and forms the Lower Sedimentary Belt.

3.1 Norway Lake Area

The Norway Lake area consists of the western-most section of the Upper Sedimentary Belt (Fig.3.01), and contains the coarsest-grained clastic sedimentary rocks. Clastic rocks include conglomerates and fine- to medium-

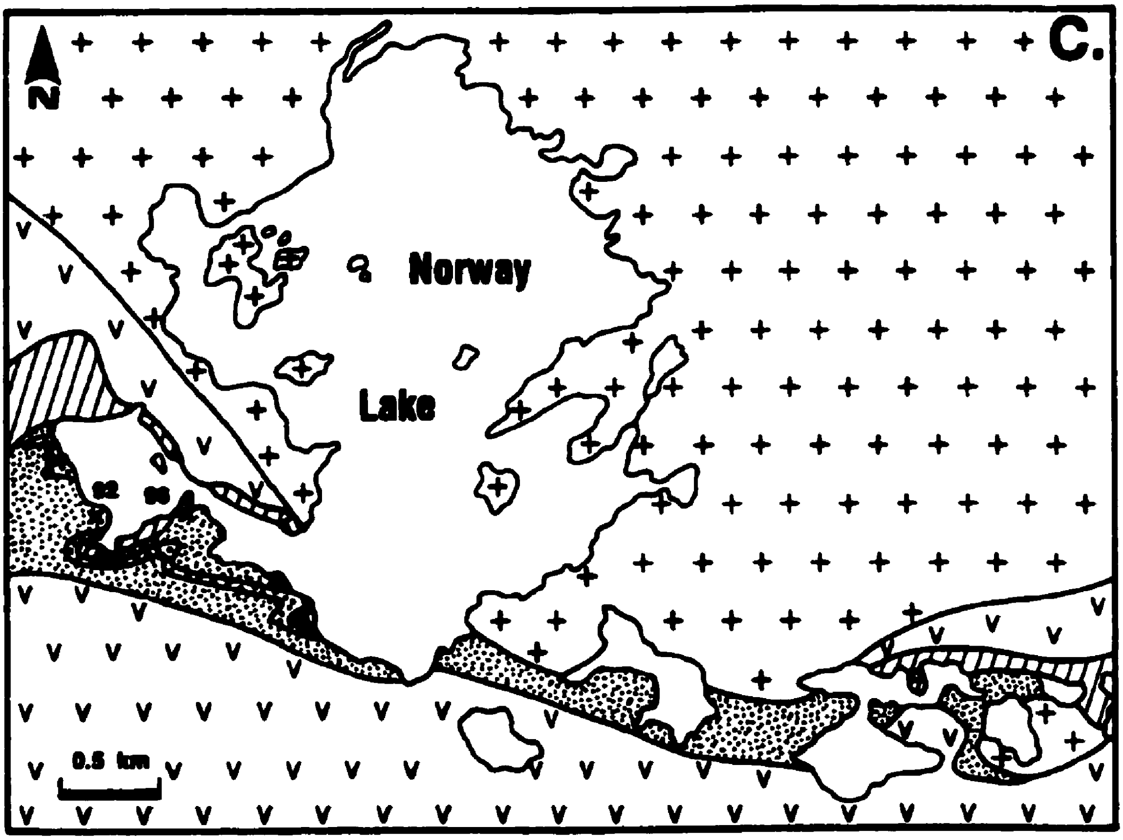


Figure 3.01 General geology of the Norway and Pinecone-Cryderman Lakes area. Refer to Figure 1.04 for geological legend. Refer to Figure 1.01 for location within the Lumby Lake Greenstone belt. Numbers refer to lithogeochemical sample locations.

grained sandstones. Chemical sedimentary rock units are not as variable in the Norway Lake area as they are further to the east. Chert and magnetite iron formation are present, and marble is more abundant in this area than to the east. The higher metamorphic grade in the area, due to the intrusion of the Norway Lake Pluton (Jackson, 1985), makes it difficult to identify many primary structures that may be used as younging indicators.

Clastic Sedimentary Rocks

Conglomerate

Conglomerate in the Norway Lake area consists of polymictic conglomerate in a fine-grained sandstone matrix. The matrix is chlorite-rich, and contains many crystals of metamorphic garnet within it. The conglomerate is clast supported and composed of pebble to cobble size clasts (Fig. 3.02). Clasts are subrounded, and highly stretched in a plane parallel to the dominant schistosity in the area. Clast size ranges from <1 cm in diameter, up to approximately 45 x 7 cm, with an average clast size of approximately 2 x 5 cm. Felsic volcanic, and mafic volcanic rock fragments are the dominant clast lithologies, with rare gabbro, chert, and iron formation clasts also present.

Norway Lake conglomerate is poorly sorted and non-graded, with respect to the clasts. Conglomerate appears massive with no indication of bedding contacts or lenses within the thicker conglomerate units. It is possible that

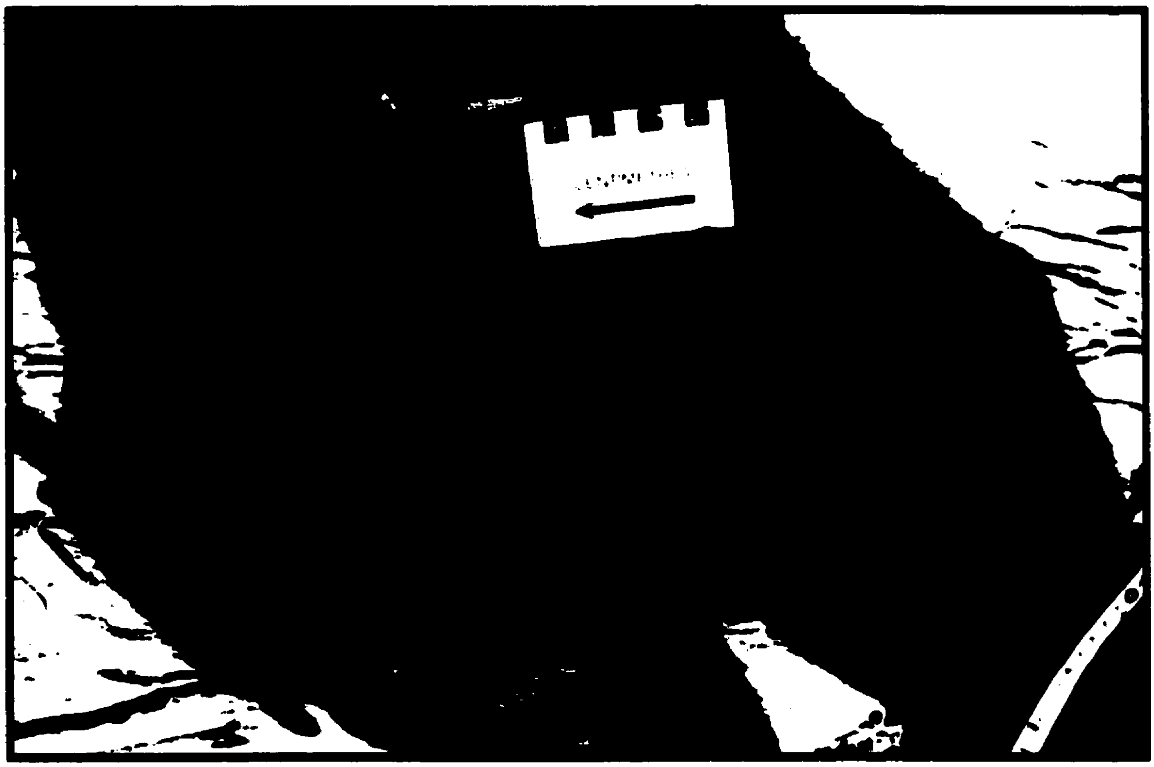


Figure 3.02 Photograph of the conglomerate facies present in the Norway Lake area showing sub-rounded, highly deformed clasts.

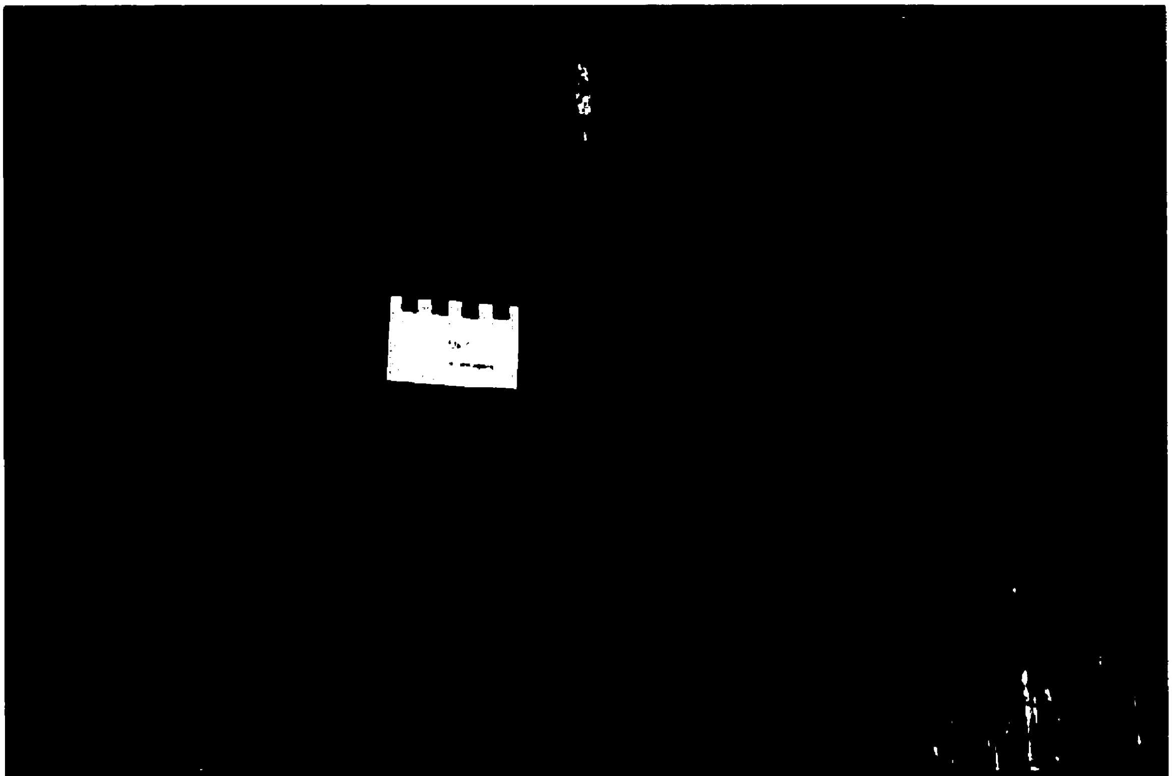


Figure 3.03 Photograph of the fine-grained sandstone and slate facies present in the Norway lake area showing sharp bedding contacts and centimeter scale bedding.

bedding contacts are not visible due to the poor and limited outcrop exposure, as well as the homogenous nature of the conglomerates. A slight decrease in average clast size, in a southward direction, was noticeable over an outcrop approximately 10 meters wide. This may be a true gradation within a single bed, or simply changes reflecting different beds, with contacts which are not easily visible. There does appear to be some subtle variation from one outcrop of conglomerate to another. Some outcrops appear to have a slightly finer-grained more clay-rich matrix than others, and a higher matrix to clast ratio.

Fine- to Medium-Grained Sandstone

Fine- to medium-grained sandstone is common in the Norway Lake area, and is similar to sandstones in the Finlayson Lake area. Each bed commonly consists of a medium- or coarse-grained base, coupled with a finer-grained siltstone to slatey top (Fig. 3.03). Sandstones are poorly sorted and contain a substantial percentage of matrix material. The clay-rich top of a bed can often be distinguished in the field by an increase in number of metamorphic garnet porphyroblasts visible. Primary internal structures are rare in the sandstones, possibly due to the locally higher metamorphic grade.

Bedding thickness ranges from approximately 5 cm to 20 cm, averaging in the 10 cm range. In some beds grading is visible from a fine- or medium-grained base to a clay or silty clay top. Beds are commonly laterally persistent over outcrop lengths of a few meters, with little indication of thinning or thickening. Both upper and lower bed contacts are sharp, and occasionally, loading of the medium-grained base of one bed, into the clay rich top of the underlying bed is visible.

Chemical Sedimentary Rocks

Marble

Marble is fairly common in the Garnet Bay area of Norway Lake. Carbonate units occur as isolated pockets, up to 20 meters thick, along the south shore of Garnet Bay. Carbonate appears grey to brown on the weathered surface, and white to bluish-grey on a fresh surface.

The layering and intricate folding within the carbonate is visible through the interlayering of carbonate and positively weathered siliceous layers (Fig. 3.04). Carbonate layers from 0.2 cm to approximately 4 cm in thickness, are interlaminated with siliceous layers ranging from approximately 0.1 cm to 1 cm thick. The contacts between the layers is often sharp. These two components are arranged into units which are either dominated by carbonate layers, or dominated by siliceous layers. Carbonate dominated units range from approximately 5 to 10 cm in thickness, while siliceous units are usually slightly thinner, ranging from approximately 2 to 6 cm thick. The transition from one unit to another appears gradational in some cases. A carbonate dominated unit, with a few thin siliceous layers, will show a progressive increase in the frequency and

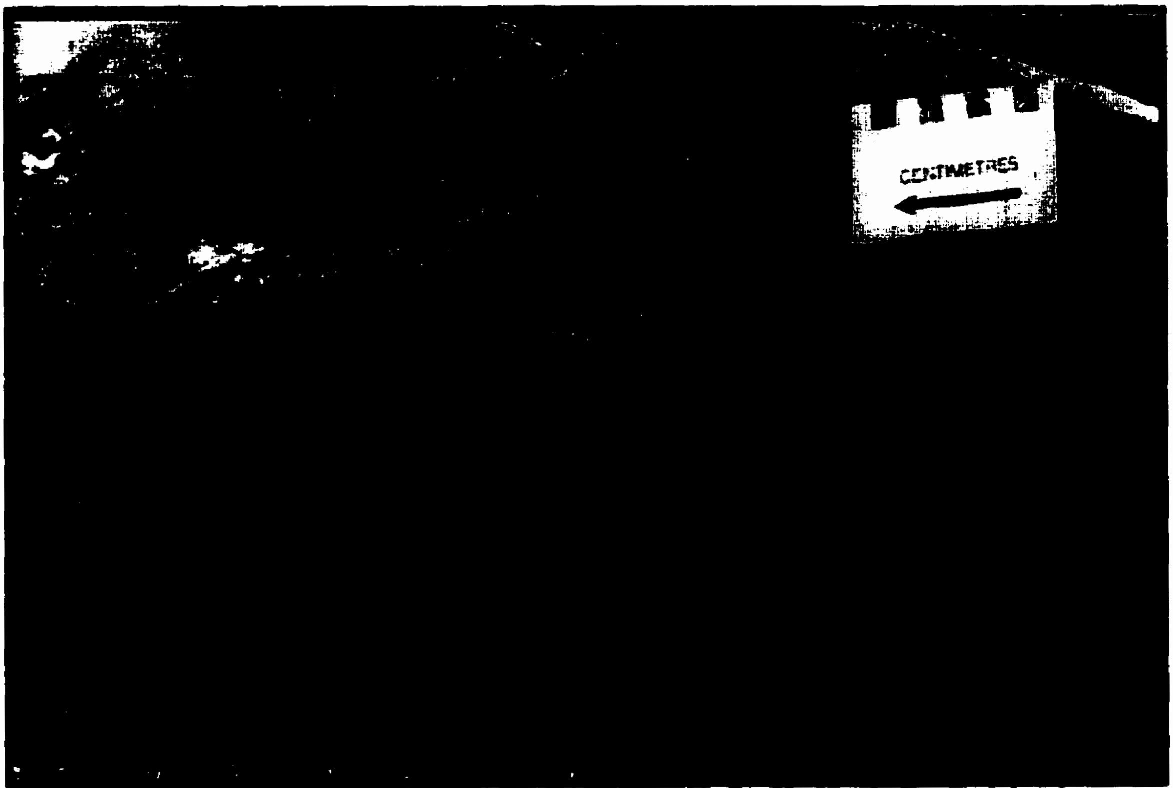


Figure 3.04 Photograph of the carbonate facies found in the Norway Lake area showing folding of sub-centimeter scale siliceous and recessive weathering carbonate layers.

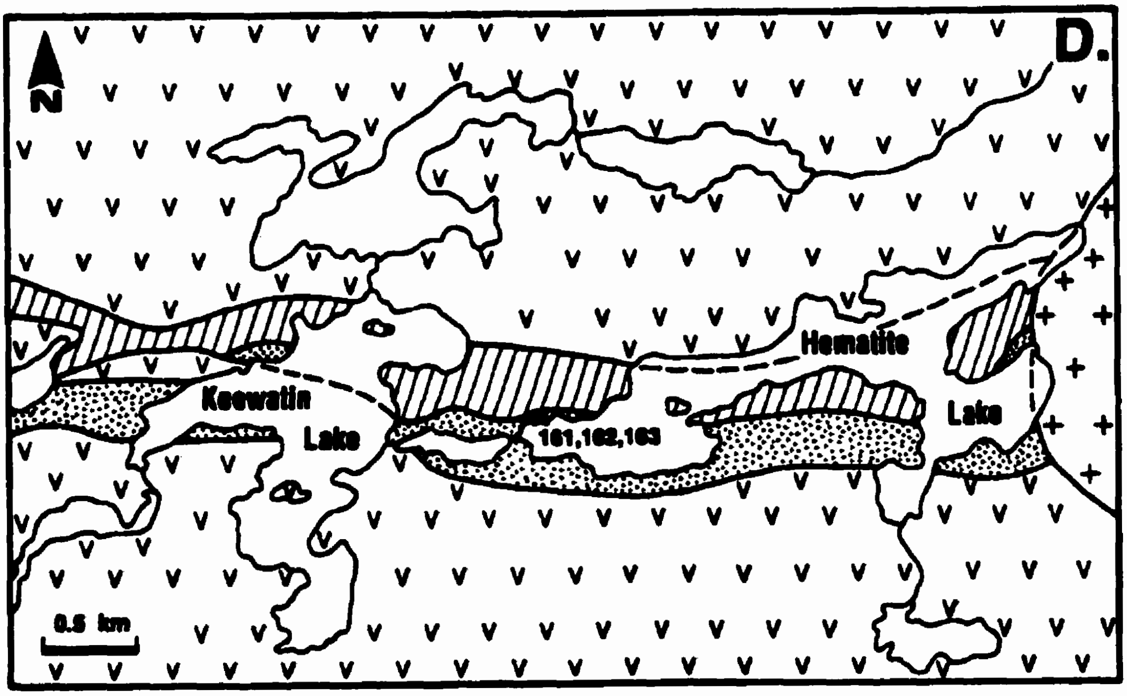


Figure 3.05 General geology of the Keewatin and Hematite lakes area. Numbers refer to samples lithogeochemical samples taken. Refer to Figure 1.01 for location within the Lumby Lake Greenstone Belt and 1.04 for geological legend.

thickness of siliceous layers, until the unit becomes dominated by siliceous layers with a few thin carbonate layers.

Carbonate is coarse grained, and highly recrystallized, with the diameter of carbonate grains varying from 0.1 to 0.3 cm. Recrystallization of the carbonate disrupts some of the thinnest silicate layers (0.2 cm or less) in the carbonate dominated units. All carbonate is calcite, with very little variation in the main cation concentrations (Fig. 3.06). This carbonate, with a low iron composition, is different from other carbonates within the Lumby Lake Greenstone Belt. Most other carbonate in the belt has either an ankerite or siderite composition (determined from SEM-EDS).

Chert

Chert in the Norway Lake area is intimately associated with magnetite and fine-grained, matrix-rich sandstone. Chert is interbedded with either magnetite-rich chert, or sandstone layers. Chert appears white to grey in colour depending on the amount of clastic material contained within it.

In Garnet Bay, Norway Lake, an outcrop of chert with lenticular bedding is present. This consists of lenses of chert approximately 5 to 15 cm thick, by approximately 50 to 100 cm in length. Chert here is recrystallized and fairly coarse grained, with a sugary appearance on a fresh surface.

Chert at Garnet Bay is also interbedded with magnetite-rich chert. The purer cherts are white in colour, and the magnetite-rich chert is dark grey to

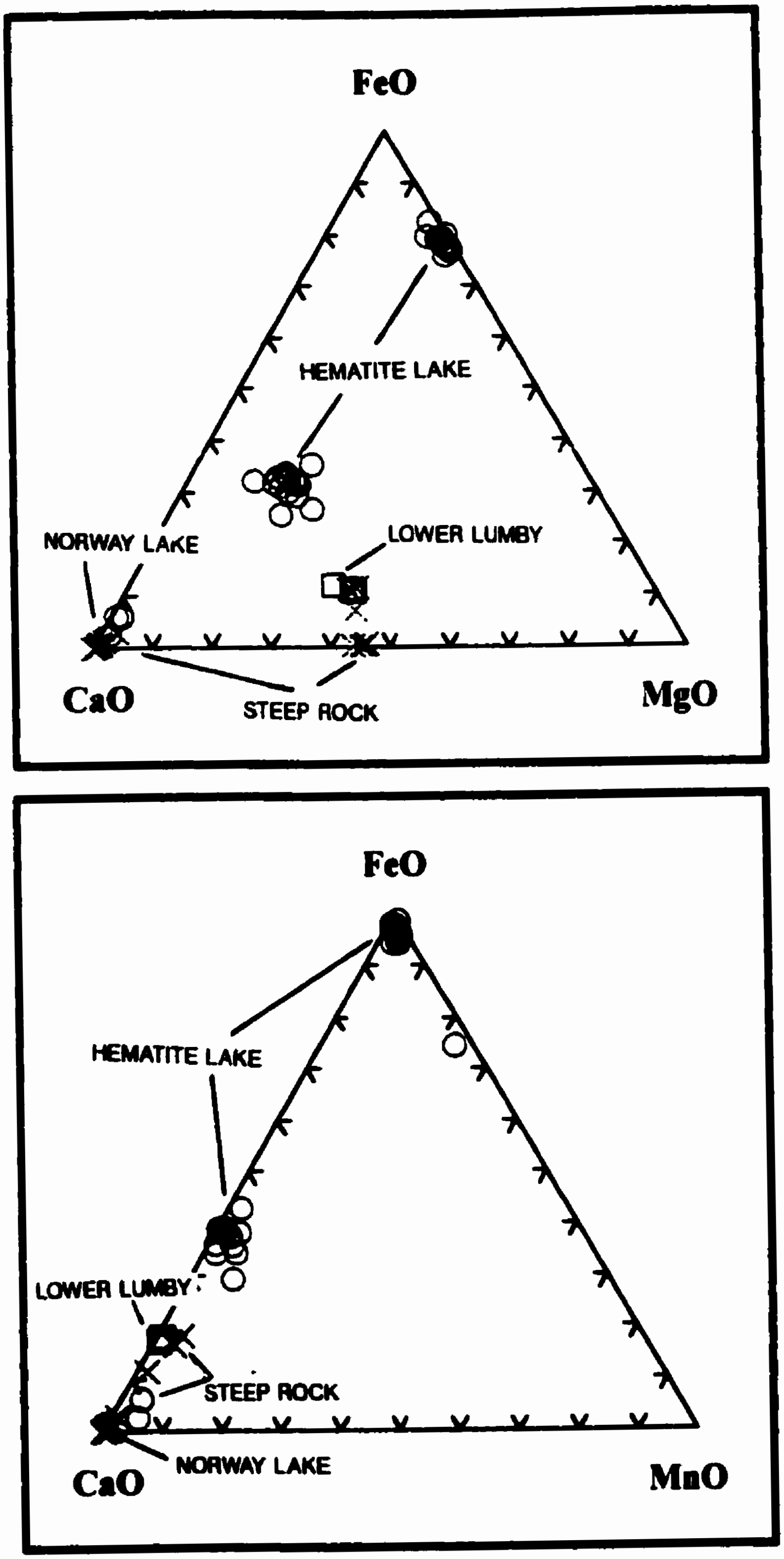


Figure 3.06 Carbonate compositions determined by SEM-EDS of selected samples taken from within the Steep Rock, Finlayson and Lumby Lake areas. Squares-Lower Lumby Lake, Open circles- Hematite Lake area, X-Steep Rock Lake, Closed circles-Norway Lake.

brown in colour on the weathered surface. Chert layers are between 0.5 and 3 cm in thickness and are laterally persistent. Magnetite-rich layers are 0.2 to 2 cm in thickness, and are, on average, thinner than the chert layers. The contacts between chert and magnetite-rich chert are sharp in some cases, and are gradational in others.

Magnetite

Magnetite is always associated with chert layers in the Norway Lake area. Layers of magnetite-rich chert are interbedded with chert layers. The magnetite-rich layers often appear dark grey to brownish-red on the weathered surface. Magnetite-rich layers range in thickness from approximately 0.2 cm to 2 cm, with an average thickness of 1 cm. Contacts with purer chert layers may be sharp or gradational, and the layers are laterally continuous.

3.2 Pinecone and Cryderman Lakes Area

The Pinecone and Cryderman Lakes area is in the central portion of the Upper Sedimentary Belt, east of the Norway Lake area and west of the Keewatin and Hematite Lake area (Fig. 3.01). Clastic sedimentary rocks here are less abundant and finer-grained, with only fine-grained sandstones present. The chemical precipitates are more varied, and include carbonate, chert, magnetite, and sulphides, with associated graphitic slates.

Clastic Sedimentary Rocks

Fine-Grained Sandstones

This facies consists dominantly of fine-grained sandstone and shale couplets. Each couplet has a fine-grained sandstone base, which is dark grey in colour, and a fine siltstone to clay-rich top. The ratio of fine-grained sandstone base to clay-rich top, ranges from approximately 50/50 to 70/30. The fine-grained top portion often contains a high concentration of garnet porphyroblasts, each up to a few millimetres in diameter. Fine-grained sandstones are poorly sorted. The sandstones consist of fine-grained quartz in a matrix of chlorite, muscovite, biotite and hornblende. The fine matrix composes up to 40 % of the rock, making these sandstones very argillaceous.

This facies is thinly bedded, with bed thickness ranging from approximately 1 cm to 10 cm, with an average of 3-5 cm. The fine-grained sandstone base often has an abrupt gradation to the slate top. The base of some beds is parallel laminated fine-grained sandstone and siltstone. Both upper and lower bed contacts are sharp, and bedding planes are laterally persistent over outcrops a few meters in length.

In some cases the very fine-grained, clay-rich top of the beds may be absent, and beds are recognised by a fine-grained sandstone base which grades continuously to a very fine-grained sandstone or siltstone top.

Chemical Sedimentary Rocks

Carbonate

Carbonate occurs in small outcrops at the west end of Pinecone Lake, and is similar to carbonate on Garnet Bay, Norway Lake. The carbonate is coarse-grained and highly recrystallized calcite. The white to bluish-white carbonate contains thin, black cherty mudstone layers running through it. The extreme folding, evident in the silicate layers, has disturbed any original bedding in the carbonate.

Chert

Chert ranges in colour from white to dark grey, depending on the amount of clastic material or magnetite in it. On the northern shore of Pinecone Lake the chert is interbedded with magnetite-rich layers. Chert units, ranging in thickness from 0.2 to 3 cm, alternate with magnetite-rich units of similar thickness. The chert units often show thin laminations, 0.1 to 0.5 cm thick, of purer white chert and slightly darker grey chert. The darker layers contain scattered grains of clastic material, giving them a darker colour. Contacts between the layers are in most cases sharp, however, rarely, they grade into one another. Some magnetite-rich layers have a sharp lower contact and grade from magnetite into magnetite-rich chert and finally into chert. Chert units are laterally continuous over outcrops up to approximately 5 meters in strike length.

Magnetite

Units of magnetite are interbedded with chert on the north shore of Pinecone Lake and on Cryderman Lake. A colour variation from dark black to grey exists, depending on the chert content within the magnetite bands, with the darkest layers being the most magnetic. Magnetite units range in thickness from <0.5 cm to approximately 2 cm, and commonly have sharp upper and lower contacts. The thickest magnetite units commonly contain several laminae, up to a few millimetres in thickness. These are composed of nearly pure magnetite laminae and chert-rich magnetite laminae. The laminae have either sharp upper and lower contacts or are in some cases gradational into the chert-rich laminae. The magnetite rich bands are laterally continuous over several meters.

Graphitic Slate

Dark black coloured, graphitic slate occurs in units up to a few meters thick, and is associated with sulphides on Pinecone and Cryderman Lakes. Graphitic slate units are often highly schistose, and slates have a greasy appearance on fresh surfaces. The graphitic slate units consist of interlaminated chert-rich and clastic-rich layers, both of which contain graphite up to approximately 15 % by volume, giving a dark black colour. They are thinly laminated, with laminations ranging from <0.1 cm to 0.3 cm thick. The laminations do not appear to be laterally persistent for more than approximately 5-10 cm, which is likely the result of the cleavage disrupting original layering.

Sulphides

Both pyrite and pyrrhotite are found in the Pinecone-Cryderman Lakes area, and are associated with the graphitic slates. Sulphide may be found as massive layers, up to 15 cm thick, or disseminated throughout the graphitic slate. The massive sulphide does not appear to be layered internally, possibly due to later recrystallization, which may have disturbed any original layering.

Pyrite nodules are found within graphitic slate units at Cryderman Lake. The nodules are not scattered randomly throughout the slate, but are concentrated within bands up to 25 cm thick, which are parallel to bedding planes in the slate. The nodules range in diameter from <1.0 cm to approximately 3 cm, with an average diameter from 1-2 cm. Internally the nodules are composed of several concentric laminae. The laminae average 0.1-0.2 cm in thickness and are visible on the broken surface of most nodules.

3.3 Keewatin and Hematite Lakes Area

The Keewatin and Hematite Lakes area is the eastern-most area within which the Upper Sedimentary Belt is developed. Clastic sedimentary rocks become increasingly finer-grained and consist of fine-grained sandstones and mudstones. Chemical precipitates dominate in the area, and many different facies are present (Fig. 3.05). These include chert, magnetite, carbonate, sulphides, and interbedded graphitic slates. Many of the chemical units are interbedded with one another.

Clastic Sedimentary Rocks

Fine-Grained Sandstone

Fine-grained sandstone and slate are common along the southern shores of Keewatin and Hematite Lakes. This facies consists of fine-grained sandstone to siltstone and slate couplets. Some beds are lacking the fine-grained sandstone division, and consist only of a siltstone and slate divisions.

Bed thickness ranges between 1 cm and 6 cm, and only the thicker beds contain fine-grained sandstone. Where the unit consists of only siltstone and slate the average thickness is only a couple of centimetres. Each unit begins with a fine-grained sandstone or siltstone division, which has a sharp lower contact. This basal division may show a gradual gradation, or a rapid gradation, into an overlying slate. The fine-grained sandstone or siltstone to slate ratio is commonly between 40/60 and 60/40. Occasionally parallel laminations of fine-grained sandstone and siltstone are present in the lower portion of the beds.

Contacts between each bed are sharp and the layers are laterally continuous over outcrops a few meters in length.

Green Mudstone

Very fine-grained siltstone and mudstone is found in the Keewatin-Hematite Lakes area. Mudstone commonly has a greenish colour, which is due to a high chlorite content. Chlorite, muscovite, quartz and feldspar are the dominant minerals in these layers.

Green mudstone forms layers <0.1 cm up to several centimetres thick. The layers are often interbedded with chert, and chert-rich mudstone, as well as thin, millimetre-scale, sulphide layers. Contacts with sulphide and chert layers are commonly sharp (Fig. 3.07). Grading is visible within some layers, from a siltstone base to a fine-grained chlorite-rich upper portion. Grading is gradual and takes place over the entire thickness of the layer.

Figure 3.07 SEM photomicrograph of the Green mudstone facies found in the Keewatin-Hematite lakes area.

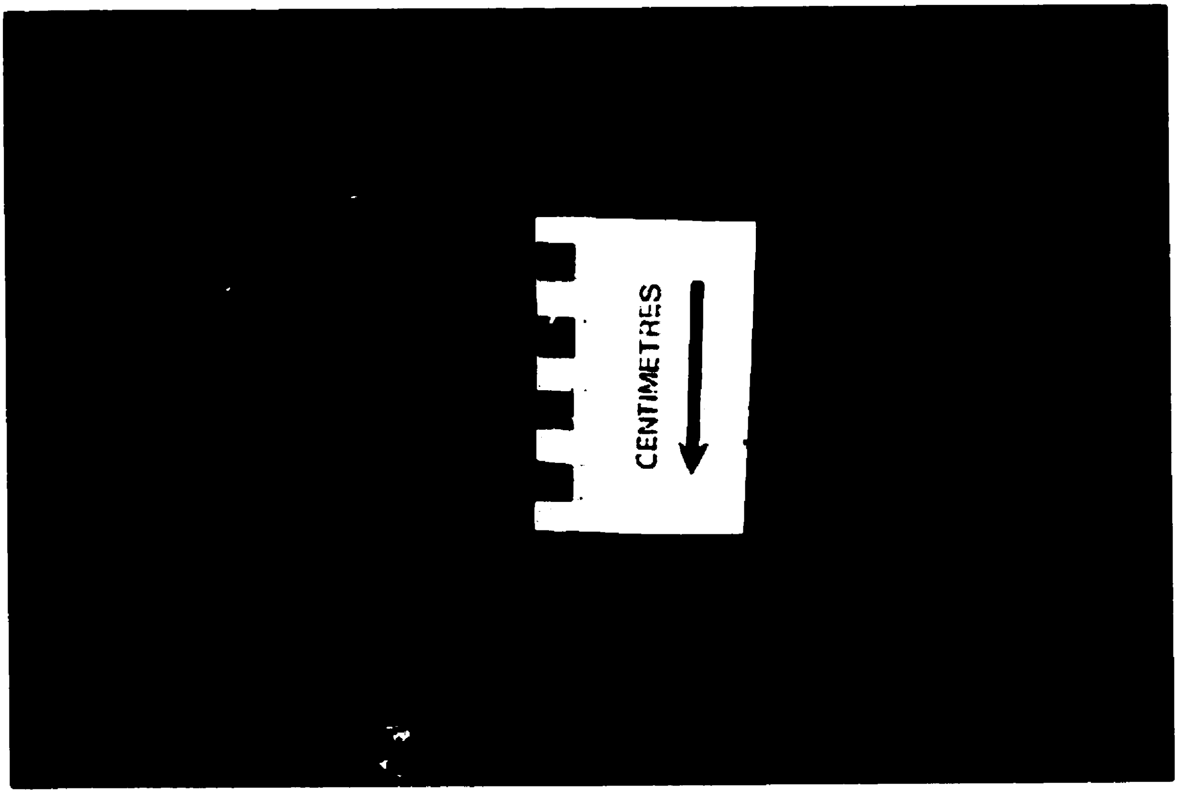


Figure 3.08 Photograph of the chert-magnetite iron formation present within the Keewatin-Hematite lakes area. Thinly bedded magnetite dominated (dark) and chert dominated (white) layers are visible.

Chemical Sedimentary Rocks

Chert

Chert deposition was widespread in the eastern portion of the Lumby Lake Greenstone belt. Chert can be found interbedded and mixed with all other chemical precipitates, including carbonate, sulphides and magnetite, as well as clastic mudstones.

Banded chert and magnetite-rich iron formation are common on northern Keewatin Lake. These consist of units containing several layers, which are either chert dominated or magnetite dominated (Fig. 3.08). Chert dominated units range in thickness from only a few millimetres, up to approximately 5 cm, with an average thickness between 0.5 - 1.0 cm. The thickest units are thinly laminated, with laminae ranging in thickness from <0.1 cm to 1.0 cm. Contacts between the units are commonly sharp, however some appear gradational, grading from magnetite dominated to chert dominated. The chert units that are gradational from a magnetite base, have sharp upper contacts with the next magnetite dominated unit. The chert layers, whether thick or thin, appear laterally continuous over outcrops several meters in length.

Chert is also found interbedded with carbonate as chert dominated layers and carbonate dominated layers. The chert ranges in colour from white to brownish-red on weathered surfaces, depending on the amount of iron carbonate present. Chert layers range in thickness from 0.1 cm to approximately 2 cm, with the thicker units containing several laminations of chert from <0.1 cm to 1 cm thick. Even the chert dominated units contain micro-laminations of carbonate within them. Millimetre-scale chert layers often show a gradation from carbonate layers (Fig.3.09A). The chert grades from a carbonate layer, becoming purer upwards, and has a sharp upper contact with the next carbonate dominated layer. Contacts are often wavy, suggesting some degree of soft sediment deformation.

Thick units of chert containing thin laminations of sulphide are also present. Very pure white chert is thinly laminated, from <0.1 cm to approximately 0.5 cm thick. The chert layers show sharp upper and lower contacts with the sulphide-rich layers, which are up to approximately 0.1 cm thick.

Chert dominated units can also be found with thin interbeds of clastic material. The clastic layers are often very thin and have sharp contacts with the chert.

Magnetite

Magnetite is found interbedded with chert in the Keewatin and Hematite Lakes area. Magnetite-rich units range in thickness from approximately 0.2 cm to 3.0 cm. Magnetite dominated units contain thin laminations of magnetite and magnetite-rich chert. The laminations range in thickness from approximately 0.01 cm to 0.1 cm. On a microscopic scale the micro-laminations of chert and

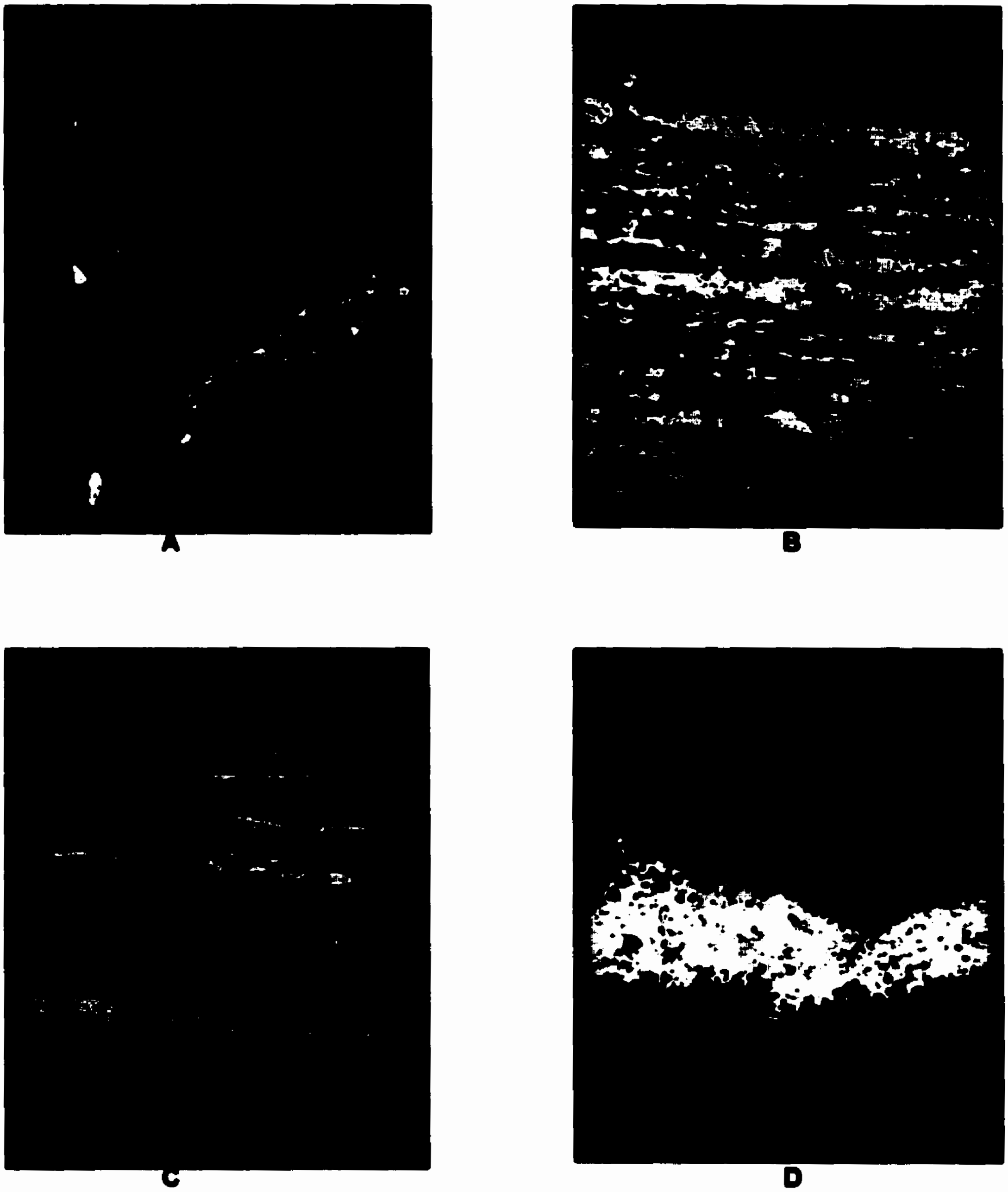


Figure 3.09 SEM photomicrographs taken from core present in the Keewatin Hematite Lakes area. A)Chert-interbedded with carbonate. B) Chert-magnetite iron formation (chert-black, magnetite-white). C)Chert-magnetite iron formation. D)Sulphide-chert iron formation (chert-black, sulphide-white).

magnetite have fairly sharp upper and lower contacts (Fig 3.09B). However, in some cases there appears to be a gradation within units from magnetite domination to chert domination. The lower contact of the magnetite dominated units is commonly sharp, and there is a progressive decrease in the number of magnetite micro-laminations upward, until chert becomes the dominant lithology (Fig. 3.09C).

Carbonate

Carbonate is often found interbedded with chert. Carbonate layers range in thickness from approximately 0.05 cm to 1 cm. The thicker carbonate layers will often show laminations, <0.1 cm thick, visible because of varying amounts of chert mixed with the carbonate. Carbonates are high in iron and are of ankerite or siderite composition (Fig. 3.06). These two carbonate phases are usually found intergrown with one another. The carbonates also have a lower magnesium content than carbonate found in either the Lower Lumby Lake sequence or the Steep Rock sequence. The carbonate layers often have a sharp lower contact and show a gradational decrease in carbonate, and increase in chert, upward, until the layer becomes chert dominated (Fig.3.09A). Alternatively, both the upper and lower contacts with the chert layers may be sharp.

Sulphides

The sulphide in the Keewatin and Hematite lake area is commonly associated with either graphitic slate or chert. Both pyrite and pyrrhotite are the common phases present.

Within the cherts, the sulphides commonly occur as thin laminations, approximately 0.05 to 0.2 cm thick. Sulphide layers have sharp upper and lower contacts with the chert. In some cases carbonate grains, usually siderite, are present at the sulphide-chert contacts (Fig. 3.09D)

Sulphides associated with the graphitic slate are found as thin layers or as nodules within the slate. Thin layers are <0.1 cm to approximately 0.5 cm thick. The layers are usually recrystallized and do not show any internal lamination. Contacts with the slate are sharp and in some cases the contacts are wavy, suggesting soft sediment deformation.

The sulphide nodules that are found within the graphitic slate range in diameter from approximately 0.5 to 2.0 cm. The nodules are commonly randomly located within the slate. However, in some cases they form bands which are parallel to bedding planes in the slates. When cut in half, the larger nodules show millimetre-scale concentric laminations within them.

Graphitic Slate

Graphitic slate in the Keewatin and Hematite Lakes area range from a dense, hard, black form to a dark black, "greasy", fissile form. This is due to the varying proportions of chert, clastic material, and graphite that make up the slate. The hard slates contain more chert with lesser clastic material, while the softer slates are richer in clastic material and graphite.

In most slates thin, millimetre-scale laminations are visible and consist of chert-rich, and clastic-rich layers. Figure 3.10 shows a pyrite layer overlain by a chert layer, with a gradation from chert with scattered clastic grains to a clastic dominated layer in the middle of the photomicrograph. The clastic layer has a sharp contact with a chert layer containing scattered clastic grains. The clastic material is dominantly sericite and chlorite.

Graphitic slate is often associated with sulphides, which may be found disseminated throughout, as thin layers, or nodules within the slate. Contacts with sulphide layers are commonly sharp (Fig. 3.10)

3.4 Hook Lake Area

The Hook Lake area contains sedimentary rocks which are part of the Lower Sedimentary Belt in the Lumby Lake greenstone Belt (Fig. 3.11). Sedimentary rocks in this area are laterally continuous with felsic volcanic rocks

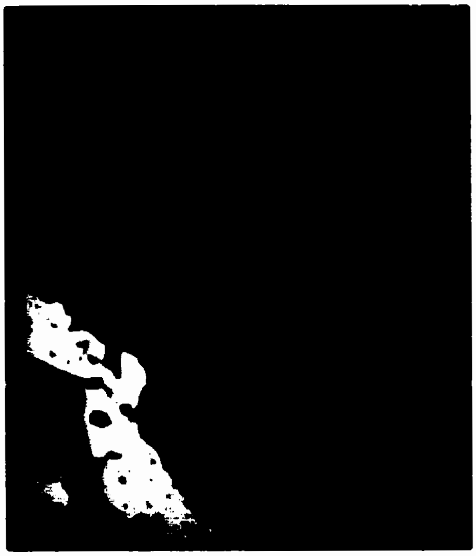


Figure 3.10 SEM microphotograph of pyrite layer (white) overlain by a mixed chert and clastic (grey), overlain by a chert layer (black).

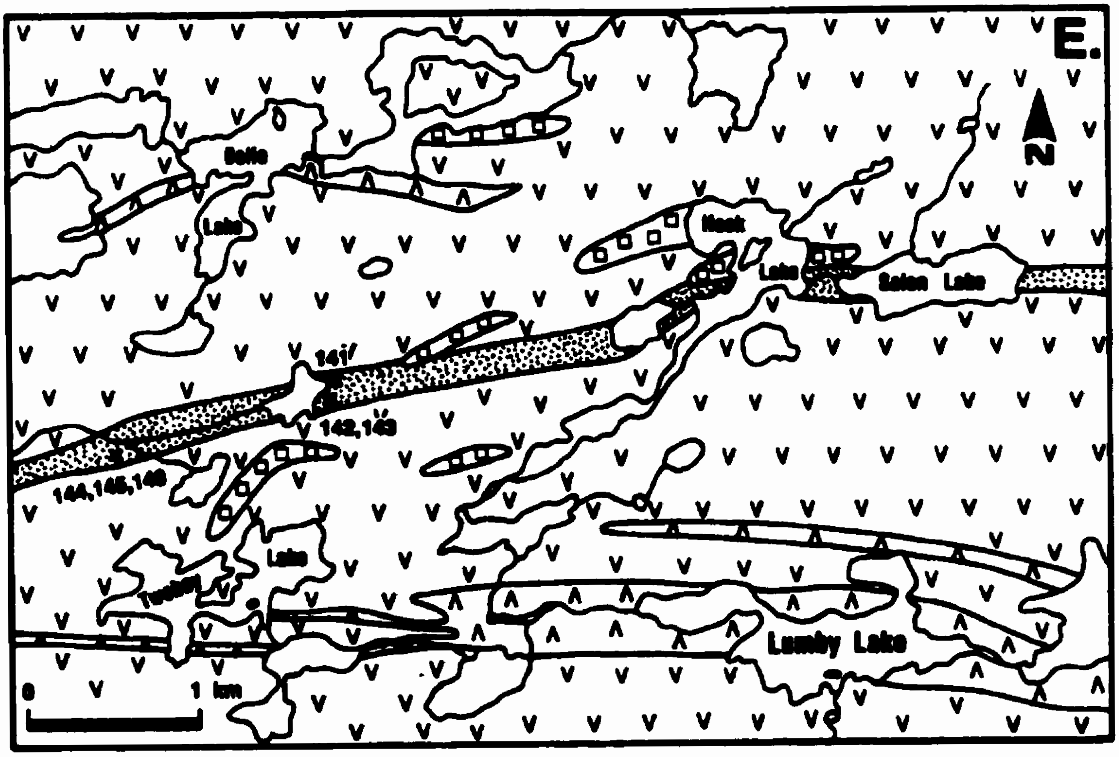


Figure 3.11 General geology of the Hook Lake area. Numbers refer to lithogeochemical sample locations. Refer to Figure 1.1 for location within the Lumby Lake Greenstone Belt and 1.04 for geological legend.

present further to the east (Jackson, 1985). Clastic sedimentary rocks here have a higher proportion of quartz than the main metasedimentary belt and consist of both medium- and fine-grained sandstones, as well as slate. These are overlain by a unit of chemical sedimentary rocks which includes chert and carbonate.

Clastic Sedimentary Rocks

Medium-Grained Sandstone

Sandstone in the Hook Lake area ranges from fine-grained to medium- and coarse-grained. Sandstones are quartz-rich compared to other areas, and contain quartz granules that are commonly 1-3 mm in diameter, but may be up to 6 mm in diameter. The subangular to angular granules are poorly sorted, and are scattered throughout the beds.

Both upper and lower bed contacts are sharp, and easily recognised in the field. Bed thickness appears to be bimodal, with one set ranging from approximately 5 - 10 cm, and the others between 20 - 30 cm. This bimodality is not grainsize dependant. In some areas beds from both thickness groups interbed randomly, while in others the beds are arranged in packages. Often 4 - 5 thin beds (5-10 cm) are overlain by a few thicker beds (20-30 cm), which are in turn overlain by another set of thin beds. These sandstones lack internal structures, and only rarely show grading from a medium- or coarse-grained base to a fine- to medium-grained top. Beds are laterally continuous and show no evidence of thickening or thinning over outcrop lengths of up to 20 meters.

Fine-grained Sandstone

Thinly bedded fine-grained sandstone is common in the Hook Lake area. These are up to approximately 20 cm in thickness, but average a few centimetres or less. Fine-grained sandstone beds are commonly found interbedded with thin slate beds, in units up to approximately 100 cm in thickness. These units consist of alternations between fine-grained sandstone, 0.5 - 3 cm thick, and thin slate beds. Contacts between these layers are sharp, with no indication of any gradation between them.

Fine-grained sandstones are poorly sorted and contain a substantial chlorite matrix component, which gives them a dark grey colour. Bedding contacts are sharp and laterally persistent.

Slate

Slate units are commonly found between the coarser-grained sandstone beds in the Hook Lake area. The thicker bedded, medium-grained sandstones often have a thin slate bed between them that may be up to a few centimetres thick, though it is commonly only 1 cm or less.

Slate is also present interbedded with the fine-grained sandstones, in units up to approximately 100 cm thick. The slate layers average 0.5 to 1.0 cm in thickness. Contacts between the sandstone and slate are sharp, with no indication of a gradation between the layers.

Slate is also found in thick units without the interbedded fine-grained sandstone. These units are up to approximately 100 cm in thickness. Although they are often heavily sheared, there is some indication that the thicker units consist of several thinner layers. Laminations are a few millimetres in thickness, and sometimes visible through slight colour changes within the slate.

Chemical Sedimentary Rocks

Carbonate

A massive iron carbonate unit, up to approximately 800 cm thick, is present in a pit west of Hook Lake. The carbonate is of ankerite composition (SEM-EDS identification), but differs from other carbonates in the Lumby Lake area. The Hook Lake carbonates have high magnesium concentrations similar to the carbonate found in the Steep Rock succession (Fig. 3.09). The carbonates of ankerite composition in other areas within the Lumby Lake Belt have much lower magnesium concentration, and a higher iron to magnesium ratio.

Carbonates appear to be massively bedded. They are coarse-grained and highly recrystallized, which may have disturbed any of the original bedding that was present. In some areas possible beds 20 - 40 cm thick are present. This is not very distinct and may not be a true reflection of bed thickness for the entire unit.

Chert

Chert in the Hook Lake area is found interbedded with darker clay-rich chert layers. The purer chert is white in colour, while the clay-rich cherts are a dark grey to black. The cherts are thinly laminated, 0.1 to 0.5 cm thick, and show well developed alternations between white chert and grey chert. Contacts between layers are sharp, and they do not appear to grade into one another. Although cherts are thinly laminated, the individual layers are laterally persistent over outcrop widths of 1-2 meters.

CHAPTER 4

STRATIGRAPHY OF THE FINLAYSON LAKE GREENSTONE BELT

The metasedimentary belt in the Finlayson Lake area can be divided into two sequences. The metasedimentary rocks in the Southern Finlayson Lake, and the Northern Finlayson Lake areas form a laterally continuous assemblage which constitute a well developed coarsening upward sequence, and conformably overlies the surrounding mafic volcanic rocks. The metasedimentary sequences exposed in the Little Falls Lake area and in the lower Finlayson Lake area are lower in the overall stratigraphy of the Finlayson Lake Greenstone Belt and are likely lateral equivalents.

4.1 Little Falls Lake Area

The nature of the Little Falls Lake sedimentary rocks makes it difficult to give a detailed stratigraphic analysis. Sedimentary rocks in the Little Falls Lake area are often poorly sorted and massive, with unclear bedding contacts and only rarely visible younging indicators such as graded bedding or cross-stratification. For these reasons only a broad stratigraphy of the metasedimentary rocks and surrounding volcanic rocks can be given.

Along the south-eastern part of the Little Falls Lake area the metasedimentary unit is intruded by a gabbroic unit which separates the metasedimentary unit from the adjacent mafic volcanic rocks. The contact with the volcanic unit along the north-western edge was not observed in outcrop. However, Stone et al. (1992) suggest that the adjacent mafic volcanic rocks underlie the metasedimentary sequence, which occupies the core of a large scale syncline with the fold axis striking northeast, parallel to the trend of the sedimentary belt. The few observed younging indicators in the sedimentary sequence suggest that the sedimentary rocks do form a syncline. The maximum width of the sedimentary belt is approximately 1000 meters in the Little Falls Lake area, which suggests a vertical thickness of approximately 500 meters for the sedimentary sequence. The coarsest-grained sandstones and the conglomerates are found near the centre of the sedimentary belt, suggesting that the Little Falls Lake metasedimentary rocks form a coarsening upward succession.

Section LF-1 was taken through the metasedimentary rocks near the edge of the belt. The section is approximately 50 meters thick and is dominated by medium-grained, arkosic sandstone (Figure 4.01). There are several interbeds of mafic detritus-rich sandstone and slate. These are more common in the lower part of the section. Near the top of this section medium-grained arkosic sandstones and coarse-granular sandstones dominate, and the average bed thickness is greater.

Section LF-2 was taken near the middle of the sedimentary belt (Figs. 1.05; 4.02). The section is approximately 13 meters in vertical thickness and begins

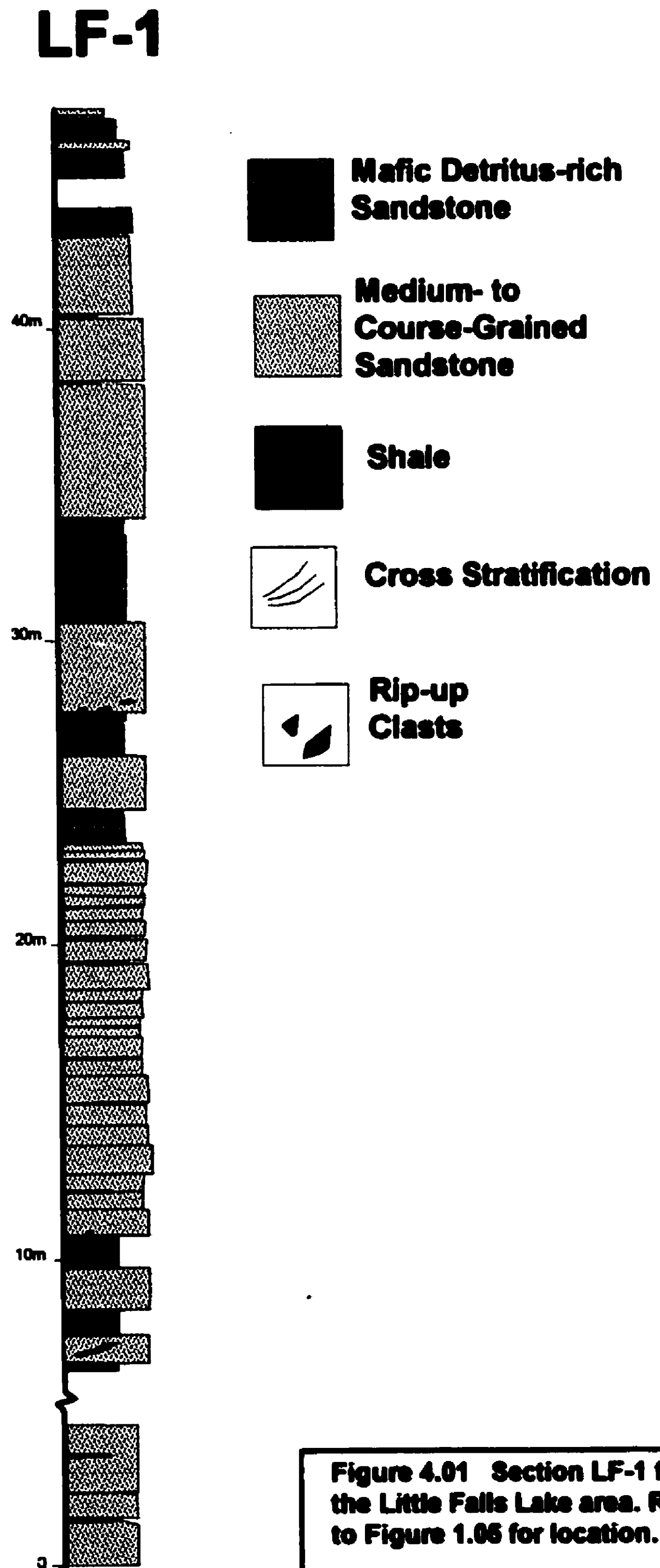


Figure 4.01 Section LF-1 from the Little Falls Lake area. Refer to Figure 1.06 for location.

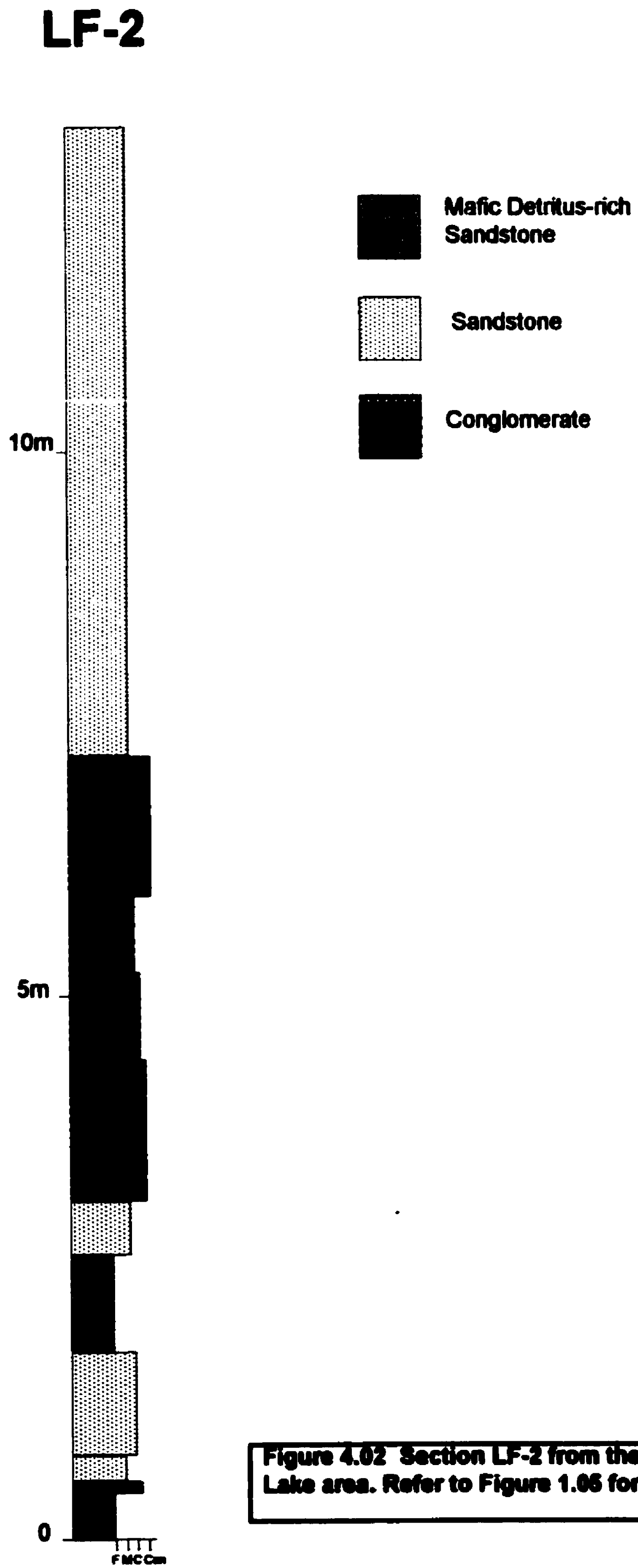


Figure 4.02 Section LF-2 from the Little Falls Lake area. Refer to Figure 1.05 for location.

with 3 meters which is dominated by medium-grained arkosic sandstones interbedded with mafic detritus-rich sandstone beds. These are overlain by approximately 4 meters of monomictic conglomerate, containing medium- to coarse-grained arkosic sandstones. Above the conglomerate unit the section is dominated by coarse granular sandstones and medium-grained arkosic sandstones.

4.2 Southern Finlayson Lake

The Finlayson Lake greenstone belt trends in a general northeast-southwest direction, striking approximately 40° - 220°. Schistosity in the rocks nearly parallels the trend of the belt and ranges in strike from approximately 40° - 70°, dipping between 70° - 90° northwest or southeast (Fenwick, 1976). Several major folds have been mapped within the sequence, with fold axis striking parallel to the trend of the belt (Fenwick, 1976, Stone et al., 1992). Younging directions and top indicators are common in the metavolcanic rocks in the form of pillow tops, and abundant in the metasedimentary rocks as graded bedding. These outline the major fold axis. The metasedimentary rocks in the Finlayson Lake area occupy the core of a large-scale synclinal axis that runs nearly parallel to the trend of the belt. The contact between the metavolcanic rocks and the metasedimentary rocks is conformable, on both the southern and northern limbs of the syncline, with the sedimentary rocks directly overlying the metavolcanic rocks. The contact is visible in several outcrops in both the southern and northern Finlayson Lake area.

Metavolcanic rocks at the base of the sequence, near the contact with the metasedimentary rocks, contain thin units of graphitic argillite and sulphide facies iron formation, and are gradational to the metasedimentary sequence. The sequence of metasedimentary rocks in the southern Finlayson Lake area reaches a maximum thickness of approximately 220 meters, with excellent exposure on the southeast shore of the lake and the small islands near the shore. The Finlayson Lake metasedimentary rocks exhibit well developed vertical and lateral trends. Metasedimentary rocks become more thickly bedded and coarser-grained up section. A lateral fining and thinning trend is also evident (Fig. 4.03). The coarsest metasandstones and conglomerates are present in the southern Finlayson Lake area, while the metasedimentary rocks in the northern area tend to be somewhat finer-grained. It is possible that only the basal portion of the succession is preserved in the northern Finlayson Lake area.

A well developed coarsening and thickening upward sequence is developed in the metasedimentary rocks in the southern Finlayson Lake area. The metasedimentary rocks can be divided into several units which are dominated by one of the sedimentary lithofacies or a group of associated lithofacies. The division between each unit is not fixed, as there is a gradation between most sedimentary facies found in the Southern Finlayson Lake area.

A fine-grained assemblage is found at the base of the metasedimentary

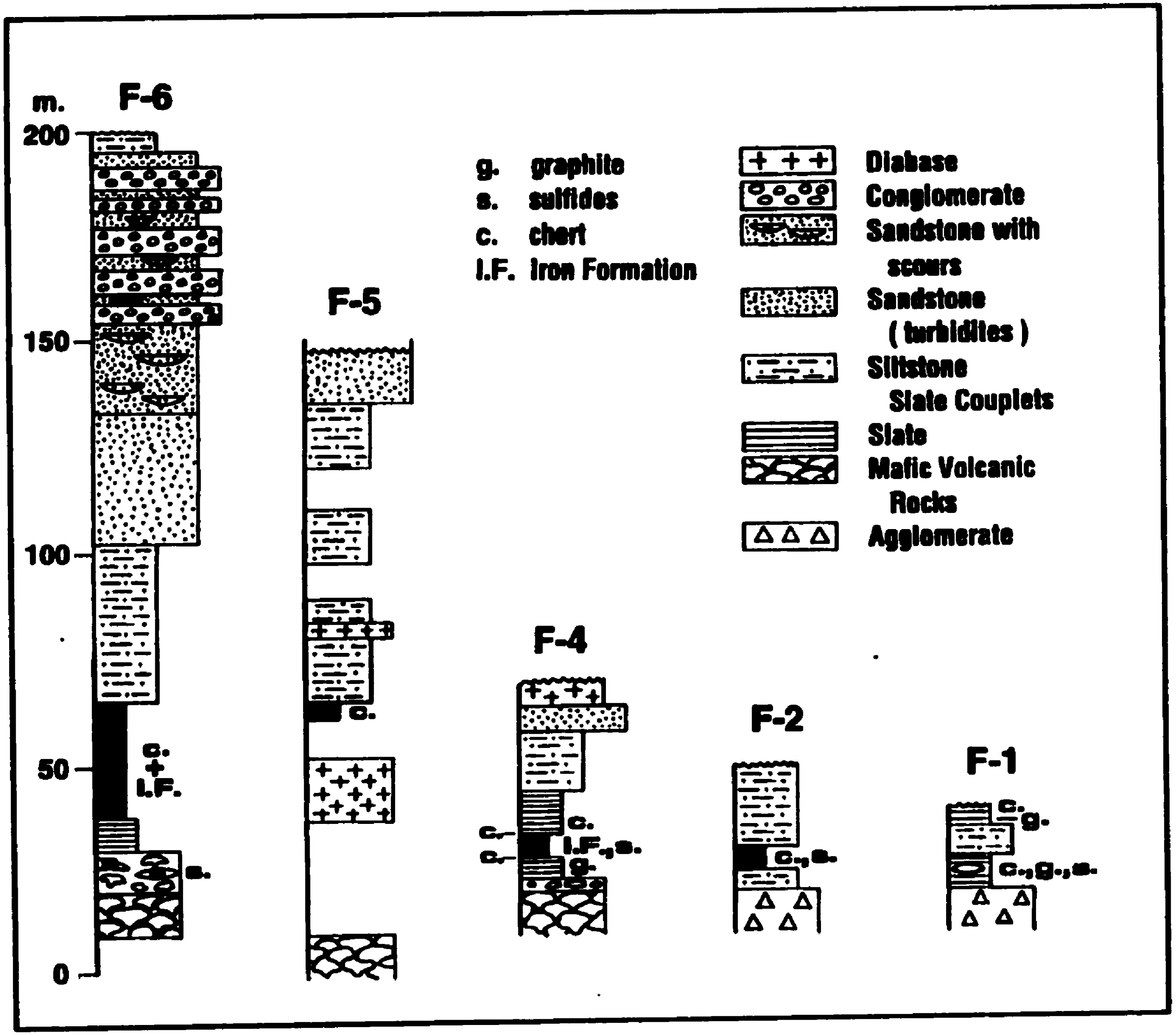


Figure 4.03 Stratigraphic sections measured within the Finlayson Lake area. Refer to Fig. 1.04 for location of section. Section F-6 is an idealized section, covered intervals are interpretated.

sequence, directly overlying the metavolcanic rocks (Figure 4.03 section F-6). This unit ranges in thickness from several meters to approximately 50 meters. Either thinly bedded slates or banded chert-magnetite iron formation dominate the fine -grained unit. These two facies can be interbedded, and can also contain thin interbeds of graphitic slate and sulphide facies iron formation. The basal chert magnetite iron formation is up to 20 meters or more in thickness. The thickest unit observed consists of approximately 5 meters of iron formation overlain by a 10 meter covered interval which was followed by another 6 meters of iron formation. The iron formation may be thicker than the basal 21 meters present in this section, as the contact with the overlying metasandstones is not visible and occurs somewhere in a 50 meter interval which is covered by the lake.

The slate and iron formation unit is overlain by a unit of thin to medium bedded fine to medium-grained sandstone (Figure 4.03 section F-6). Beds are commonly graded from a medium-grained base to a fine-grained sandstone or siltstone top, as described in previous chapters. The vertical thickness of this unit is variable and ranges from 15 meters to approximately 80 meters.

The fine- to medium-grained metasandstones are gradational to and overlain by a unit dominated by medium- to coarse-grained metasandstones bedded on a medium scale. This unit is up to approximately 30 meters in thickness. Near the upper portion of the unit the metasandstones become interbedded with the coarse-grained metasandstone facies which contains scallop shaped scours.

The coarsest-grained unit, found at the top of the sedimentary sequence, is a unit dominated by coarse-grained metasandstones and conglomerate beds. This unit is up to approximately 50 meters in thickness (Figure 4.03 section F-6). The coarse-grained metasandstones, containing low angle scours, are interbedded with lenses of polymictic conglomerate.

The units form a pronounced coarsening and thickening upward sequence, which shows a complete gradation throughout, from thinly bedded slate and iron formation, to thickly bedded conglomerate and coarse-grained metasandstone beds. Two smaller-scale, less than 10 meter thick, coarsening upward sequences were present at the top of this succession. The coarser units are abruptly overlain by a gradation from medium- to coarse-grained sandstone through to conglomerate interbedded with coarse-grained sandstone in the upper cycle.

Section F-4 was taken through the metasedimentary sequence at the northern end of Snake Island (Fig. 4.04, section F-4). The metasedimentary rocks overlie a unit of mafic volcanic rocks. However the contact is not visible due to a covered interval and the intrusion of diabase in the area. The metasedimentary sequence begins with a 3 meter thick unit of thinly layered cherts with variable amounts of clastic material mixed in. The chert unit is in sharp contact with an overlying clastic unit approximately 15 meters in thickness. This unit begins with fine-grained sandstone and siltstone with bed thicknesses between 0.1

centimetres and 1 centimetre. There is a slight increase in bed thickness up section to approximately 2-3 centimetre thick beds. A 40 meter diabase intrusive separates this unit from 4.5 meters of overlying thinly bedded, fine-grained sandstones. This unit continues for another 35 meters, with two covered intervals of 7 meters and 10 meters within the unit. Moving up section the metasedimentary sequence becomes coarser-grained. The last 9 meters of this section is dominated by medium- to coarse-grained metasandstones which are bedded on a medium scale. These are interbedded with thinner units of fine-grained sandstone and siltstone which are up to approximately 1 meter thick. The sedimentary sequence above this section is covered by the lake and not visible.

4.3 Northern Finlayson Lake

The band of metasedimentary rocks in the northern Finlayson Lake area is laterally continuous with the metasedimentary sequence found in the southern Finlayson Lake area. A similar coarsening upward trend to that observed in the Southern Finlayson Lake area is developed. The metasedimentary rocks in the northern part of Finlayson Lake are somewhat finer-grained, and bedded on a thinner scale than in the Southern Finlayson Lake area. The finer-grained metasedimentary rocks tend to form thicker units than those found in the south, and the entire metasedimentary sequence becomes thinner. It is possible that the coarser-grained upper units are not preserved. Here too, the metasedimentary rocks conformably overlie the metavolcanic rocks, although here they may overlie either a thick agglomerate unit or a unit of pillowed volcanic rocks.

Section F-5 represents a sequence measured through the southern part of the western band of sedimentary rocks in the Northern Finlayson Lake area (Fig. 1.04). The metasedimentary rocks here conformably overlie a 3 meter thick unit of pillow breccia, which in turn conformably overlies a unit of pillowed mafic volcanic rocks. The metasedimentary sequence at this location totals approximately 40 meters, which is not likely the total thickness of the sequence due to a diabase intrusive cutting its top (Figure 4.03, section F-5). The metasedimentary sequence begins with a unit of thinly bedded graphitic slates which is approximately 5 meters thick. These grade into an iron formation which is composed of approximately 5 meters of interbedded chert and graphitic chert and a 1 meter thick sulphide facies iron formation showing millimetre-scale layering. The iron formation is overlain by a 10 meter thick unit of thinly bedded slates which have a variable amount of interbedded chert. This unit grades into approximately 6 meters of fine-grained sandstone and siltstone couplets which show thin- to medium-scale bedding. The sequence continues to coarsen upward as the fine-grained metasandstones become interbedded with medium- to coarse-grained sandstones for the remaining exposed 15 meters of the metasedimentary sequence. Near the top of the section medium to thickly bedded, medium- to coarse-grained metasandstones dominate, although these have thinly bedded fine-grained sandstone and siltstone units, up to approximately 1 meter thick, interbedded with them. The section ends with an intrusive diabase cutting the metasedimentary sequence.

The metasedimentary sequences in the most northern part of Finlayson Lake show similar coarsening upward successions as those in other areas of northern Finlayson Lake and those in southern Finlayson Lake. Again the metasedimentary sequences conformably overlie the metavolcanic rocks. However, here the top of the volcanic assemblage is a thick intermediate meta-agglomerate unit (Figure 4.03, section F-1). The base of section F-1 begins with approximately 160 meters of pillowed mafic volcanic rocks. The pillowed flows contain interbedded units of thinly bedded chert, graphitic slate, sulphide facies iron formation and slate, which are up to 5 meters in thickness. These are conformably overlain by a unit dominated by thickly bedded intermediate agglomerate, which contains interbedded units of graphitic slate, sulphide facies iron formation and chert, with thicknesses commonly less than 1 meter. The agglomerate unit is approximately 120 metres thick and beds become thinner with smaller clasts upward. The agglomerate is overlain by a 6.5 meter thick unit of thinly bedded graphitic slate, which contains thin interbeds of chert and sulphide facies iron formation. This unit grades upward into a 6 meter thick unit dominated by medium- to coarse-grained metasandstone which is interbedded with fine-grained sandstone and siltstone couplets. The section ends with approximately 2 meters of thinly bedded graphitic slate. This is likely not the end of the sedimentary sequence, however the rocks become covered by overburden at this point.

At outcrop #18, where section F-2 was measured, the basal contact of the sedimentary sequence with the underlying metavolcanic rocks is not visible. Section F-2 begins with approximately 13 meters of polymictic conglomerate. The conglomerate is overlain by approximately 4.5 meters of thinly bedded, fine-grained sandstone and siltstone couplets, which are succeeded upwards by a 4.5 meter iron formation. This unit consists of interbedded chert, sulphide facies iron formation and graphitic slate (Figure 4.03, section F-2). The fine-grained sandstone beds at the top of the underlying unit, near the contact with the iron formation, have chert-rich tops. The iron formation unit is overlain by approximately 18 meters dominated by thinly bedded, fine-grained sandstones. This unit contains interbeds up to a few meters thick of medium- to coarse-grained sandstone which are bedded on a medium scale. The sedimentary sequence here is capped by approximately 2 meters of dominantly medium- to coarse-grained sandstones which is then covered by overburden.

CHAPTER 5

STRATIGRAPHY OF THE LUMBY LAKE GREENSTONE BELT

The Lumby Lake greenstone belt trends in an east-west direction from the Redpaint Lake area east to near Upsala. The metasedimentary rocks in the Lumby Lake area are separated into an upper and a lower sequence. The upper sequence extends continuously from west of Norway Lake, near Seahorse Lake, through the Pinecone-Cryderman Lakes area, and east to the Hematite Lake area where they are intruded by the Van Nostrand Lake Stock. The upper metasedimentary rock sequence ranges in total thickness from approximately 800 meters in the Norway Lake area to approximately 150 meters in the eastern area, and is composed of a clastic dominated unit and a chemical precipitate dominated unit.

The Lower Lumby Lake metasedimentary rock sequence follows the east-west trend of the belt, and extends from Redpaint Lake, just north of Two Bay Lake, east to Rea Lake, where it appears to be continuous with a felsic volcanic unit extending further east (Jackson, 1985). This sequence is dominated by clastic metasedimentary rocks, with lesser chemical precipitates. The lower sequence is consistently much thinner than the upper sequence, and attains a maximum thickness of only 200 meters.

5.1 Norway Lake area

The metasedimentary rocks in the Norway Lake area are the western extension of the Upper Metasedimentary Belt. The metasedimentary belt is surrounded by mafic volcanic rocks, dominantly massive, with lesser amounts of pillowed flow rocks. Younging direction, obtained from pillowed flows, suggest that the mafic volcanic sequence youngs towards the metasedimentary sequence in both the northern metavolcanic rocks and the southern metavolcanic rocks. The metasedimentary sequence appears to be a homoclinal sequence composed of a clastic dominated unit and a chemical precipitate dominated unit. The clastic unit reaches a thickness of approximately 600 meters near Garnet Bay, Norway Lake, however folding may have repeated some of the sequence exaggerating the thickness of the unit. The unit of chemical precipitates in this area reaches a total thickness of nearly 200 meters. Younging directions in the metasedimentary rocks are rare, and evidence for younging in both a north and a south direction are visible. It has been suggest by Woolverton (1960) and Jackson (1985) that a synclinal fold axis parallels the southern contact between the metasedimentary belt and the southern mafic volcanic rocks, however the sequence of metasedimentary rocks are not repeated as would be expected on the other limb of the fold axis. Another possibility is that the metasedimentary rocks are in fault contact with the mafic volcanic rocks along either the northern or southern edge of the metasedimentary belt. This would suggest either of two possibilities for the stratigraphy of the metasedimentary rocks. The clastic unit may form the base of the sequence, with the chemical precipitates overlying, or the chemical precipitate unit may form the base of the sequence and the clastic unit overlies them.

If the clastic unit forms the base of the metasedimentary sequence, the stratigraphy in the Norway Lake area would begin with fine- to medium-grained metasandstones overlying the mafic volcanic rocks. The sandstone unit is between 100 and 300 meters thick in the Garnet Bay area. A conglomerate unit overlying the fine- to medium-grained sandstones is up to approximately 50 meters in thickness. A rough gradation in average clast size, from smaller to larger in a northward direction, over an outcrop of conglomerate approximately 10 meters thick , may represent younging to the north and a coarsening upward trend in the sequence, similar to correlative units in the Finlayson Lake area. A unit dominated by coarse recrystallized carbonate overlies the conglomerate unit. The marble is up to approximately 80 meter in thickness and is interbedded with siliceous layers up to a few centimetres in thickness. Overlying the marble is a unit that is dominated by banded chert magnetite iron formation. This unit begins with chert-rich, fine-grained sandstones near the base, which grade to purer chert interbedded with magnetite and magnetite-rich chert layers near the top. The iron formation unit reaches a thickness of nearly 200 meters in the Norway Lake area.

The alternative possibility is that the unit dominated by chemical precipitates overlies the mafic volcanic unit, and the clastic sedimentary rock unit forms the upper portion of the sequence. If this is the case the sequence described previously would be reversed. Iron formation would form the base of the sequence, and overlie the mafic volcanic rocks. The iron formation would exhibit an upward increase in disseminated clastic material. Marble would overly the iron formation and in turn be overlain by conglomerate. In this scenario, the conglomerate unit is overlain by fine- to medium-grained metasandstones which form the top of the metasedimentary sequence.

5.2 Pinecone Lake Area

The metasedimentary rocks in the Pinecone Lake area are continuous with the metasedimentary rocks found near Norway Lake. The unit of clastic sedimentary rocks becomes somewhat thinner in this area, approximately 350 meters, and the sandstones are finer-grained. The chemical metasedimentary unit thickens to approximately 200 meters.

The sequence here is similar to that in the Norway Lake area. Several fine-grained sandstone beds along the southern shore of Pinecone Lake suggest that younging is to the north. Further evidence for north younging comes from the formation of garnet porphyroblasts within the northern, argillaceous portion of the sandstone beds. Garnet porphyroblasts up to a few millimetres in size form in the finest-grained material of each bed, and are found consistently in the northern portion of the sandstone beds. If the clastic unit forms the base, as suggested by the younging indicators, the sequence begins with approximately

350 meters of fine-grained sandstones and interbedded slate. There are isolated pockets of graphitic slate and associated sulphide facies of iron formation found within the sequence of fine-grained sandstones. The graphitic slates form lenses which may be up to a few meters thick and are usually found near the southern margin of the clastic unit, near the base of the metasedimentary rocks if the package consistently youngs to the north.

A unit of recrystallized carbonate overlies the clastic sedimentary rocks at the west end of Pinecone Lake. The carbonates are tightly folded, and it is difficult to give a true thickness of this unit due to the possibility of fold repetition. Outcrops of marble are up to several meters in width and total thickness may be up to 100-200 meters as suggested by Jackson (1985). The carbonates are overlain by a unit of chert and oxide facies iron formation. Chert directly overlying the carbonates is fairly impure, containing a high proportion of disseminated clastic material. Moving up the sequence the chert layers become "cleaner" and are interbedded with layers of magnetite and magnetite-rich chert. The total thickness of the chert and iron formation unit is up to approximately 150 meters in the Pinecone Lake area.

5.4 Keewatin and Hematite Lakes

The stratigraphy in the Keewatin and Hematite Lakes area is unclear due to a scarcity of outcrop, and a lack of younging direction indicators within the rocks. However, some indication of the relative stratigraphy can be gained from old diamond drill core present in the area. In the early 1950's, Candella Development Corporation drilled several diamond drill holes in the Keewatin and Hematite Lakes area. Some of the core is still present near an old core shack on the shore of Hematite Lake. Much of the core has been scattered around, but several intact boxes are present. Although there are few indications of younging, or the specific hole each box of core came from, continuous sections of core within a single box can be useful in determining the relative stratigraphy and how the different units relate to one another.

A wide range of sedimentary rocks and chemical precipitates are found in the core including medium grained sandstones, slate, graphitic slate, chert, magnetite and sulphides. All of these different facies are intimately associated with one another and several may be present within a single short measured core section. Some representative descriptions of sections follow.

Section H-4 (Figure 5.01A) represents a transition from chert to a more clastic dominated sequence near the top. The section begins with approximately 1 meter of chert containing disseminated pyrite and pyrrhotite. Recrystalization of the sulphides has destroyed any original layering that may have been visible. Above the chert unit there is 90 centimetres of alternating chert and fine silt, within layers ranging from 1 mm to 1 cm thick. The layers have a sharp lower contact with pure chert and grade into a darker silt-rich chert near the top of each laminae. Recrystallized sulphides are again present within the more pure chert layers and the upper silt-rich portion of each bed is magnetite rich. This

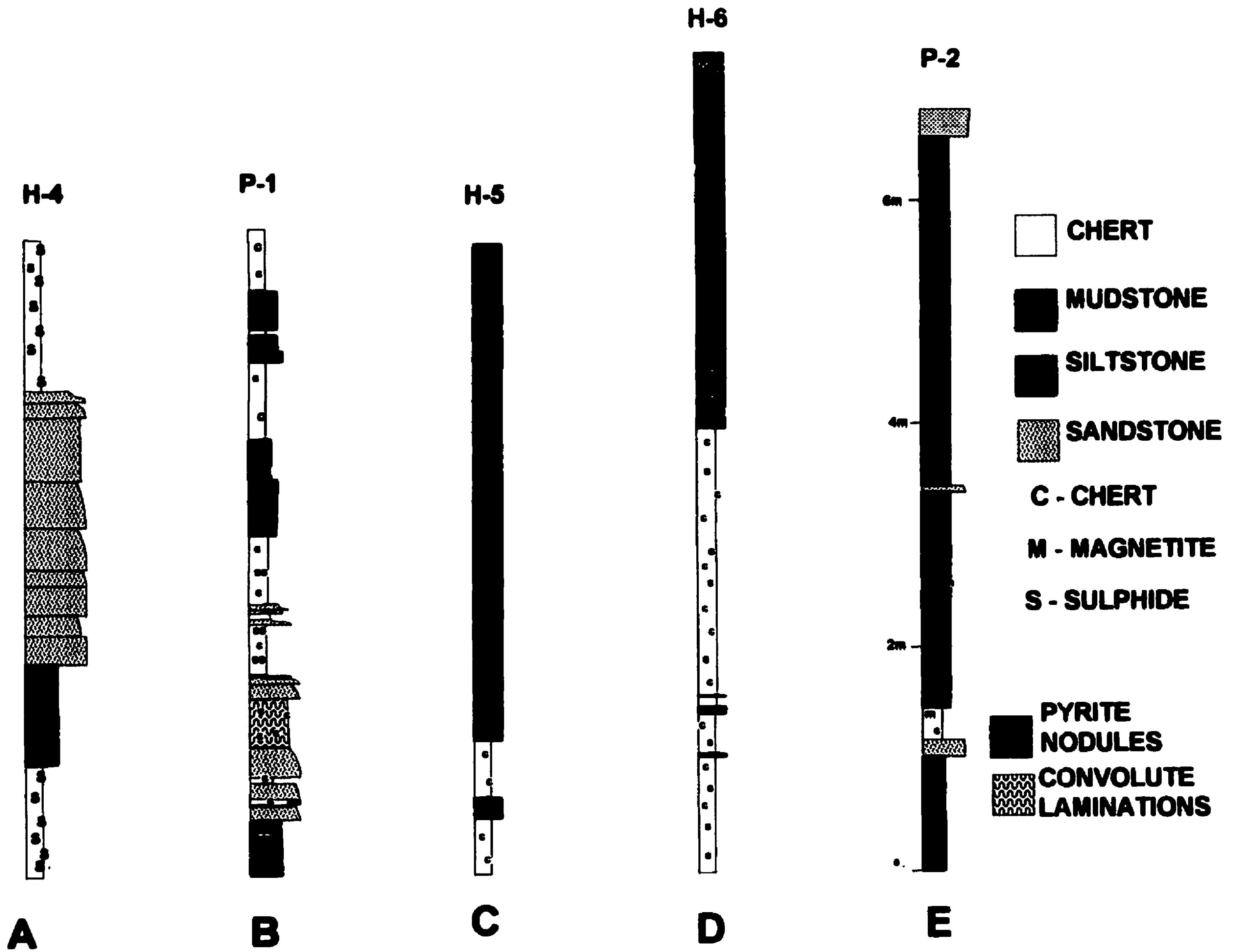


FIGURE 5.01 Stratigraphic sections measured from drill core within the Keewatin Lake area.

unit has a sharp upper contact with 2.45 meters of medium-grained sandstone containing 15-55 centimetre thick beds. Each bed has a sharp lower contact with medium-grained sandstone and grades into a fine-grained sandstone or siltstone top. Above the sandstone unit there is 70 cm of chert with disseminated sulphides similar to the beginning of this section.

Section H-6 (Figure 5.01D) begins with approximately 4 meters of chert containing disseminated pyrite as 1-2 mm cubes with no indication of layering. This unit contains rare 2-5 cm thick interbeds of graphitic argillite within it. The graphitic argillite beds have sharp upper and lower contacts and contain thin, sub-millimetre scale laminations of chert within them. The chert unit has a sharp upper contact with a 3.3 meter graphitic argillite dominated section. The graphitic argillite can be divided into three types. The argillite may be fairly pure or contain thin sulphide laminations or ovoid pyrite concretions. The concretions are 0.5 to 1 centimetre in diameter and contain concentric internal laminations. Where the argillite contains laminations of pyrite the laminations are very thin, usually less than 1 millimetre.

Section H-5 (Figure 5.01C) also shows a transition from a chert dominated unit into a graphitic argillite dominated unit. The section begins with approximately 50 centimetres of chert containing variable amounts of silt, giving the core a greyish white colour. The chert contains disseminated sulphides as small cubes scattered throughout. Above this is 22 centimetres of interbedded chert and graphitic argillite. The chert layers average approximately 0.5 but range from 0.1 to 2.0 centimetres in thickness. Chert layers have sharp upper and lower contacts with dark black graphitic argillite layers 0.3-1.0 centimetre thick. The alternating chert and argillite is succeeded upward by a 50 centimetre thick unit of chert, similar to the beginning of the section, but containing less disseminated sulphides. This is overlain by 10 centimetres of alternating chert and argillite layers containing disseminated sulphides. The chert layers are more pure, white in colour, and have wavy bedding planes, possibly due to soft sediment deformation. Overlying this is 450 centimetres of graphitic argillite containing <1 to 1millimeter thick sulphide rich laminations.

Section P-1 (Figure 5.01B) contains units of sandstone interbedded with very fine-grained green mudstones and chert. The section begins with 50 centimetres of 1-10 millimetre thick, black to green coloured mudstone layers. Each layer has a sharp basal contact with fine-grained, black siltstone which grades upward into a dark green mudstone. Layers are commonly separated by a thin <1-1 millimetre sulphide laminae. The unit has a sharp contact with a series of interbedded fine-grained sandstone and siltstone layers. Sandstone beds are normally graded into a very fine-grained parallel laminated top. Intervening siltstone layers grade to a chert rich top. The siltstone layers commonly appear contorted due to soft sediment deformation. The section becomes more chert-rich upward, consisting of thin chert layers, commonly separated by thin sulphide layers averaging 1 millimetre thick. At approximately 3.1 meters the section consists of 1-3 centimetre thick siltstone to black mudstone beds. Each bed has a sharp lower contact with 1-8 millimetres of siltstone, grading rapidly upward into very fine-grained black mudstone. The lower silty division is parallel laminated on a sub-millimetre-scale. This unit is overlain by 35 centimetres of chert-rich mudstone containing disseminated pyrrhotite, changing upward into more pure chert, and then into parallel laminated fine siltstone and chert again. Overlying this is a series of dark green mudstone layers, up to approximately 40 centimetres thick, interbedded with chert-rich intervals. The chert-rich intervals consist of 0.5-2 centimetre chert beds separating thin mudstone layers up to 2-3 millimetres thick.

Section P-2 (Figure 5.01E) is dominated by interbedded magnetite and green mudstone layers with local cherty layers and rare fine-grained sandstone beds. The section begins with massive magnetite containing contorted layers and wisps of green mudstone. The interval from 40 to 88 centimetres is dominated by thicker green mudstone layers up to 4 centimetres thick, and interbedded magnetite layers up to 2 centimetres thick. This unit is overlain by a series of 1-2 centimetre thick, fine-grained sandstone layers interbedded with thin green mudstone layers. The rest of the section is dominated by massive magnetite containing abundant layers and wisps of the green mudstone with rare fine-grained sandstone to siltstone beds. The section ends with 6 centimetres of green mudstone, showing loading into the underlying magnetite, which is overlain by 20 centimetres of fine-grained sandstone beds. The sandstone beds are 4-5 centimetres thick and are graded to a very fine-grained dark grey siltstone top.

5.5 Lower Lumby Lake

The metasedimentary rocks in the Lower Lumby Lake sequence are continuous with a felsic volcanic unit to the east. The sequence reaches a maximum thickness of approximately 200 meters north of Two Bay Lake near Hook Lake.

The lower contact of the metasedimentary rocks with the underlying volcanic rocks is covered and not visible in section LL-1(Figure 5.02). The section begins with approximately 21 meters dominated by medium- to coarse-grained sandstones, with lesser fine-grained sandstone interbeds. After a 14 meter covered section the sequence resumes with a 55 centimetre unit of thinly laminated chert with thin slaty interbeds. Approximately 8.5 meters of interbedded medium- to coarse-grained and fine-grained sandstone overlie the chert unit. These sandstones are bedded on a medium scale. Overlying the sandstone is an 8 meter thick unit of massive marble, in which bedding contacts are not visible. Above this the marble becomes interbedded with fine-grained sandstone beds for a thickness of approximately 4.5 meters. The next 7.5 meters of section LL-1 is a covered interval, after which the fine -grained sandstone beds become interbedded with units of slate up to approximately 50 cm thick. This unit is approximately 7 meters thick and becomes covered to the south by

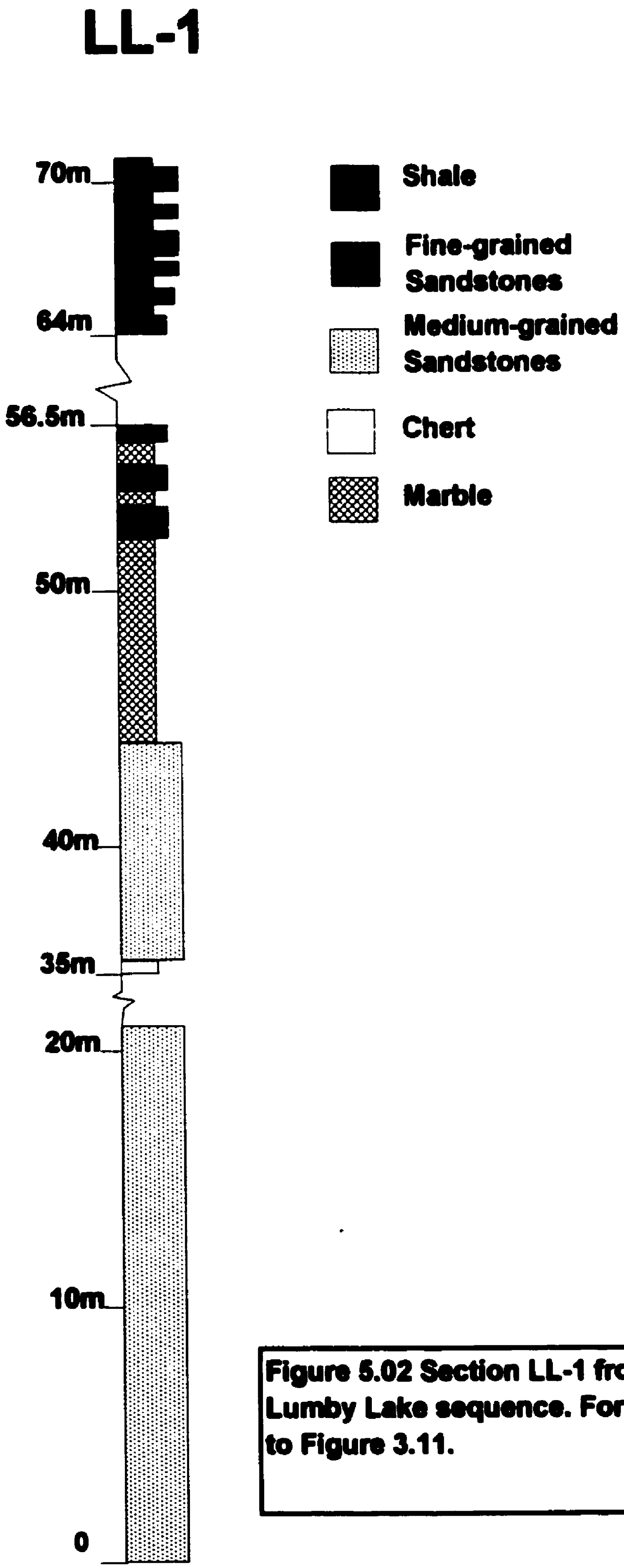


Figure 5.02 Section LL-1 from the Lower Lumby Lake sequence. For location refer to Figure 3.11.

overburden.

Section LL-2 is a short section through a portion of the clastic sequence near the small lake just west of Hook Lake. The section begins with approximately 1 meter of thinly bedded medium- to coarse-grained sandstones (Figure 5.03). Approximately 1.5 meters of thinly bedded fine-grained sandstone and slate overlie the medium-grained sandstones. The next 1.5 meters consist of a few medium- to coarse- grained sandstone beds with interbedded fine-grained sandstone and slate units.

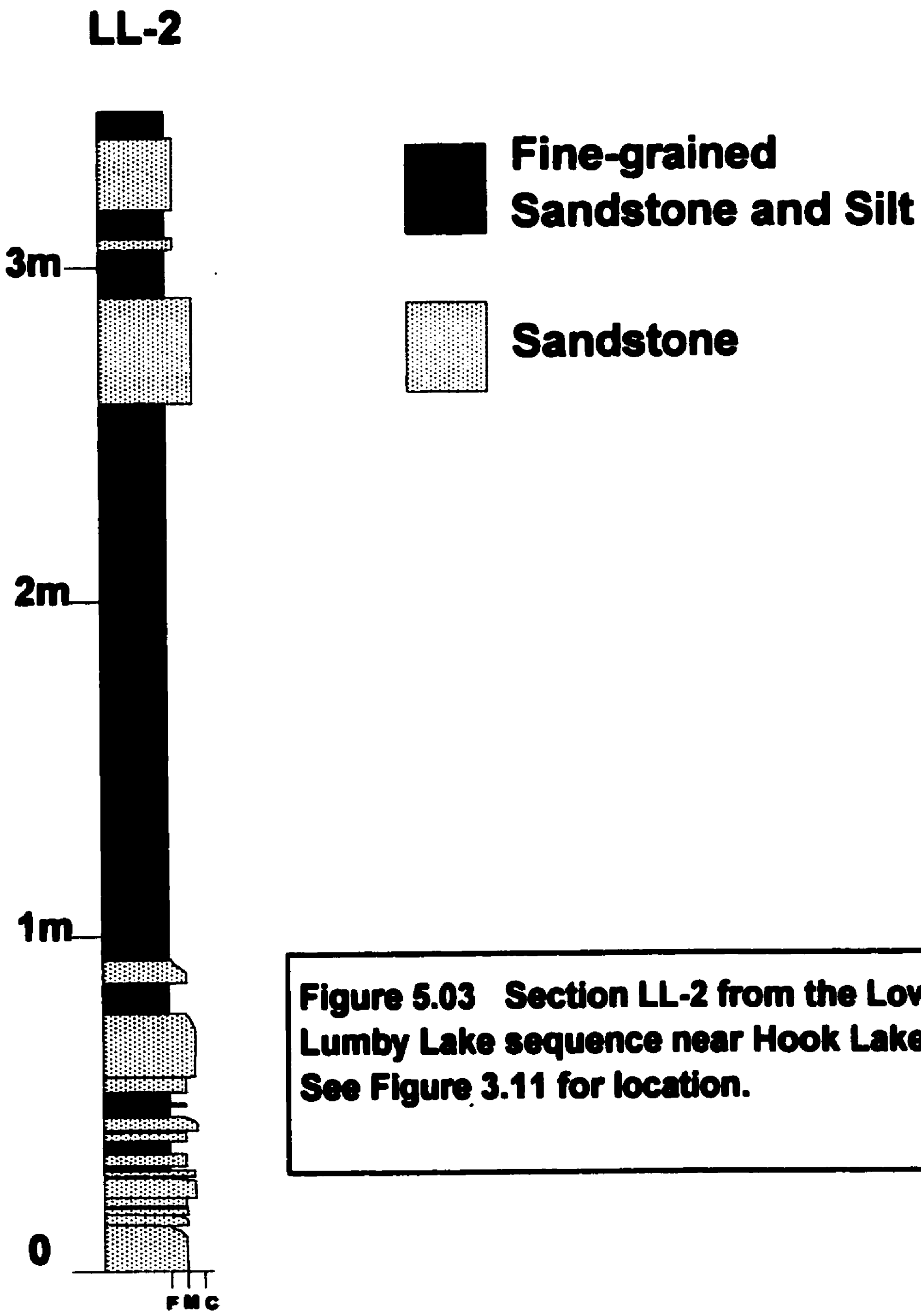


Figure 5.03 Section LL-2 from the Lower Lumby Lake sequence near Hook Lake. See Figure 3.11 for location.

CHAPTER 6

PROVENANCE OF THE FINLAYSON AND LUMBY LAKE SEDIMENTARY ROCKS

6.1 Clast Lithologies

The lithologies of the clasts within conglomerate facies present in the different areas in the Finlayson and Lumby Lakes Greenstone Belts gives some indication of the provenance of the sedimentary rocks. The variation in lithology, as well as the relative abundance of the different types of clasts can give an indication of the composition of the source areas contributing sediment into these basins.

The conglomerate present within the Little Falls Lake area is monomictic, consisting only of felsic volcanic clasts. The clasts are similar in appearance to the matrix sandstones as well as the sandstone facies found within the area, suggesting a single source area, with very little compositional variation. The conglomerate clasts are irregularly shaped, sub-angular to sub-rounded, and the sedimentary sequence is laterally transitional to felsic volcanic rocks. All of these characteristics suggest a nearby source and that the sedimentary rocks in this area represent the more distal, reworked equivalent of the felsic volcanic rocks to the south.

Conglomerates present within the southern Finlayson Lake area are polymictic, and contain clasts with granitic, felsic volcanic, and mafic volcanic compositions. Of 62 clasts within outcrop #6, 50 % are granitic, 26 % felsic volcanic, 13 % mafic volcanic and 11 % were quartz. The granitic clasts are more abundant than either the felsic volcanic, or mafic volcanic clasts. The clasts are subrounded which may indicate that they were transported some distance from their source, and the polymictic nature of the conglomerate suggests a mixing of sediment from several source rocks of different composition and possibly separate source areas.

The conglomerate within the northern Finlayson Lake area is fairly local and contains clasts of felsic volcanic and mafic volcanic composition, as well as rare clasts of banded chert-magnetite iron formation. Of 62 clasts surveyed in outcrop #18, 52 % were felsic volcanic, 19 % intermediate volcanic, 19 % mafic volcanic and 5 % of each chert-magnetite iron formation and chert. The felsic volcanic and mafic volcanic clasts suggest some similarities in source to the southern Finlayson Lake area. However the presence of the iron formation clasts, and lack of granitic clasts may represent a more localized contribution.

Norway Lake conglomerates contain clasts which are similar in composition to those found in the northern Finlayson Lake conglomerates. Felsic and mafic volcanic clasts dominate, with rare chert and iron-formation clasts. Of 203 clasts, 53 % were mafic volcanic, 44 % felsic volcanic, and the remaining 3 % were gabbro, granite, and chert clasts. This suggests a similar source to the northern Finlayson Lake conglomerates

6.2 Sediment Geochemistry

Introduction

If used with caution the whole rock geochemistry of sedimentary rocks can be an extremely useful tool as an indicator of provenance. Sediment geochemistry and source rock geochemistry are rarely linearly related, making it a formidable task to determine provenance. The sediments are inevitably effected by several processes which may alter their chemical composition. A major control on sediment chemistry is the composition of the source rock. The chemical signature of the source rock, or rocks, may be altered by weathering processes, as well as hydraulic sorting effects, which tend to differentiate minerals, by size and mass, during transport and deposition. After deposition the effects of diagenesis, metamorphism, and hydrothermal activity may further alter the composition of the sediments. The combined effects of these processes make a linear relationship between the source rock geochemistry and sediment geochemistry very unlikely.

Many of the earlier researchers focused on the possibility that sediment geochemistry could distinguish between the possible tectonic settings basins form in (Bhatia, 1983; Roser and Korsch, 1986, 1988). These studies found a correlation between major element composition and the nature of the continental margin or oceanic islands sediment was eroded from. Bhatia (1983) showed a progressive decrease in Fe_2O_3 + MgO, TiO_2, Al_2O_3/SiO_2 and an increase in

K_2O/Na_2O and $Al_2O_3/(CaO + Na_2O)$ in sandstones from oceanic island arc to continental island arc to active continental margins to passive margins. Roser and Korsch (1986, 1988) also found that discriminant function analysis using SiO_2/Al_2O_3 and K_2O/Na_2O ratios and $Fe_2O_3 + MgO$ will distinguish tectonic environments and tend to reduce overlap between the primary fields. The basic principle behind these discriminating major element ratios is that the degree of differentiation, and therefore the abundance of acid volcanic rocks relative to basic volcanic rocks, differs within each of these settings. If these rocks differ in their geochemistry, so should the sediments that are derived from them. These studies can be useful, but are restricted to a fairly broad classification of the tectonic setting in which the sediments were deposited. They may not be useful in identifying specific provenance areas, which shed sediments at different stages in the development of the basin. The methods described above are based on major element abundances, which tend to be highly variable depending on differences in compatibility and mobility. The effects of secondary processes such as hydraulic sorting, weathering during transport, diagenesis and metamorphism are also not accounted for with the methods described above.

Recently there has been research into the effects of these processes on the final chemical composition of sedimentary rocks. Weathering processes acting upon source rocks with contrasting composition may result in products having a distinctly different chemistry. The weathering of ferromagnesian minerals tends to yield clays of the smectite-vermiculite type, while weathering of feldspars produces clays of the illite type (Deer et al., 1967). The susceptibility to weathering of minerals with variable composition may lead to loss or gain of certain elements, changing the overall chemistry of the sediment. The susceptibility of feldspars to weathering processes is in the order:

Ca-rich plagioclase > Na-rich plagioclase > K-feldspar

for Archean metasandstones of the Quetico metasedimentary belt (Sawyer, 1986). The preferential removal of Ca, Na, and K by weathering has been determined by Nisbett (1979). These authors have shown that Ca is not retained by the clay minerals produced and is therefore lost to weathering solutions. Almost all of the K released by the feldspars during weathering is retained by the clay minerals that are formed. Na is more likely to be retained by new clay minerals than Ca, but may also be lost to weathering solutions. Effects such as these may result in a depletion of Na_2O and CaO, and relative enrichment of K_2O in the sediments, as compared to the rocks from which they were derived.

Elements such as the rare earth elements (REE), which are commonly used in geochemical studies and thought of as a reliable indicator of provenance, may not be completely immobile. A study by Nesbitt et al. (1980), suggests that weathering processes may cause some degree of mobility and fractionation of REEs, altering the geochemical signature of sediments as compared to that of the source material. Nesbitt suggests that the uptake of CO_2 by rainwater percolating through the organic rich soil profile may cause a pH decrease. These acidic waters may mobilize some of the REE elements and transport them or carry them further into the soil profile where the waters react with less altered rock, increasing the pH values, releasing REE's into suitable sites within the clay phases being produced. This study also suggests that fractionation may take place and is due to differing stabilities of minerals during chemical weathering. Minerals such as hornblende will preferentially take up heavy REE, while light REE may be fractionated into the residual phases. Thus the preferential breakdown of ferro-magnesian phases alters the REE pattern.

Sorting may also have significant effects on REE concentrations. Fractionation is not as common or as significant as might be expected from the Nesbitt (1979) study, and he suggests that this may be due to mechanical mixing and homogenization of the different sized fractions of sediment. Cullers et al. (1979) found that REE's reside mainly within the siltstone and clay fraction, rather than the sandstone fraction of sediments. However the REE patterns of the finer grainsizes and the sands are very similar suggesting that the major control may be a dilution effect, due to the higher quartz content in the sand sized fraction of the sediment.

Hydraulic sorting may effect the concentration of trace elements other than the REE's. The original geochemical composition of the source material may be divided between the minerals found in the clay-size fraction and the sand-size fraction of the derived sediment (Argast and Donnelly, 1987). Elements such as zirconium tend to be extremely immobile but concentrate, along with quartz, in the sand-sized fraction. Hydrodynamic fractionation of the sediment by partitioning of heavy minerals into specific depositional sites, as evidenced in extreme cases by placer deposit formation, is another, often neglected geochemical aspect of sorting (Fralick and Kronberg, 1997).

Theoretical Aspects

These secondary processes make a direct comparison of source rock geochemistry and sediment geochemistry nearly impossible. Even the concentrations of elements within the sediments which are considered to be immobile will not reflect their original concentrations in the source rock due to the constant sum problem. As mobile elements are lost or gained within the system the concentration of immobile elements will be driven upwards or downwards. However, the ratio of immobile element pairs should remain constant provided that the elements behave in a hydraulically similar manner.

The method used in this study is outlined by Fralick and Kronberg (1997). Immobile element pairs can be used provided their relative immobility can be established. A technique developed by MacLean (1990; see also MacLean and Kranidiotis, 1987; MacLean and Hoy, 1991; Barrett and MacLean, 1991; Barrett et al, 1991) to determine elemental mobility during alteration of volcanic rocks, can be used to determine mobility of elements within sediments as well. As stated by Fralick and Kronberg (1997) immobile elements in a rock will increase or decrease in concentration as mobile elements are lost from, or gained by, the rock. In Figure 6.01A two immobile elements are plotted against one another and have starting composition "a". As mobile elements are lost or gained from the system the relative concentration of the immobile elements will increase or decrease, respectively, moving point "a" along a line either away from or towards the origin. As shown in Figure 6.01A the original concentration "a" moves away from the origin as mobile elements are lost from the system, resulting in a final immobile element composition "b". This process represents chemical weathering of the source rocks altering their chemical composition to that of the sediments, represented by "b". If after weathering, hydraulic fractionation occurs, the bulk composition of the sediment at point "b", will be divided between two size fractions, sand (sst) and clay (sh). If, as in Figure 6.01A, the immobile elements concentrate in the clay phases, the "sh" point will have a higher immobile element concentration than the sand fraction (sst), however, both of these points will remain on a tie line passing through the original concentration and the origin. The situation will be reversed if the immobile elements are concentrated in the sand (sst) fraction of the sediment during sorting. If these elements remain immobile during all other processes, such as diagenesis or metamorphism, their concentration will remain on the tie line joining them and the origin. If they do become mobile, as depicted by the dotted lines in Figure 6.01A, they move off the tie line.

The situation for a mobile-immobile element pair is depicted in Figure 6.01B. As the mobile element is lost from the system the point moves from

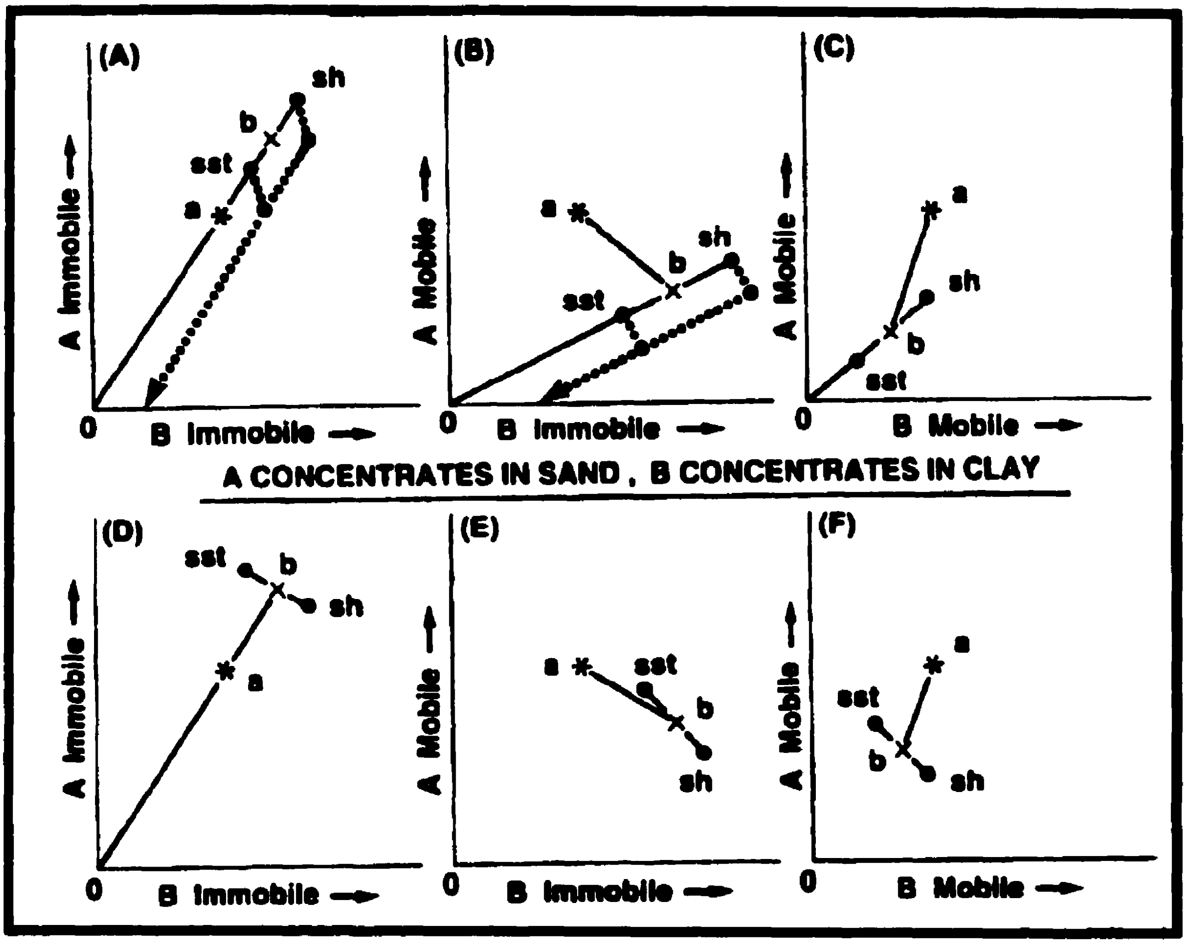


Figure 6.01 Theoretical scattergrams for assessing element mobility. Refer to text for explanation. Reproduced from Fralick and Kronberg (1997).

original concentration "a" to concentration "b". If hydraulic fractionation then acts upon the system, concentrating the immobile elements in the clay "sh" fraction, the sand (sst) and clay (sh) points will lie on a line extending to the origin. If they become mobile at a later time, the line will not extend to the origin, as depicted by the dotted lines.

Figures 6.01C, D, E, and F show typical graphical representations of concentrations for different immobile-mobile element pairs, under different weathering and hydraulic sorting conditions.

One of two patterns will develop if a number of sediment samples with the same source are plotted. If the elements have both remained immobile and have behaved hydraulically similarly, they will plot as a linear function, extending to the origin, as in Figure 6.01A. If these conditions are not met, they will plot as a scatter of points. This situation occurs as a result of either loss from or addition to the system during weathering or differences in physical sorting between the mineral phases containing the elements which are plotted.

A second technique also exists that may be used to determine the effects of chemical weathering and hydraulic sorting. The basis of this method is that chemical weathering serves to destroy all major mineral phases in a rock except quartz, and assumes that a higher quartz content reflects maturity of the sediment. As explained by Fralick and Kronberg (1997), the element suspected of being immobile is plotted against the SiO_2 content of the sediment. A sediment weathered from any source rock will split into two fractions, sand (sst) and clay (sh). When an element which is concentrated within the clay fraction of the sediment is plotted against SiO2 the concentration in the sand (sst) fraction and the clay (sh) fraction will lie upon a vector extending from the original composition to the 100% SiO_2 point on the vertical axis (Figure 6.02A). For an immobile element which concentrates in the sand (sst) fraction the position of the clay and sand points will lie upon a vector extending from the origin through the original concentration (Figure 6.02B).

When several sediment samples are plotted against SiO_2, a linear trend extending to the 100% SiO_2 point on the vertical axis will develop if the element has remained immobile and has concentrated in the clay fraction. A linear trend extending to the origin will develop if the immobile element in the samples concentrated in the sand fraction. If the immobile element is not partitioned into either the sand or clay fraction, or a mobile element is plotted the vector will not extend to either of the end member compositions. A scatter of points will be produced if the original composition has been altered significantly as seen in Figure 6.02C, D, E, and F.

If the relative immobility of an element and its hydraulic behaviour can be established the element can prove very useful in determining the source rock composition. Once immobility and similar sorting behaviour are established the elements can be plotted as ratios which should be similar to the ratios of these elements in the source rocks. Although the concentration of these immobile elements may differ significantly from the concentration in the source material,

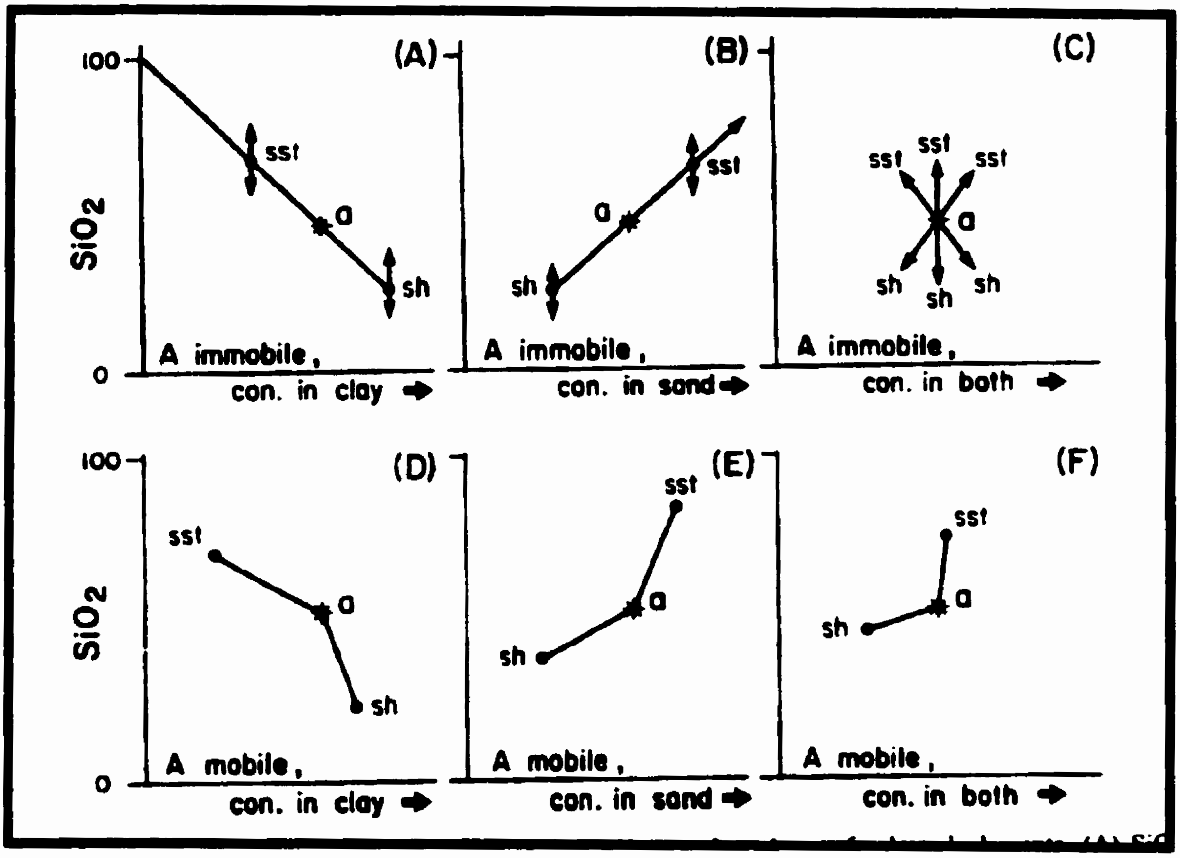


Figure 6.02 Theoretical scattergrams derived by plotting SiO2 concentrations against those of selected elements. Refer to text for explanation. Reproduced from Fralick and Kronberg (1997).

due to constant sum, the ratio of these elements to one another will remain constant if the elements have remained immobile and behaved hydrodynamically similar.

Geochemistry of Metasandstones from the Study Area

Several samples were collected from each of the areas within the Finlayson and Lumby Lake Greenstone Belts for geochemical analysis. Sandstones of a uniform, medium to coarse, grainsize were selected for analysis in order to minimize the effects of hydraulic sorting and allow comparison of the results. Clasts from within conglomerate units, as well as volcanic rock samples from the study area were also analysed. Individual clasts were cut out of conglomerates and only the matrix material from lapilli tuff samples was used.

A total of 42 samples were collected for analysis from throughout the study area. The sixteen samples collected from the Little Falls Lake area, include 12 sandstones, 2 conglomerate clasts and 2 lapilli tuff samples. A total of 13 samples from the Finlayson Lake area, including 11 sandstones, 1 conglomerate clast and 1 mafic volcanic sample, were analysed. Two sandstones were analysed from the Lower Finlayson Lake area, 6 sandstones from the Lower Lumby Lake area, and 5 sandstones from the main Lumby Lake Sedimentary Belt.

All major and trace elements, except SiO_2, were analysed using ICP-AES techniques, by the instrument laboratory at Lakehead University. SiO_2 was analysed using standard XRF techniques.

Data from other sources was also used for comparison with the data acquired for this study. Geochemical analysis of many of the metavolcanic and intrusive rocks within and surrounding the Finlayson and Lumby Lake Greenstone Belts was taken from Stone et al. (1992). Some of these rocks include the Old Tonalite (OT), Tonalite Gneiss (TG), Mafic Tonalite (MT), Steep Rock Lower Mafic unit (SRLM), Steep Rock Upper Felsic unit (SRUF) and the Finlayson Upper Felsic unit (FUF). A total of 9 sandstone samples were collected from the Steep Rock Lake area by Dr. P Fralick and analysed at Lakehead University using similar techniques as for this study.

Analysis of individual mineral grains from carbonates and iron formation samples was conducted using an SEM-EDS.

Little Falls Lake Sedimentary Rocks

Elements commonly thought of as immobile, and which may be useful in a provenance study are AL_2O_3, TiO_2, Zr, Nb, and Y. These elements are plotted against one another in Figure 6.03, to determine if they have been immobile. Zr, Nb, Y, and TiO_2 plot as linear relationships that are near parallel to a radian originating from the origin. This would suggest that they have remained immobile during all alteration processes. Although these elements show linear

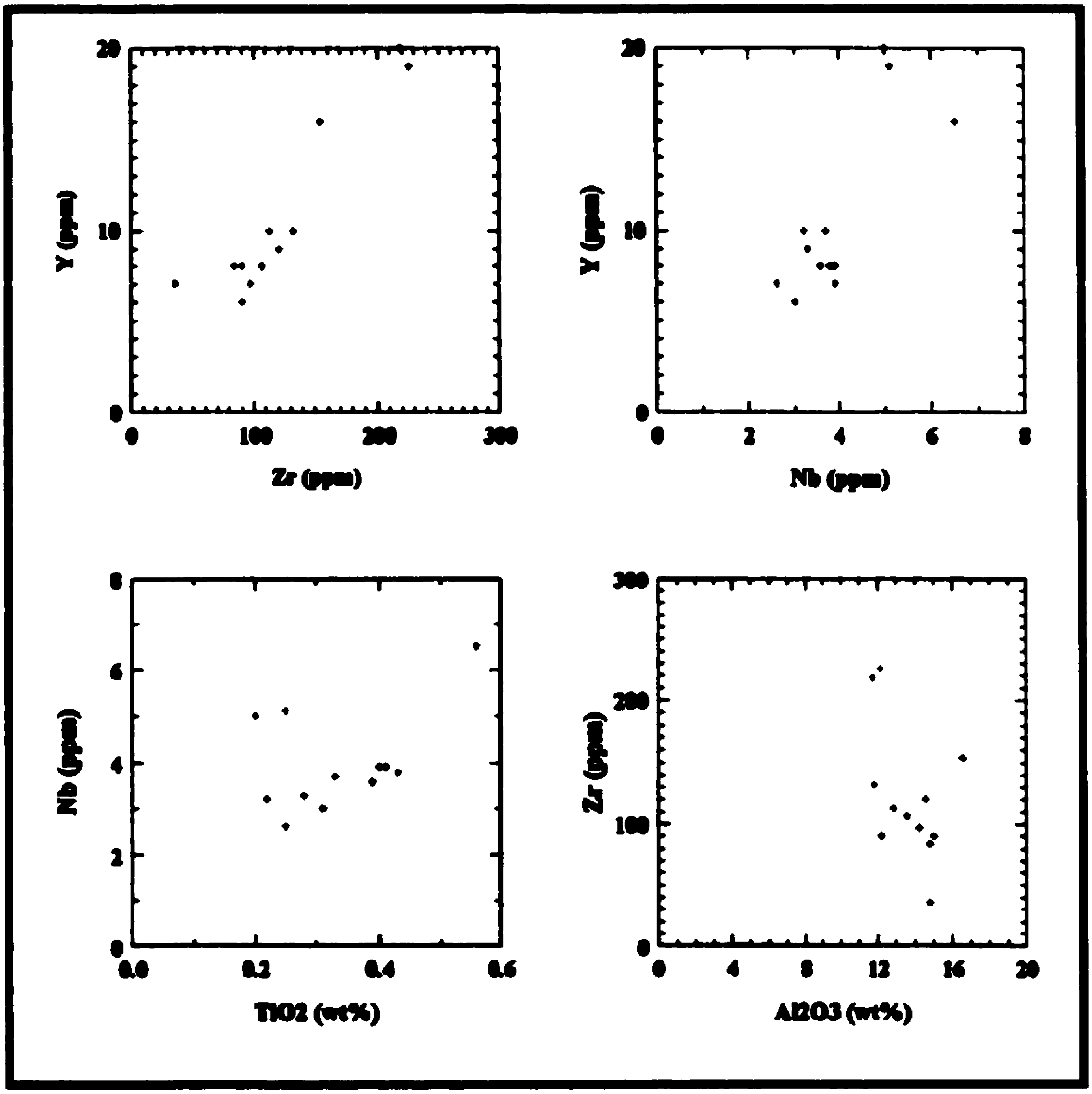


Figure 6.03 Scattergrams of selected element concentrations of samples taken within the Little Falls Lake area.

SYMBOL LEGEND

◆	Norway Lake area	+	Little Falls Lake area
●	Upper Finlayson Lake area	○	Lower Finlayson Lake area
◇	Lower Lumby Lake	◒	Little Falls Conglomerate
◒	Finlayson Lake Conglomerate	◐	Norway Lake Conglomerate
▶	Finlayson Volcanic	◀	Lapilli Tuff – Little Falls Lake
▽	Keewatin Lake core samples	×	Steep Rock Lake area
⬒	Finlayson Lower Mafic Unit	⬓	Finlayson Upper Felsic Unit
◨	Mafic Gneiss	◧	Old Tonalite
■	Steep Rock Lower Mafic Unit	□	Steep Rock Upper Felsic Unit
△	Young Tonalite		

Figure 6.03b Symbol Legend. Refer to Appendix I for geochemical analysis of samples taken for this study and Stone et al. (1992). for analysis of rocks found in the general area.

trends, they are not particularly strong ones. However, Al_2O_3 is considerably more scattered, probably due to it being slightly mobile or its major mineral phases behaving in a hydraulically different manner compared to Zr which it is plotted against. An alternative explanation may stem from the sedimentary rocks themselves. The Little Falls Lake area sedimentary rocks are very tuffaceous in nature, which may cause some variation in their chemistry. If these sedimentary rocks are indeed fairly immature, and have been derived from a volcanic source their composition may vary slightly, due to differences in the chemical composition of successive eruptions, as the volcanic source evolved. This would give rise to different compositions of sediment deposited from successive eruptions.

The hydraulic behaviour of these elements can be determined by plotting them against SiO_2 (Figure 6.04). TiO_2 and especially Al_2O_3, plot with linear trends that extend near the 100% SiO_2 point on the vertical axis. This suggests that they have behaved in a similar manner and have both been concentrated in the clay fraction of the sedimentary rocks. Zr and Y are quite different in that their linear trend does not extend to either the origin or the 100% SiO_2 point on the vertical axis. Both show very similar concentrations in the sands and clays, although there is a slight increase in these elements as the SiO_2 concentration increases, indicating a slight tendency to concentrate within the sand fraction of these sedimentary rocks.

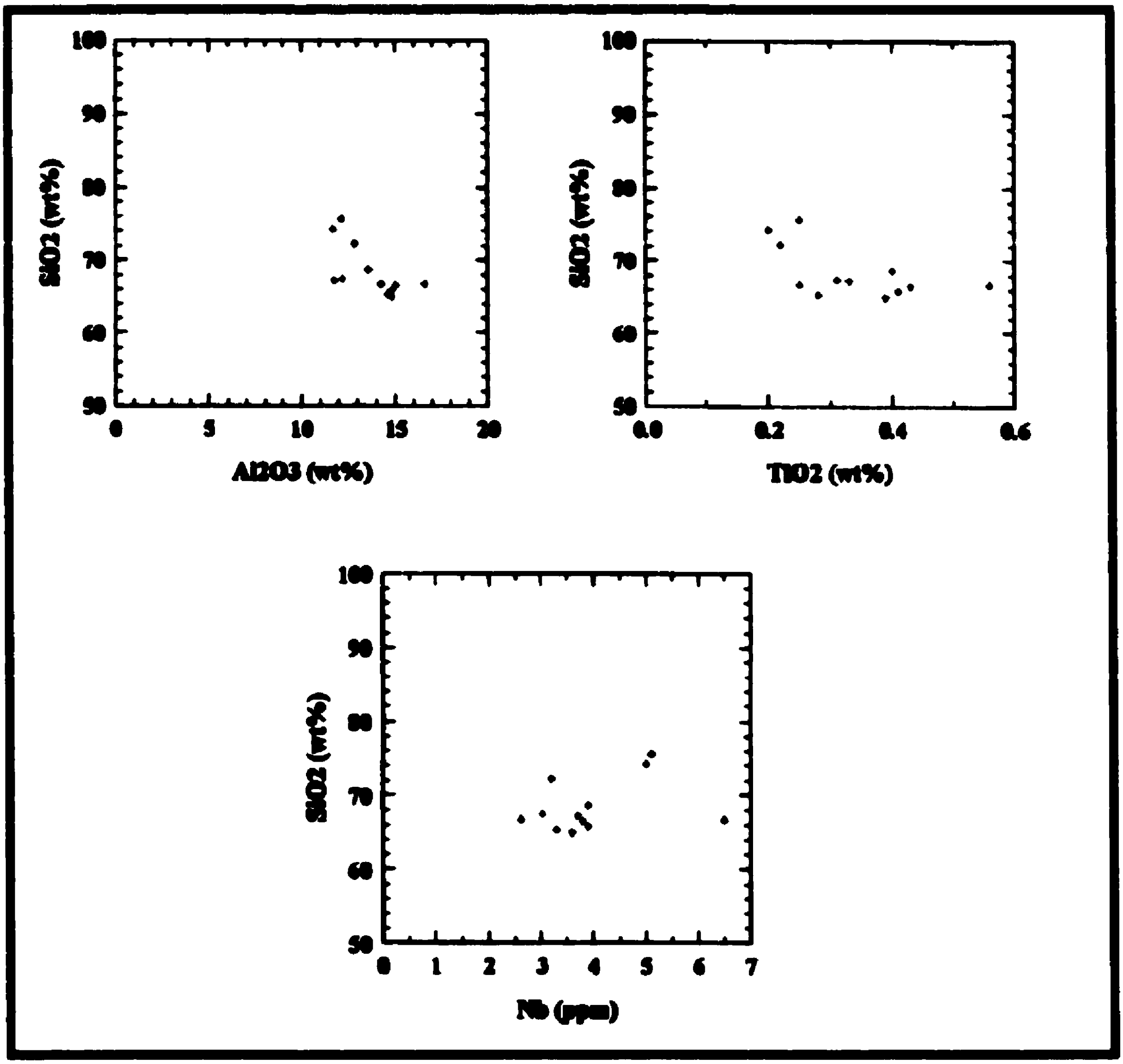


Figure 6.04a Selected elements plotted against SiO_2 for samples taken within The Little Falls Lake area.

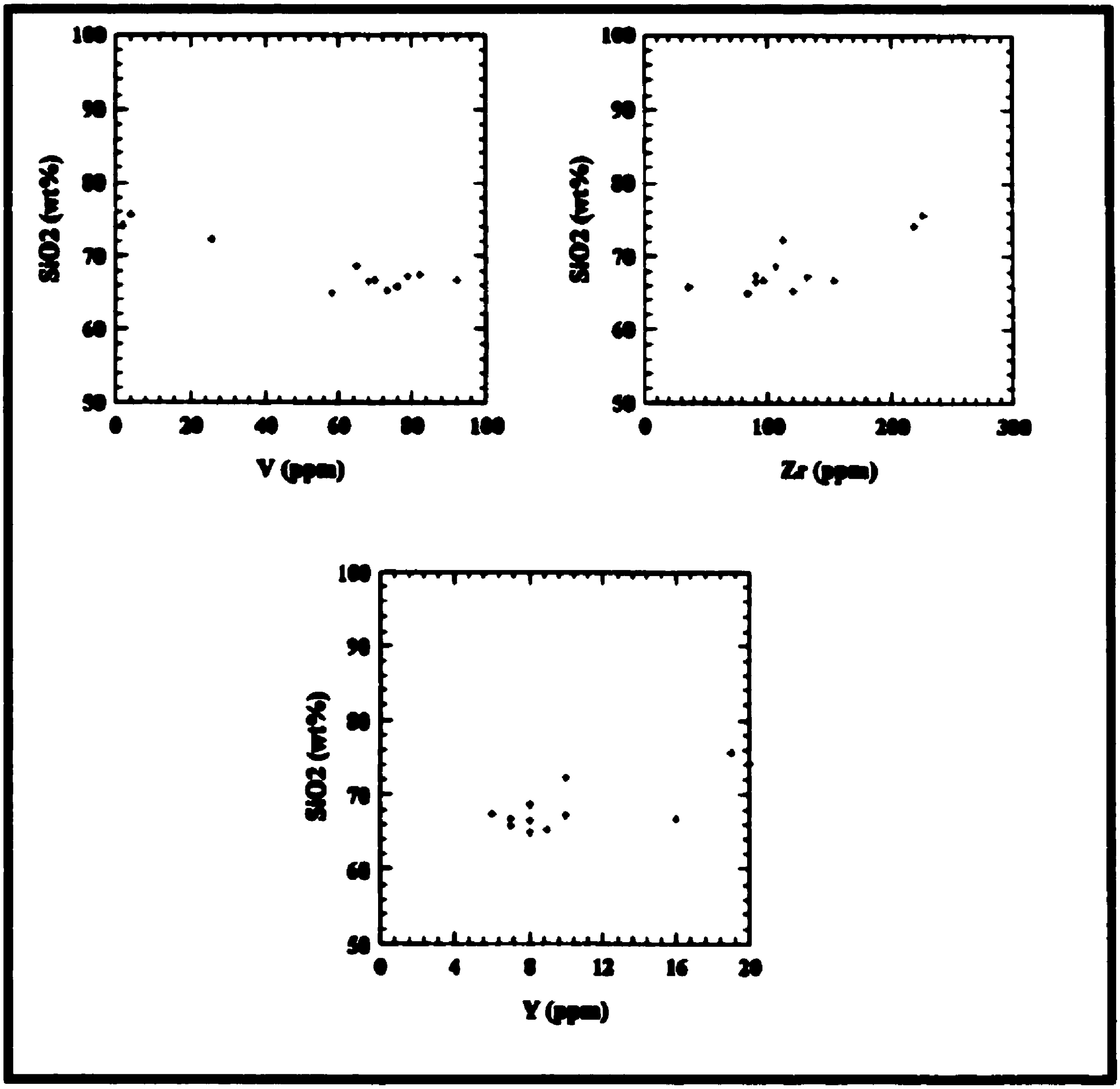


Figure 6.04b Selected elements plotted against SiO2 for samples taken within the Little Falls Lake area.

Plotting ratios of the immobile elements whose major mineral phases have behaved in a hydrodynamically similar manner will group sediments which have similar source rocks and should be useful in identifying the source rocks.

In Figure 6.05 the ratio of TiO_2/Nb is plotted against the TiO_2/Al_2O_3 ratio for the Little Falls Lake area sedimentary rocks. The data plot in a fairly tight grouping with low TiO_2 and Al_2O_3 values. The graph of TiO_2/Al_2O_3 vs. Y/Al_2O_3 shows the Little Falls Lake sedimentary rocks have low TiO_2/Al_2O_3 ratios and moderate to high Y/Al_2O_3 ratios (Figure 6.06). Figure 6.07 shows moderate to high Zr/Al_2O_3 values and low TiO_2/Zr ratios. The data again show a fairly tight grouping with only a couple anomalous samples which have higher Zr concentrations.

Finlayson Lake Area

When the typical elements thought to be immobile in the Finlayson Lake sedimentary rocks are plotted against one another they can be shown to be relatively immobile (Figure 6.08). Al_2O_3 vs. V, TiO_2 vs. Nb, as well as Al_2O_3 vs. Zr show well developed linear trends that are very nearly parallel to radians extending from the origin. Plots with Y are not as well developed suggesting that Y may have been somewhat mobile or its major mineral phases have behaved in a different hydraulic manner than those of other elements. It is interesting to note that the samples taken within the lower Finlayson Lake sedimentary belt tend to plot significantly off the trend expressed by the main group of Finlayson Lake sedimentary rocks from the upper belt. Nb and Y values in these

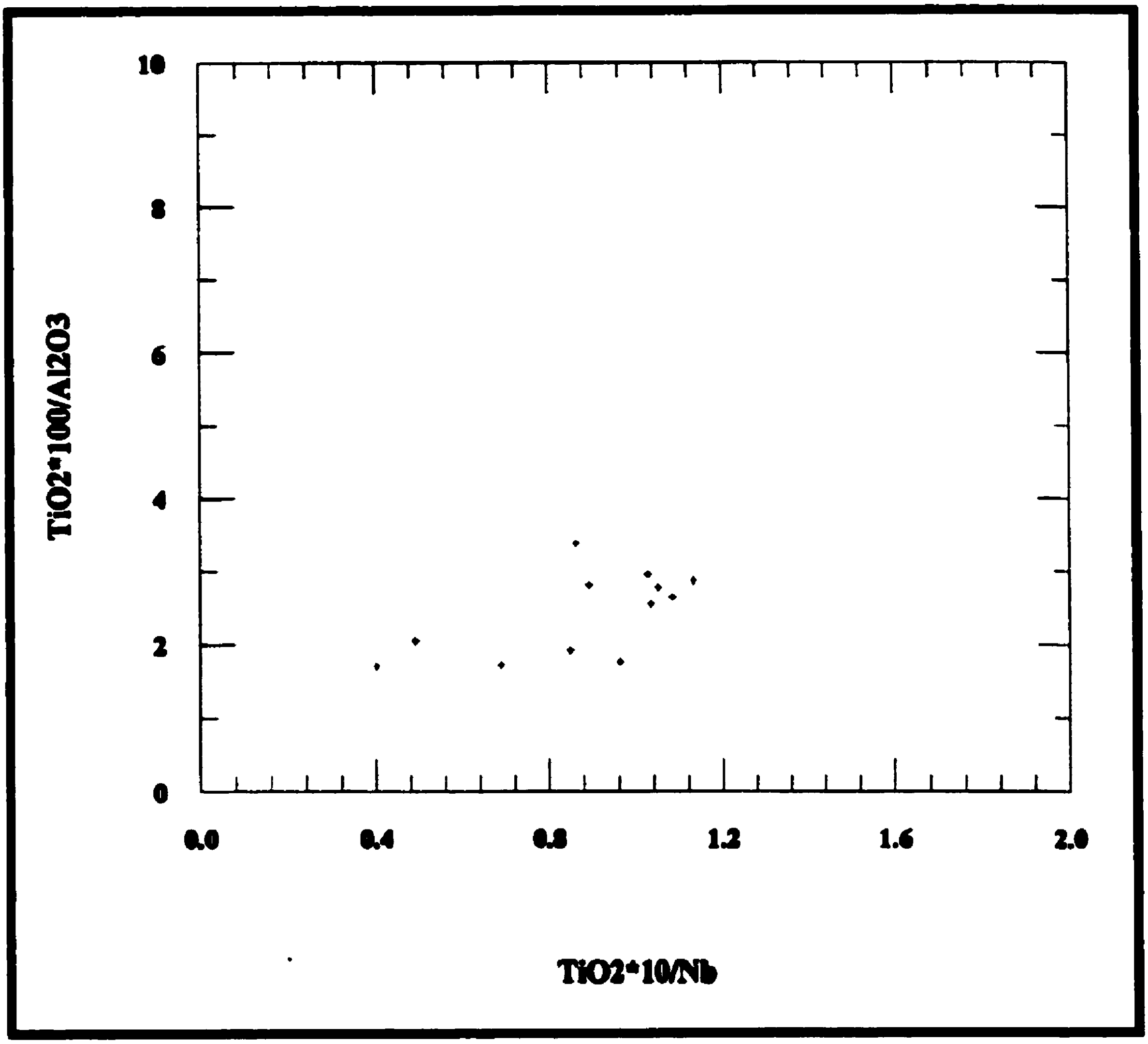


Figure 6.05 Al2O3-TiO2-Nb ratio diagram comparing sandstone samples taken from within the Little Falls Lake area.

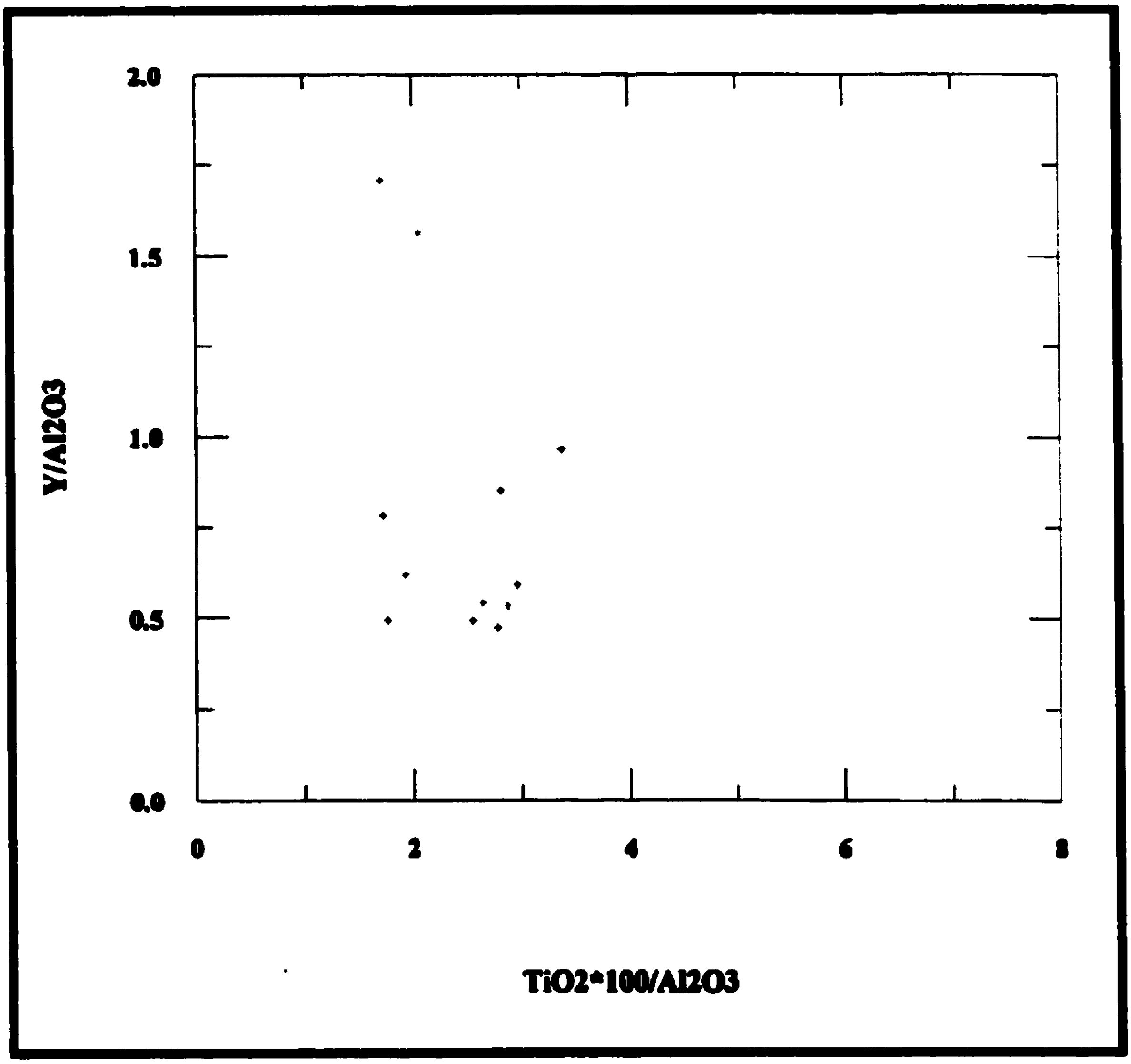


Figure 6.06 TiO2-Al2O3-Y ratio diagram comparing samples taken from within the Little Falls Lake area.

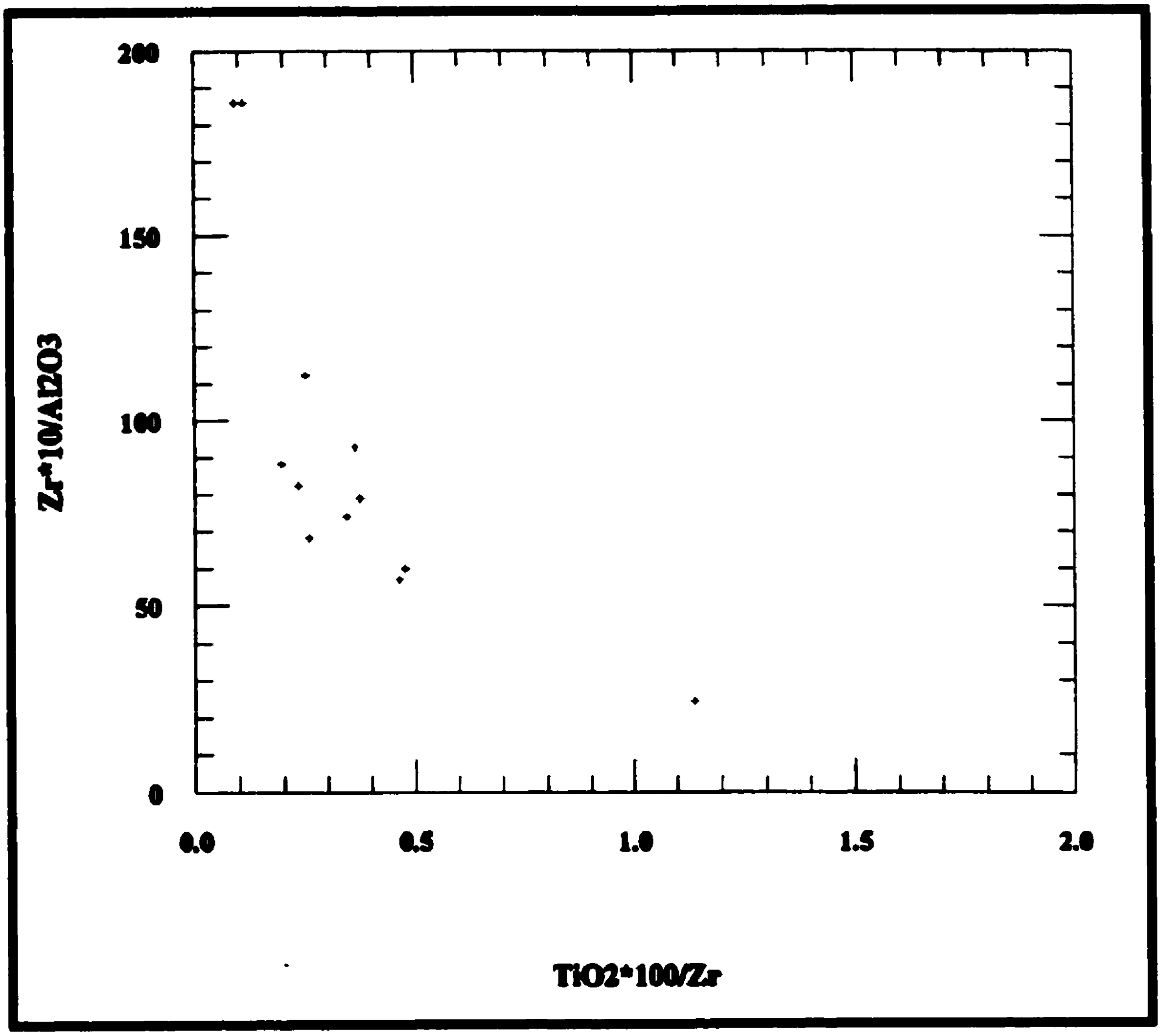


Figure 6.07 TiO2-AlO2-Zr ratio diagram for comparing samples taken from within the Little Falls Lake area.

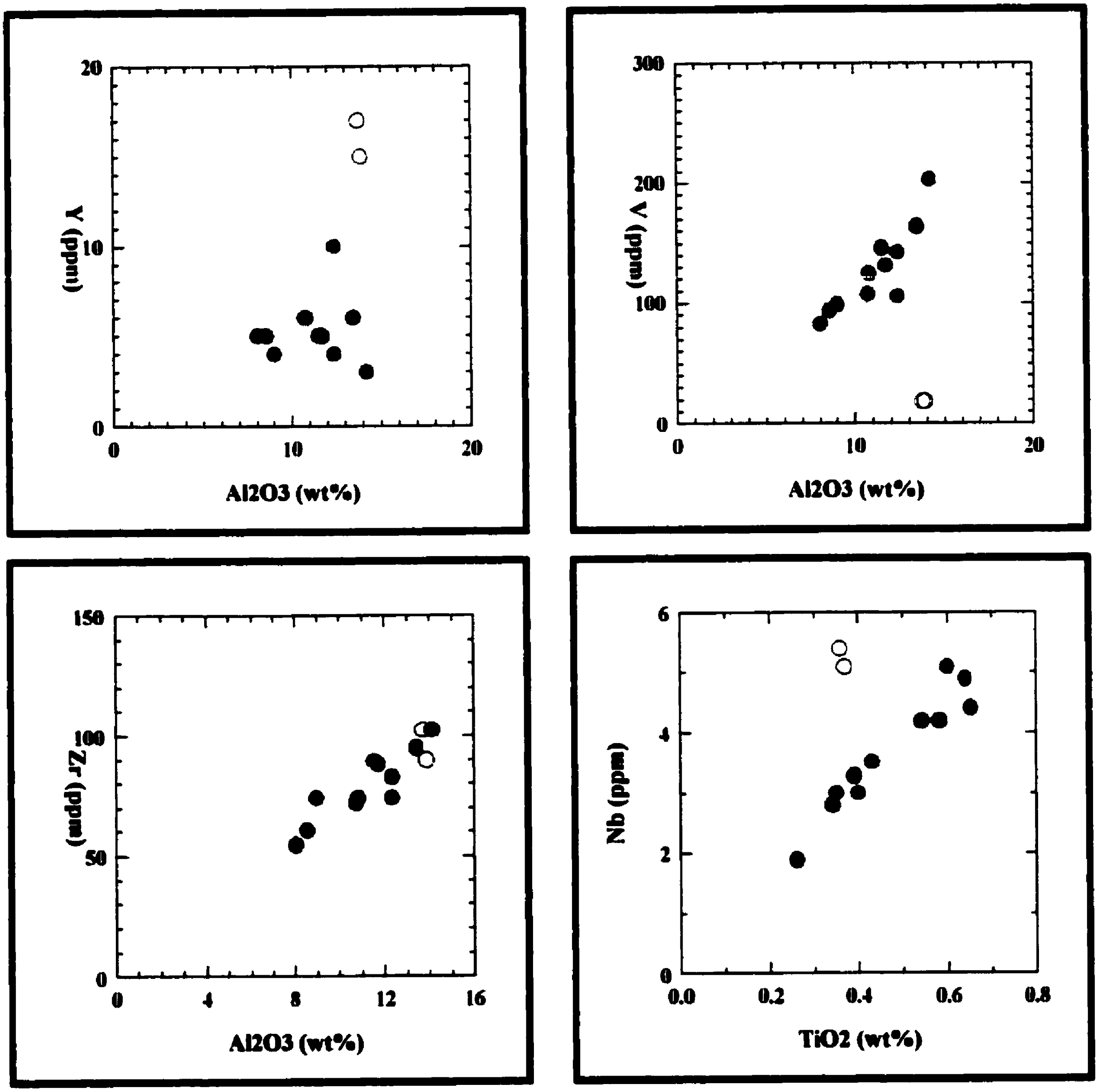


Figure 6.08a Scattergrams for selected elements from samples taken from within the Finlayson Lake area. Open circles from the Lower Finlayson Lake area, closed circles from the Upper Finlayson Lake area.

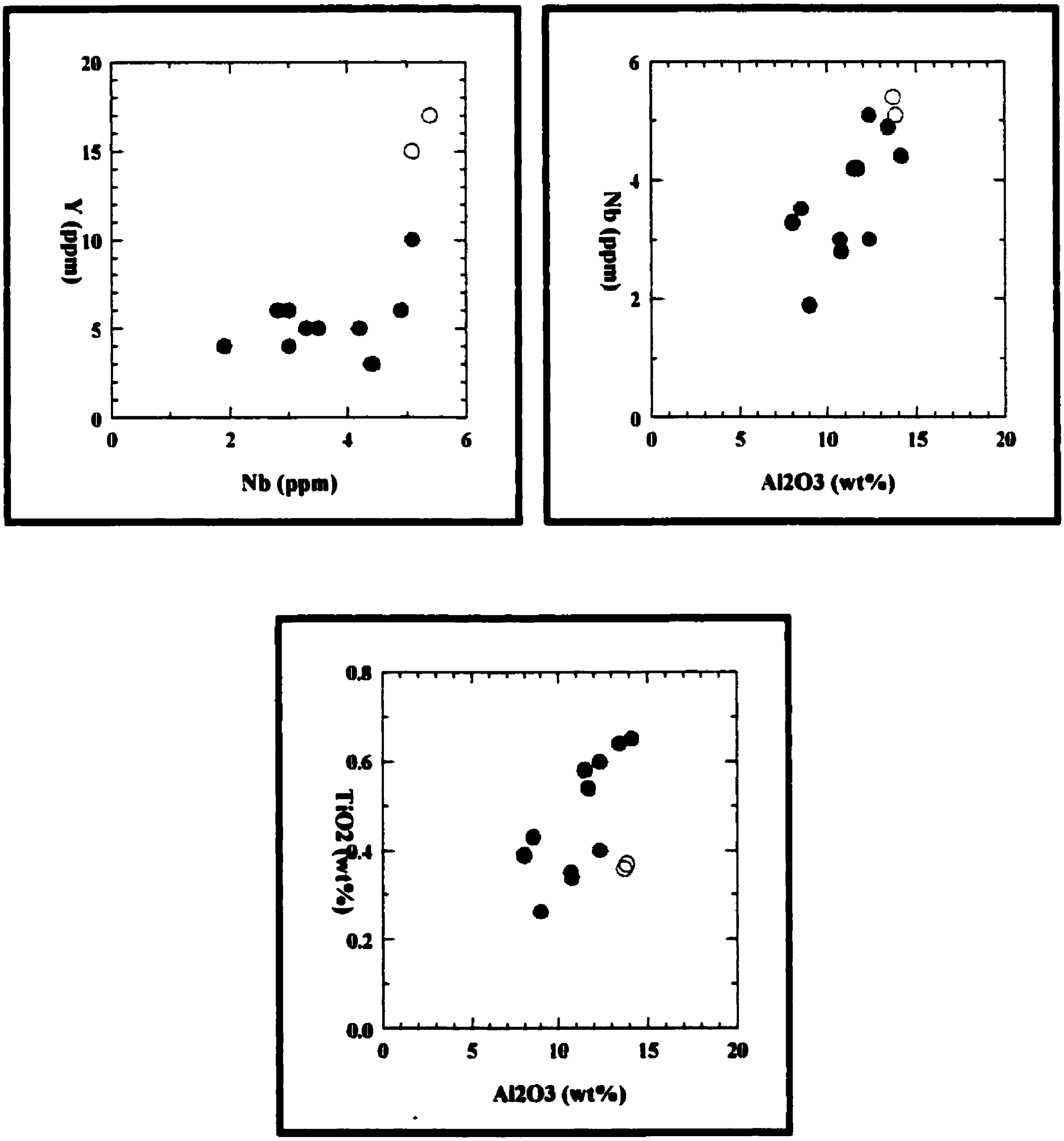


Figure 6.08b Scattergrams for selected elements from samples taken from within the Finlayson Lake area. Open circles from the Lower Finlayson Lake area, closed circles from the Upper Finlayson Lake area.

sedimentary rocks are enriched, and V is depleted as compared to the upper Finlayson sedimentary belt.

The elements can also be plotted against SiO_2 in order to determine their hydraulic behaviour. Zr, Al_2O_3, V, and to a lesser extent TiO_2 show fairly well developed linear trends that extend very near the 100% SiO_2 point on the vertical axis (Figure 6.09). This suggests that the mineral phases which contain these elements have been concentrated in the clay fraction of the sedimentary rocks, with their concentration decreasing with increasing SiO_2 content. Nb plots in a linear trend that does not extend to either the origin or the 100% SiO_2 point, suggesting Nb values are similar in both the sandstone and clay fractions. Concentrations of Nb are slightly increased in the clay fraction but not significantly. This may suggest that Nb exists as both large and small grains in similar amounts. Y shows a clustering of data points with no preferred trend with varying SiO_2 content. This suggests minimal hydrodynamic fractionation of the main mineral phases containing Y. Again it is significant to note that the lower Finlayson Lake sedimentary rocks plot away from the trend expressed by the main group of sedimentary rocks.

Finlayson Lake sediment data have high TiO_2/Nb values and show two groupings of TiO_2/Al_2O_3 ratios (Figure 6.10). This graph suggests that samples are separating due to differences in TiO_2 content relative to Al_2O_3 and Nb, indicating differences in the amount of mafic/felsic detritus within them. The

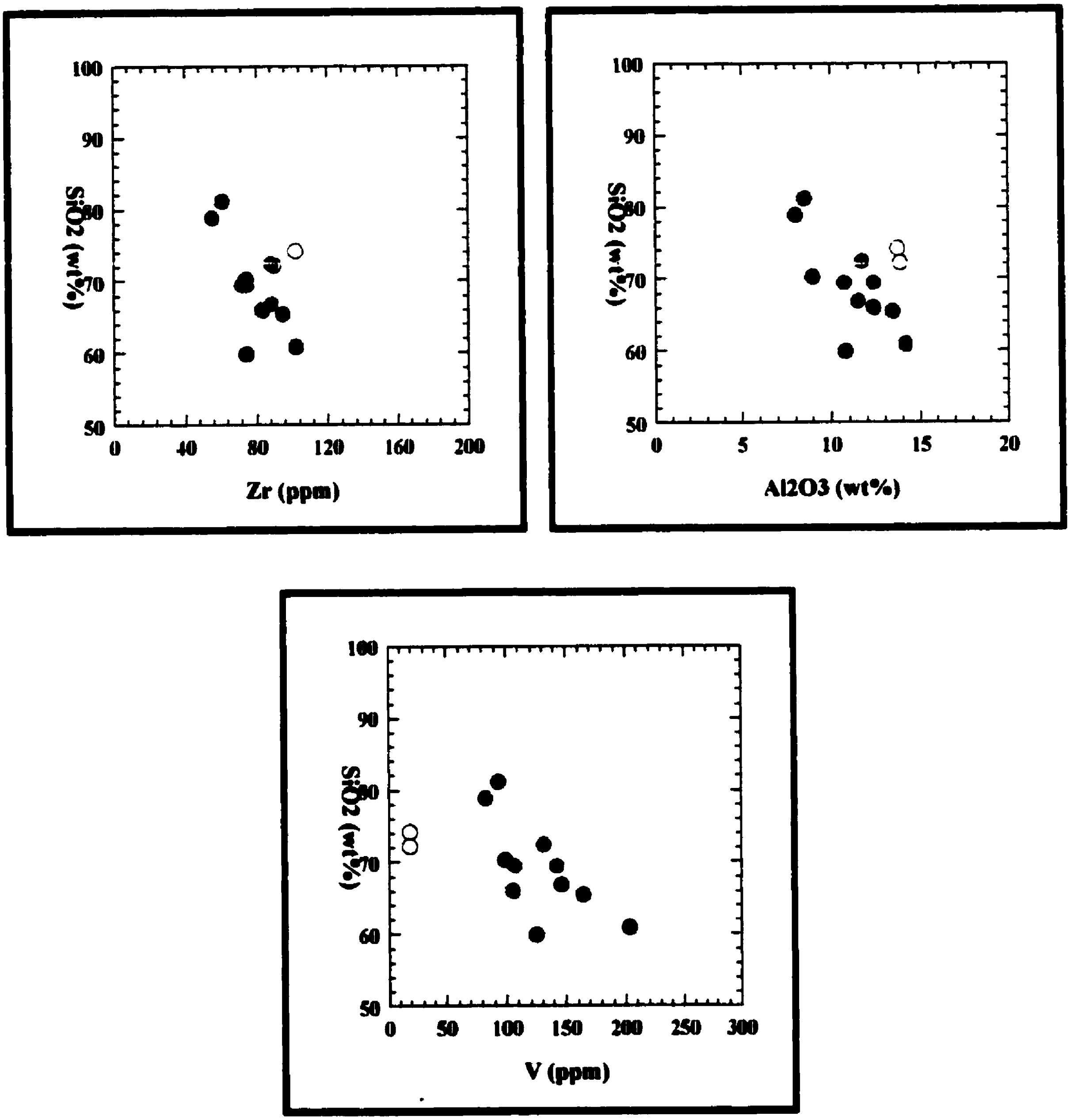


Figure 6.09a Selected elements plotted against SiO_2 for samples taken from within the Finlayson Lake area. Open circles from the Lower Finlayson Lake area, closed circles from the Upper Finlayson Lake area.

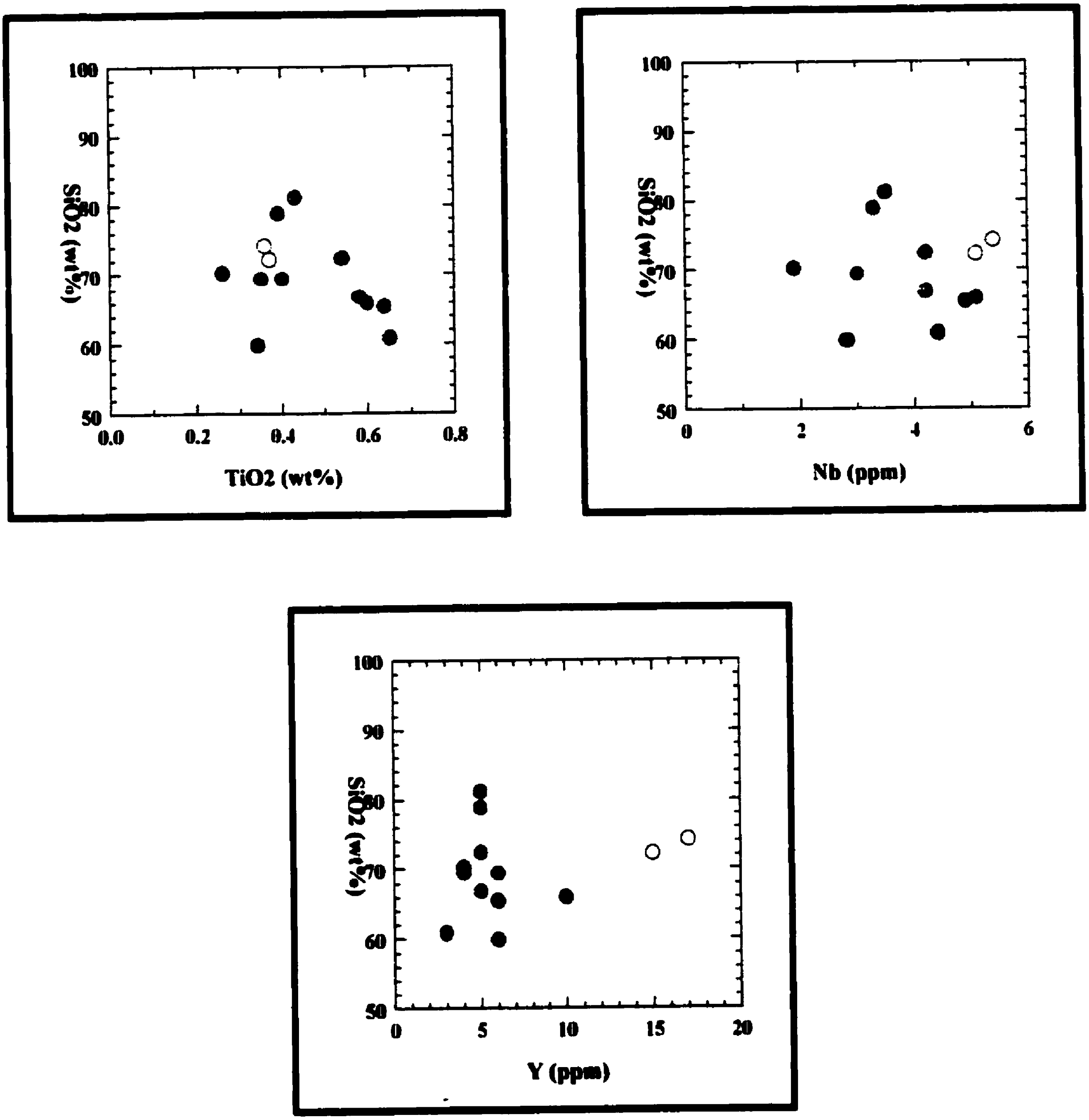


Figure 6.09b Selected elements plotted against SiO_2 for samples taken from within the Finlayson Lake area. Open circles from the Lower Finlayson Lake area, closed circles from the Upper Finlayson Lake area.

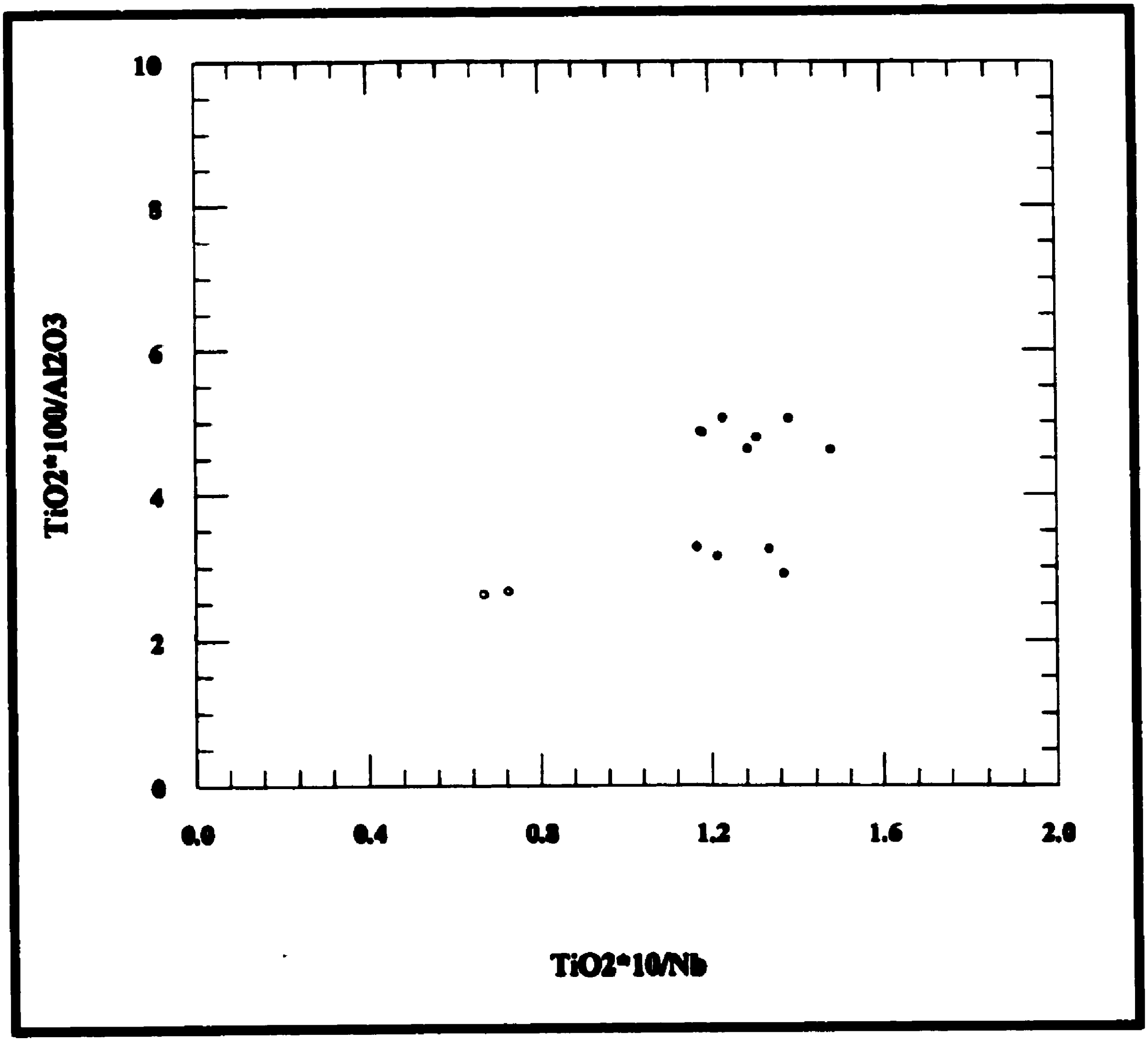


Figure 6.10 TiO2-Al2O3-Nb ratio diagram for samples taken from within the Finlayson Lake area. Open circles from the lower Finlayson Lake area, closed circles from the upper Finlayson Lake area.

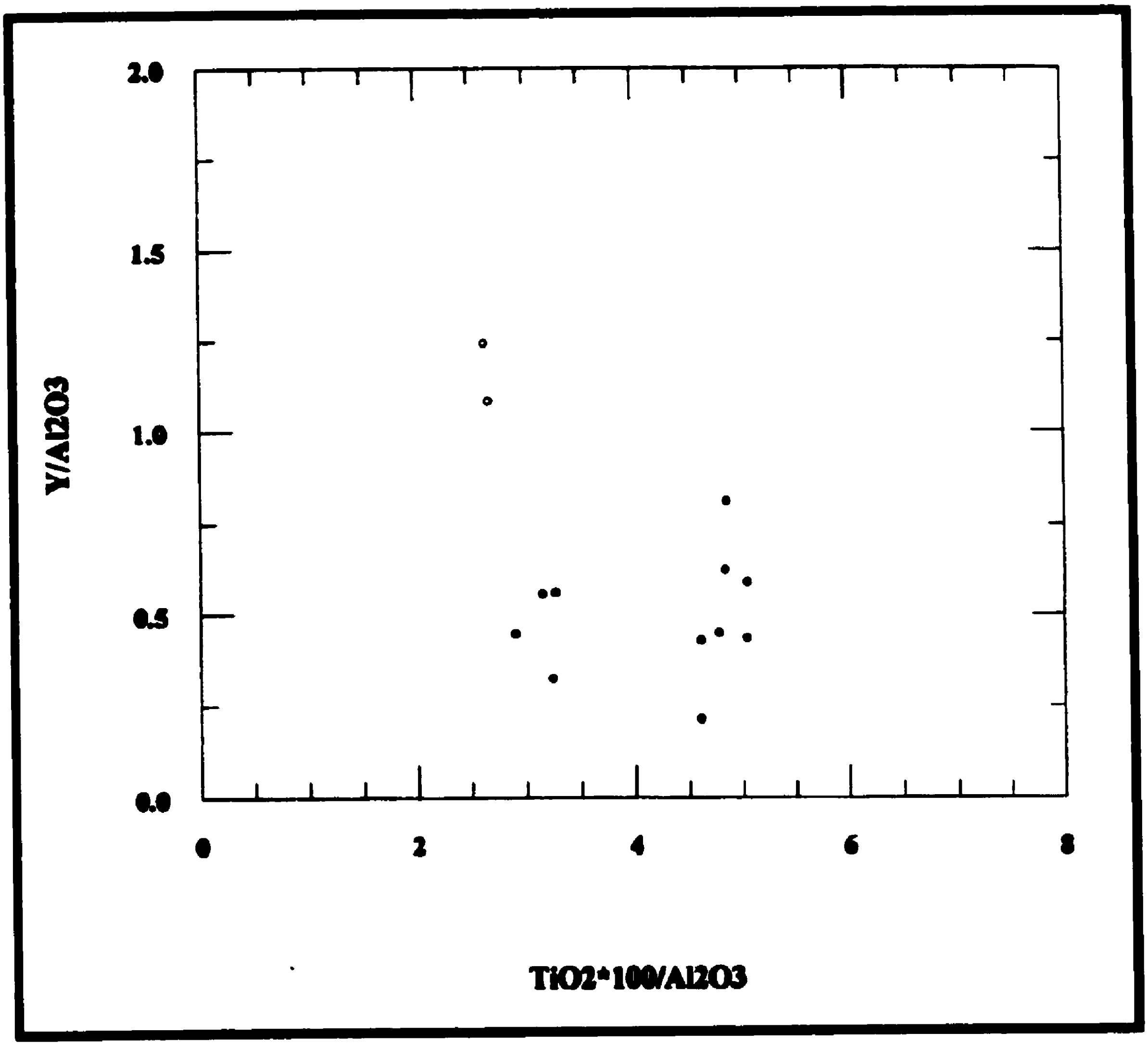


Figure 6.11 TiO2-Al2O3-Y ratio diagram for samples collected from within the Finlayson Lake area. Open circles from the lower Finlayson Lake area, closed circles from the upper Finlayson Lake area.

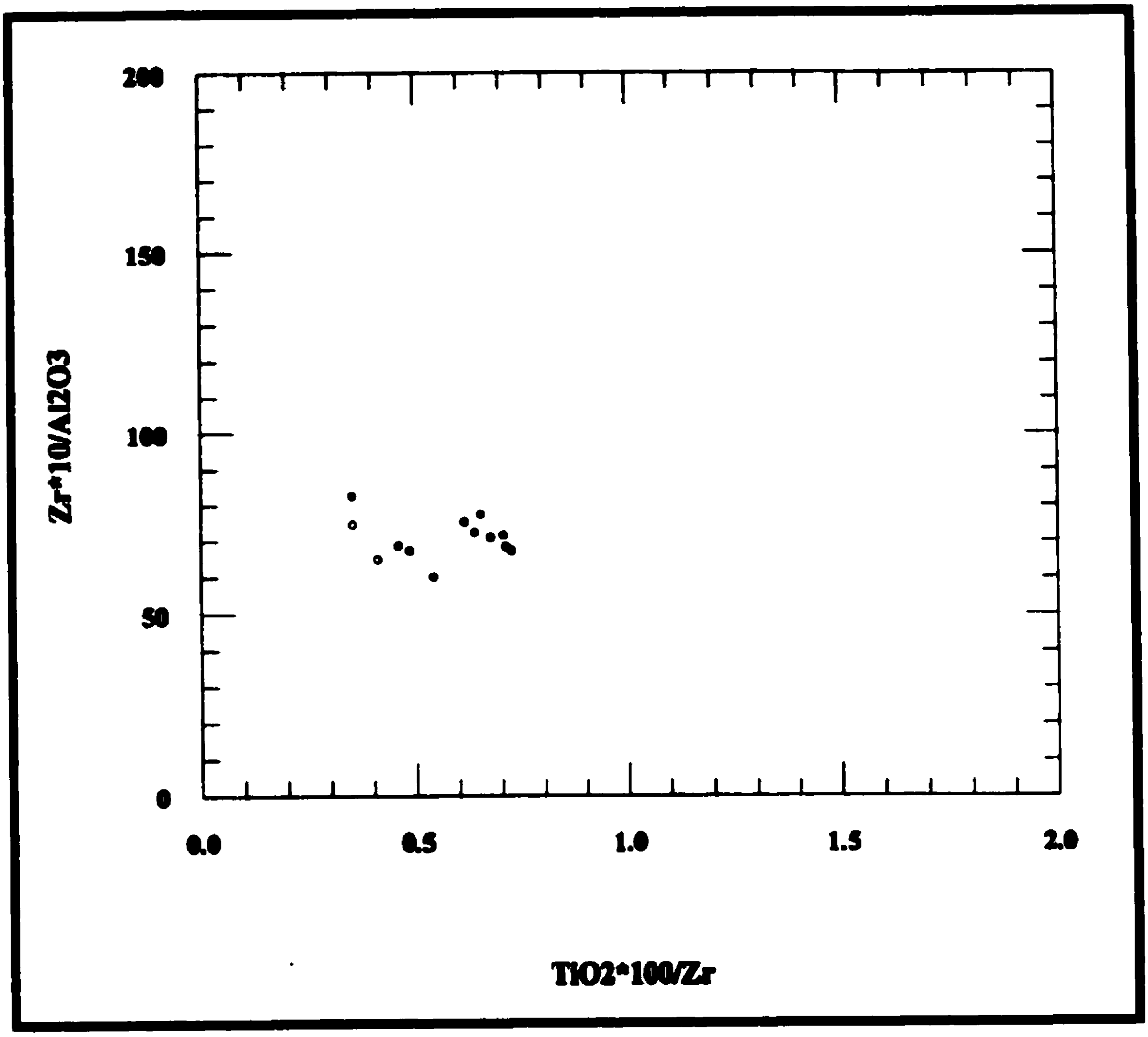


Figure 6.12 TiO_2-Al_2O_3-Zr ratio diagram for samples collected from within the Finlayson Lake area..Open circles from the lower Finlayson Lake area, closed circles from the upper Finlayson Lake area.

double grouping with respect to TiO_2/Al_2O_3 ratios is again reflected in the plot of TiO_2/Al_2O_3 vs. Y/Al_2O_3 (Figure 6.11). The data from the lower sequence of sedimentary rocks in the Finlayson Lake area plot well away from the main group of samples. A tight cluster is seen in Figure 6.12, which reflects higher TiO_2 values and lower Zr values in these sedimentary rocks when compared with the Little Falls data.

Lumby Lake Area

The Lumby Lake sedimentary rocks consist of samples taken from three separate areas. If the geochemical data from these three areas are looked at separately it tends to show that the typical elements have remained immobile in each of the data sets (Figure 6.13). The graph of Nb vs. Y shows a very well developed linear trend extending to the origin for the lower Lumby Lake and the Hook Lake data (Figure 6.13). Norway Lake data was determined for only two samples and it is therefore difficult to determine the trend which may be expressed with such a small data set. Nb vs. TiO_2 also shows a well developed trend for the lower Lumby and Hook Lake sedimentary rocks (Figure 6.13). Zr vs. Al_2O_3 shows increased Zr values relative to increasing Al_2O_3 values and may reflect some mobility of Zr or, more likely, differing hydraulic behaviour (Figure

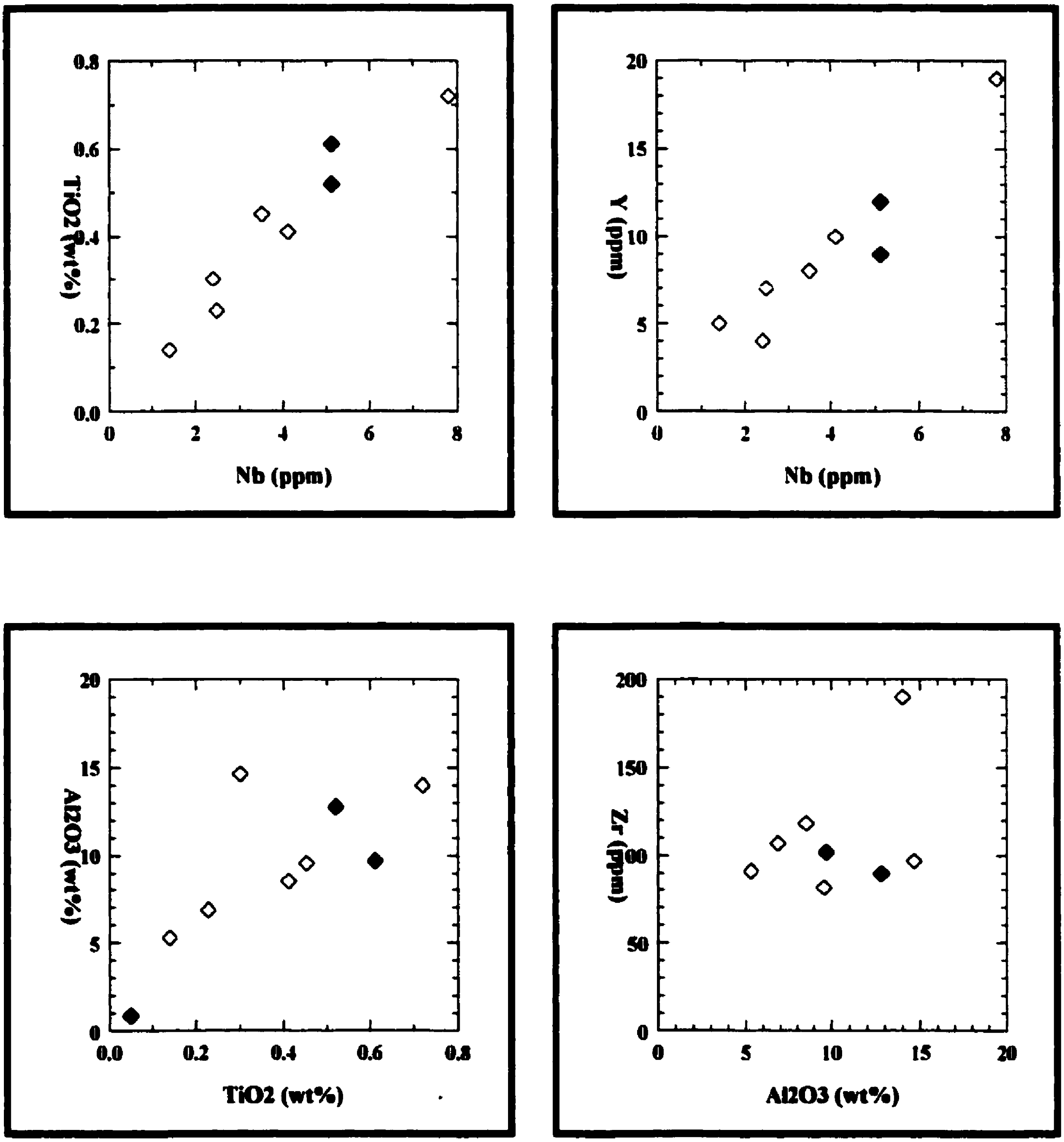


Figure 6.13 Scattergrams of selected elements for samples taken from within the Lumby Lake area. Open diamonds from the Hook Lake area, closed diamonds from the Norway Lake area.

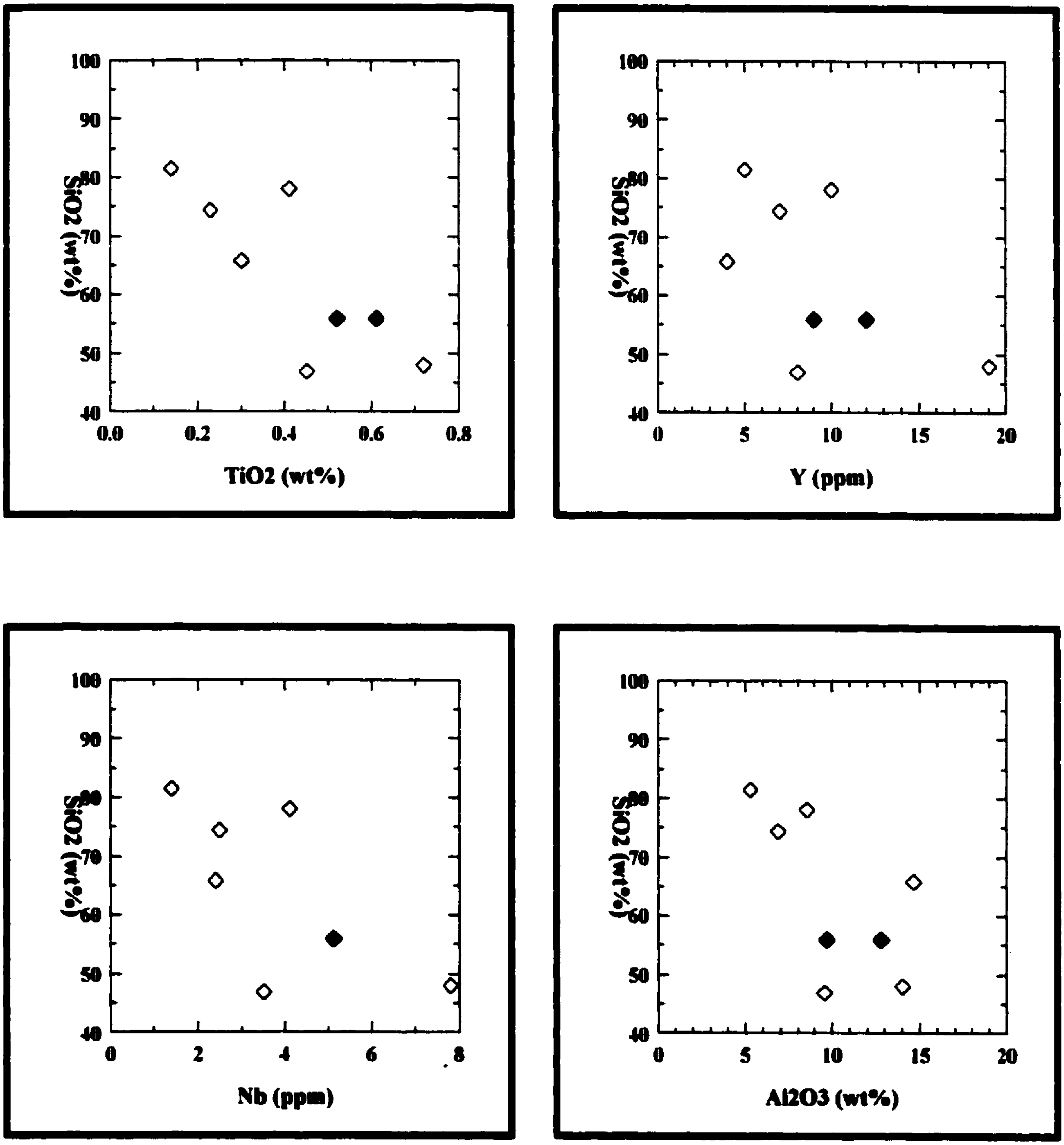


Figure 6.14 Selected elements plotted against SiO_2 for samples taken from within the Lumby Lake area. Open diamonds from the Hook Lake area, closed diamonds from the Norway Lake area.

6.13). TiO_2 vs. Al_2O_3 shows increased Al_2O_3 values with similar increases in TiO_2, trending near but not directly to the origin (Figure 6.13). These graphs suggest that Nb, Y, and TiO_2 have remained immobile and their major mineral phases have behaved in a similar hydrodynamic manner. Al_2O_3 and Zr have either been mobile or their major mineral phases have behaved in a dissimilar hydraulic manner.

Figure 6.14 suggest that Al_2O_3, TiO_2, Y, and Nb have behaved in a similar hydraulic manner, concentrating within the clay fraction of the sediment, in the lower Lumby Lake sedimentary belt. All show a correlation with SiO_2,which extends near the 100% SiO_2 point on the vertical axis. Data sets are too small for the other areas to determine their behaviour accurately.

Sediment samples taken in the Lumby Lake area show significantly more variability in their concentrations of immobile elements as compared to the other areas. These data show intermediate Ti/Nb and TiO_2/Al_2O_3 ratios (Figure 6.15). Figure 6.16 shows high variability in the data indicating higher Y/Al_2O_3 and higher TiO_2/Al_2O_3 ratios. Low TiO_2/Zr and high Zr/Al_2O_3 ratios are visible in Figure 6.17. This is probably caused by the higher Zr values in these samples.

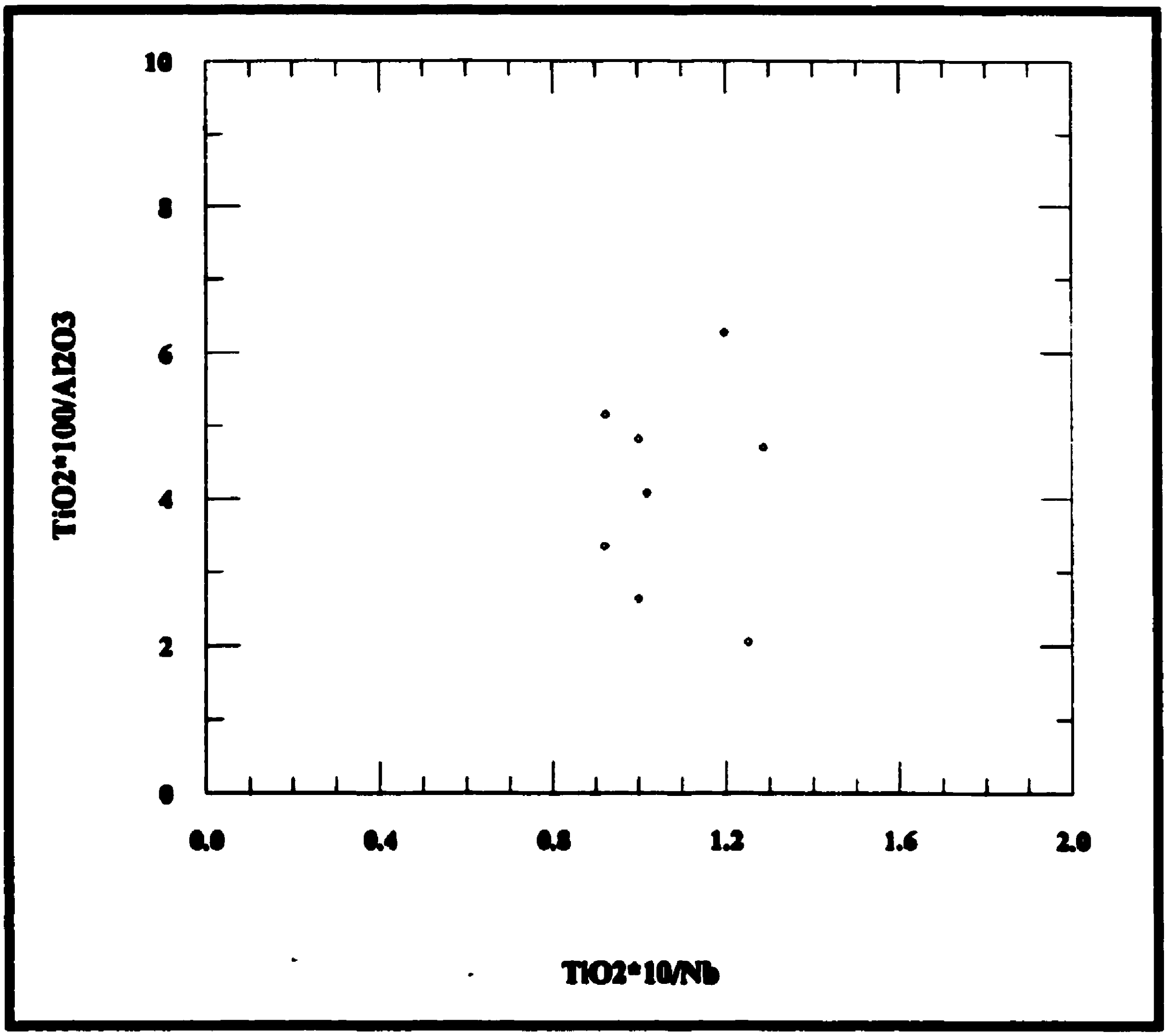


Figure 6.15 TiO2-Al2O3-Nb ratio diagram for samples collected within the Lumby Lake area. Open diamonds from the Hook Lake area, closed diamonds from the Norway Lake area.

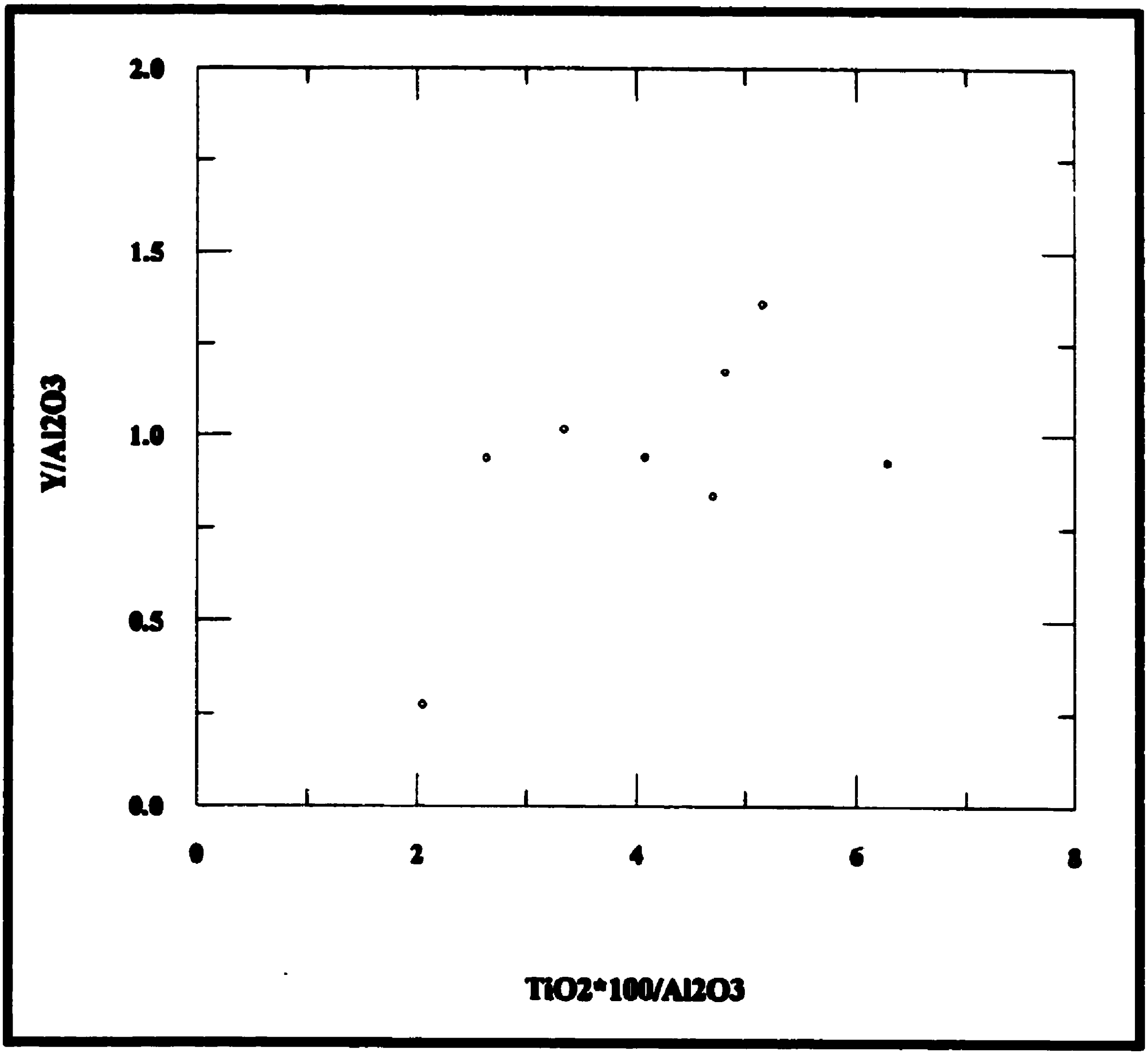


Figure 6.16 TiO2-Al2O3-Y ratio diagram for samples collected within the Lumby Lake area. Open diamonds from the Hook Lake area, closed diamonds from the Norway Lake area.

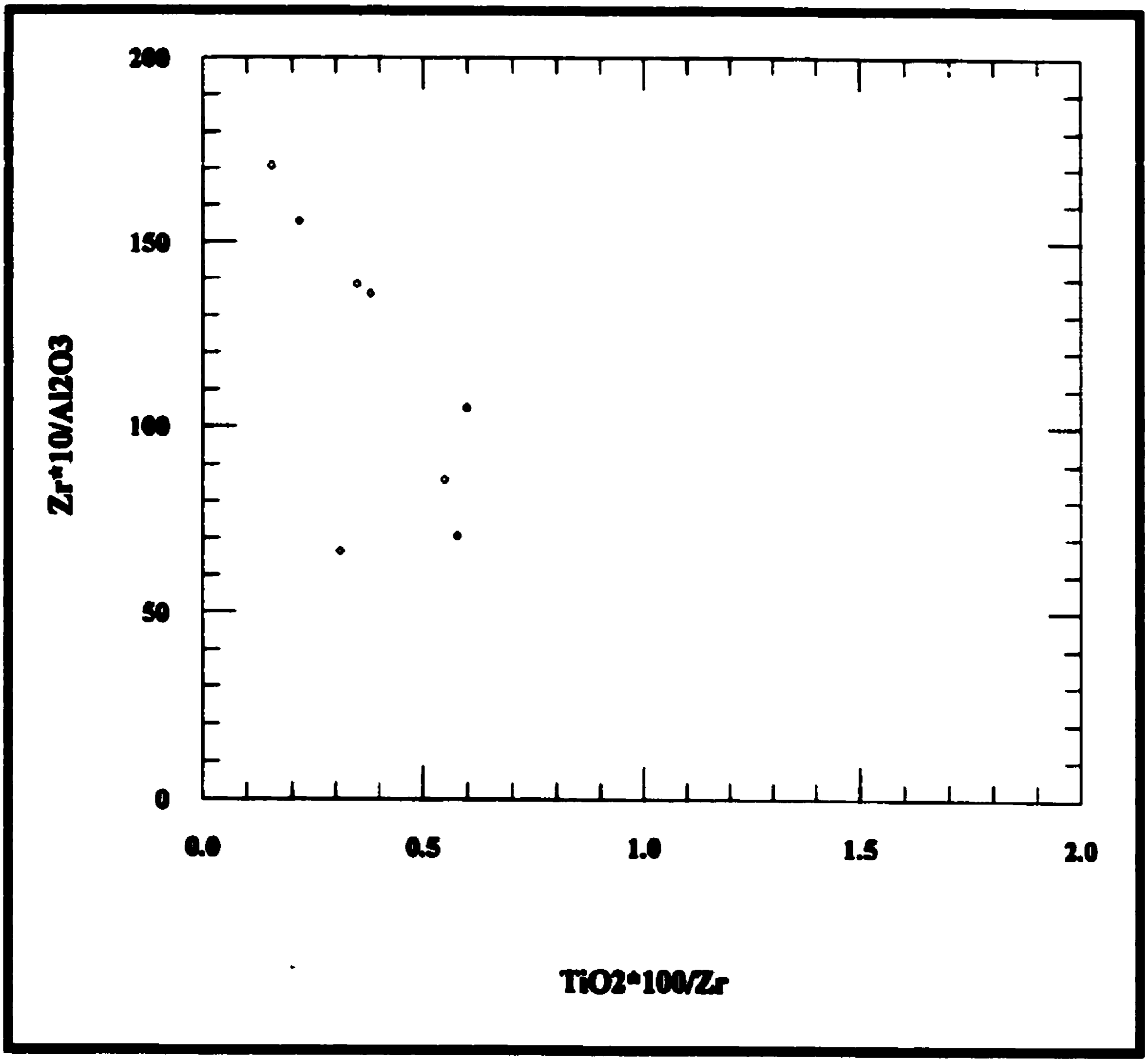


Figure 6.17 TiO2-Al2O3-Zr ratio diagram for samples collected within the Lumby Lake area. Open diamonds from the Hook Lake area, closed diamonds from the Norway Lake area.

Discussion of Sediment Geochemistry

In most areas TiO_2, Nb, V and Al_2O_3 have been shown to be relatively immobile and have behaved in a similar manner being concentrated in the clay fraction of the sedimentary rocks. Zr was usually immobile but does not seem to be partitioned into any particular size fraction; it is only slightly more abundant in the sand fraction of the sedimentary rocks. Y, in some cases, was slightly mobile, and was concentrated in the clay fraction of the sedimentary rocks in all areas.

A better understanding of the sedimentary rocks may come from plotting all sedimentary rocks on a single graph along with some possible source rocks in order to identify which rocks the different sedimentary rocks have been derived from. The graph of TiO_2/Nb vs. TiO_2/Al_2O_3 should prove to be the most promising because all elements used have remained immobile and their major mineral phases have behaved similarly during transport. The ratios of these elements in the sedimentary rocks should remain very similar to the ratios in the source rocks. Using ratios alleviates any differences that are due to mass balance effects as mobile elements are lost from or gained by the system. The sedimentary rocks from the three different areas plot in slightly different fields (Figure 6.18). Because TiO_2 and Nb are fairly compatible and are found together in some mafic mineral phases, a felsic to mafic trend is indicated only in the

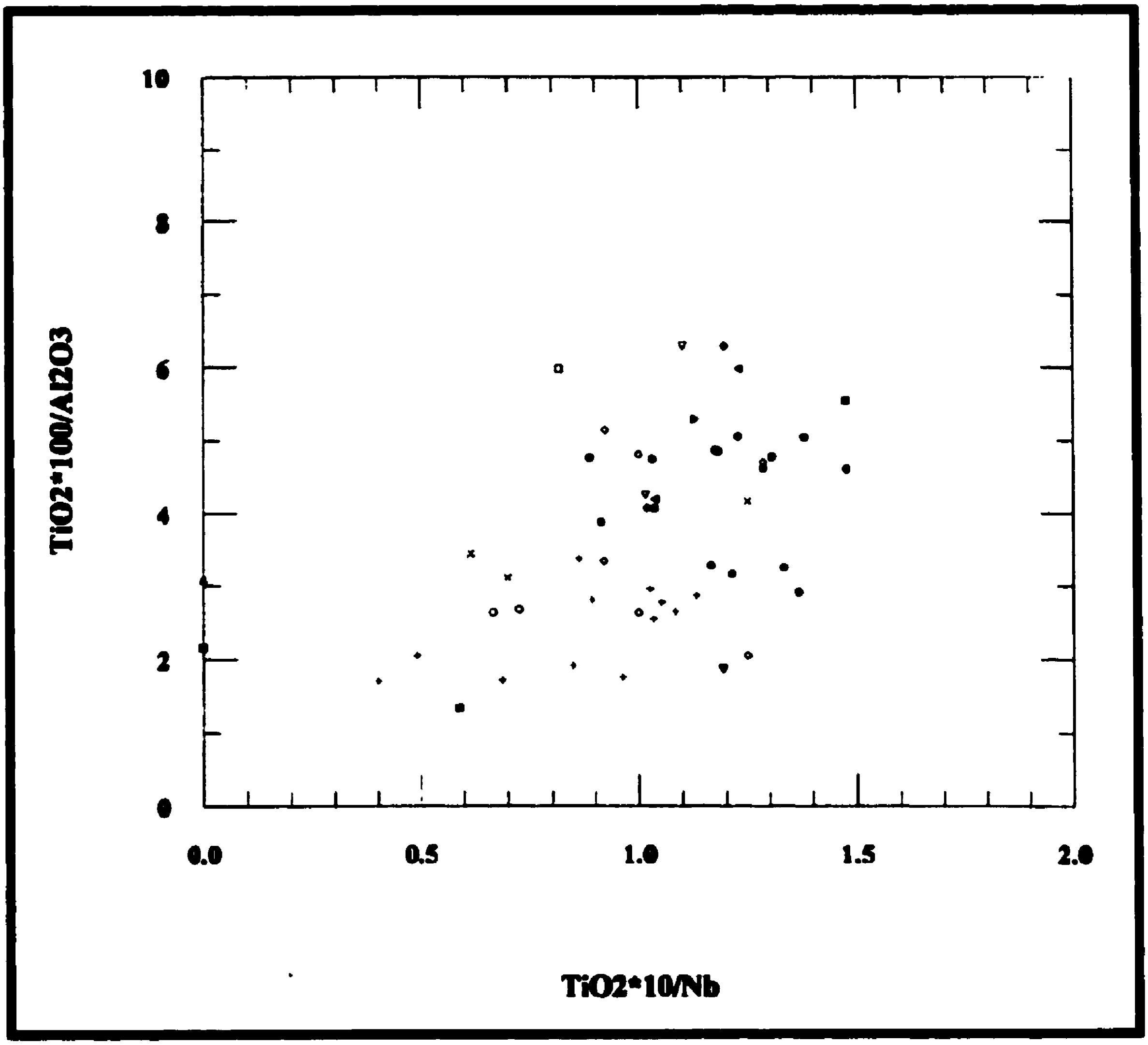


Figure 6.18 TiO2-Al2O3-Nb ratio diagram for all samples collected within the Finlayson and Lumby Lake Belts Greenstone Belts. Refer to Figure 6.03b for symbol legend.

vertical direction as the TiO_2/Al_2O_3 ratio increases. The Little Falls Lake sedimentary rocks plot in their own field and have the lowest of these two ratios, indicating a felsic source. The main group of Finlayson Lake sedimentary rocks and the sedimentary rocks from Norway Lake and some of the samples from Hook Lake in the Lumby Lake belt have higher ratios and indicate a more mafic contribution. The sediment samples taken in the lower Lumby Lake sequence and the lower Finlayson Lake sedimentary rocks are associated with the Little Falls Lake sedimentary rocks on this graph rather than the Finlayson Lake sedimentary rocks. One sediment sample taken from the steep rock group plots with the Finlayson Lake sedimentary rocks indicating it had a mafic component in its source.

Some of the potential source rocks in the area are also plotted on this graph. Lapilli tuff samples taken from Little Falls Lake and the volcanic sample from Finlayson Lake plot with high ratios in a similar position indicating the possibility of a similar magma source. These samples plot in the upper right hand region of the graph, in a position near the composition of the Steep Rock Lower Mafic unit of Stone et al. (1992). The Old Tonalite (Stone et al., 1992) has a very felsic signature, with both of the ratios low. The Steep Rock Upper Felsic unit (Stone et al., 1992) plots with a low TiO_2/Nb ratio similar to the Old Tonalite but has a much higher TiO_2/Al_2O_3 ratio. The Little Falls Lake sediments, the lower Finlayson sediments, and the lower Lumby lake sediments all have a fairly felsic signature, plotting in the vicinity of the Old Tonalite. The Finlayson Lake main group and the Norway Lake sedimentary rocks plot closer to the Steep Rock Lower Mafic unit. This indicates that the Little Falls Lake sedimentary rocks and the lower sequences had a source similar to the Old Tonalite, while the main Finlayson Lake and Norway Lake sedimentary rocks probably had contributions from the Old Tonalite and the Steep Rock Lower Mafic unit or another mafic volcanic unit. The samples taken from the Steep Rock Lake sequence are variable, with some plotting near the Little Falls Lake sedimentary rocks and some within the Finlayson Lake sedimentary rocks.

Although Zr has not been concentrated in the clay fraction of the sedimentary rocks and the other immobile elements have been, ratio plots including it can provide useful information. Just as in Figure 6.18, there is a well developed felsic to mafic trend expressed on the plot of TiO_2/Zr vs. Zr/Al_2O_3 when samples from all areas are plotted (Figure 6.19). The more felsic sedimentary rocks have higher Zr/Al_2O_3 ratios and lower TiO_2/Zr ratios. A segregation of the sedimentary rocks from the different locations similar to that of Figure 6.18 occurs. The main group of Finlayson Lake and the Norway Lake sedimentary rocks have higher TiO_2/Zr ratios than the other sedimentary rocks suggesting that they have a more mafic source. In this plot the samples taken from the lower sequence of Finlayson Lake sedimentary rocks are again affiliated with the cluster of Little Falls Lake sedimentary rocks. These two groups have lower TiO_2 values than the Finlayson Lake sedimentary rocks suggesting a more felsic

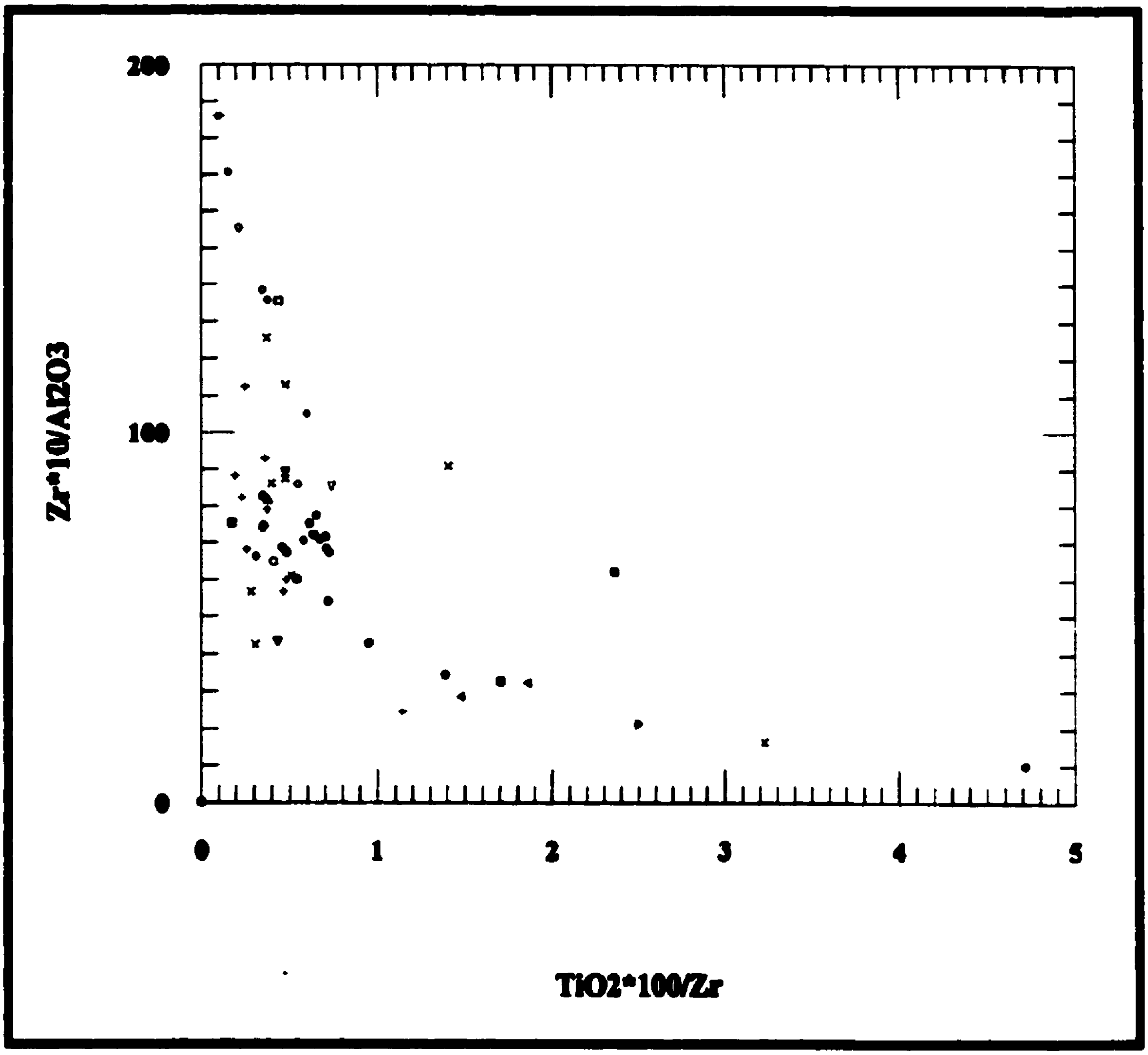


Figure 6.19 TiO2-Al2O3-Zr ratio diagram for all samples collected within the Finlayson and Lumby Lake Greenstone Belts. Refer to Figure 6.03b for symbol legend.

composition. The Little Falls sedimentary rocks are strung out along the vertical axis to the highest Zr/Al_2O_3 values observed. The sedimentary rocks taken from the Hook Lake area and lower Lumby Lake sequence are associated with these most felsic sedimentary rocks from Little Falls Lake. The high Zr/Al_2O_3 ratios may reflect the sorting behaviour of Zr. As seen earlier Zr tends to have slightly increased values in the sand fraction as compared to the clay fraction. These sediments are fairly quartz rich and the lower abundance of matrix material may be expressed by these higher ratios of Zr/Al_2O_3.

Potential source rocks are also plotted in Figure 6.19. The Lapilli tuff samples and the Finlayson volcanic sample plot in the vicinity of the Steep Rock Lower Mafic unit. The Steep Rock sediment samples are again variable, but most plot near the borderline between the Finlayson and Little Falls fields. The main cluster of samples taken from the Little Falls Lake area plot in the same field as the Old Tonalite and the Young Tonalite, suggesting that these sedimentary rocks were derived from material having similar ratios to these granitic bodies. The most felsic samples from the Little Falls Lake area and the samples from lower Lumby Lake and Hook Lake tend toward the Steep Rock Upper Felsic point. These sedimentary rocks may have a similar volcanic source. Again it appears that the main group of Finlayson Lake sedimentary rocks and the Norway Lake sedimentary rocks plot on a mixing line between the Old Tonalite and the Steep Rock Lower Mafic unit.

Figure 6.20 is a plot of TiO_2/Al_2O_3 against Y/Al_2O_3. The slightly mobile nature of Y may be expressed by the wide spread of data, suggesting more variable Y values. A clustering of the data into separate field is however still visible in this diagram. Again a felsic to mafic trend is expressed by increasing TiO_2/Al_2O_3 values. The main group of Finlayson Lake sedimentary rocks and the Norway Lake sedimentary rocks plot on a mixing line between the Old Tonalite and the Steep Rock Lower Mafic units, suggesting that rocks similar to both sources contributed to these sedimentary rocks. The Little Falls Lake samples plot in the same field as the Old Tonalite and the Finlayson Upper Felsic unit. Some of the Little Falls samples, the lower Finlayson samples, and the lower Lumby Lake samples are strung out along a mixing line between the Old Tonalite and the Steep Rock Upper Felsic unit. These points are quite variable, possibly suggesting the magma which erupted debris from which these sedimentary rocks were derived evolved from a composition similar to the Old Tonalite, to one more like the Steep Rock Upper Felsic volcanics. As in the two previous plots the Lapilli tuff samples from Little Falls Lake and the Finlayson Lake volcanic sample plot very near the Steep Rock Lower Mafic unit. In this plot the Steep Rock Lake sedimentary rocks are clearly associated with the main group of Finlayson Lake sedimentary rocks.

A plot of MnO/Zr vs. V/Zr can be seen in Fig. 6.21. Both V and MnO are compatible elements and have higher concentrations in mafic rocks than in felsic rocks. It follows that these elements should be concentrated within sedimentary

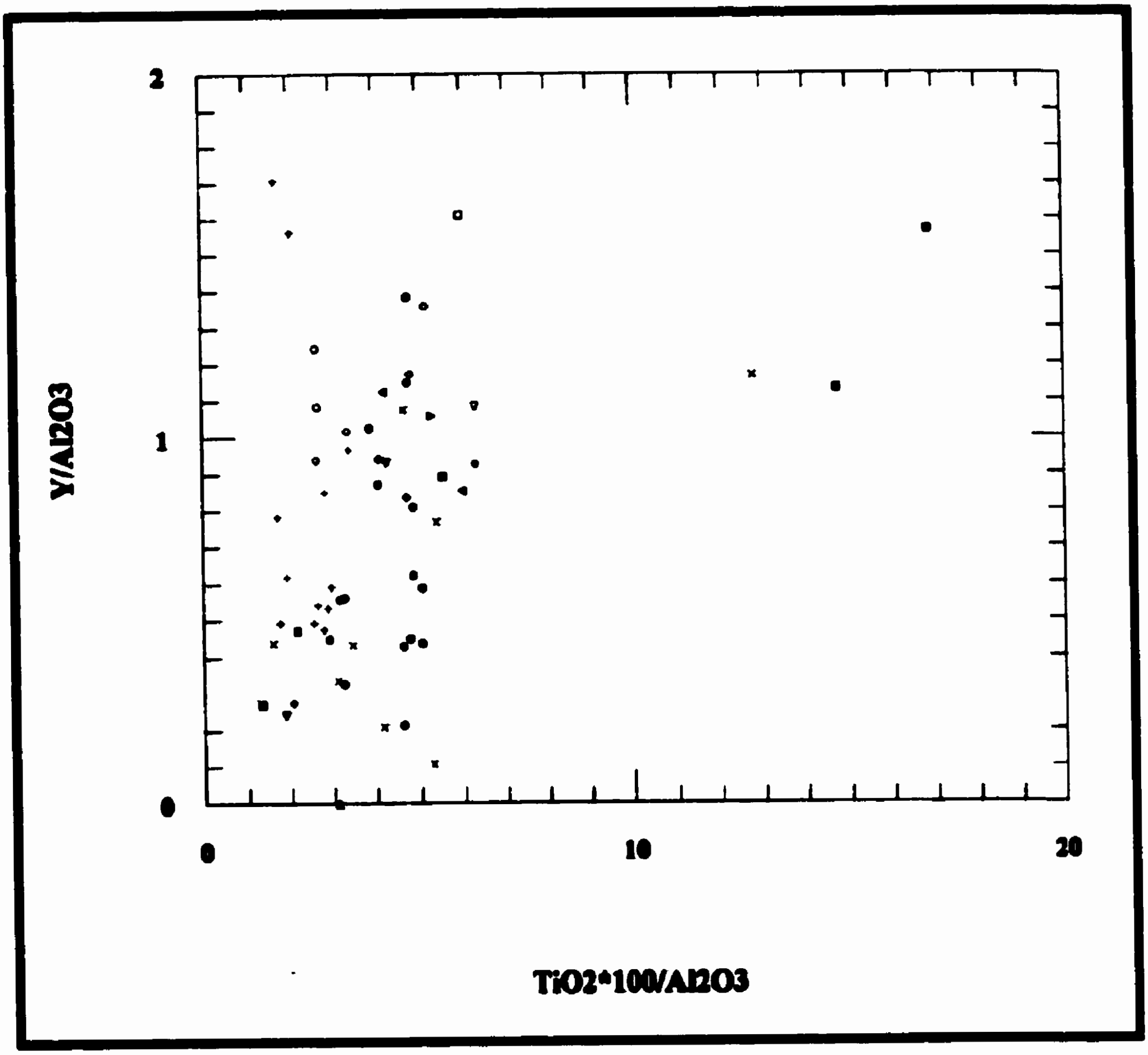


Figure 6.20 TiO2-Al2O3-Y ratio diagram for all samples taken within the Finlayson and Lumby Lake Greenstone Belts. Refer to Figure 6.03b for symbol legend.

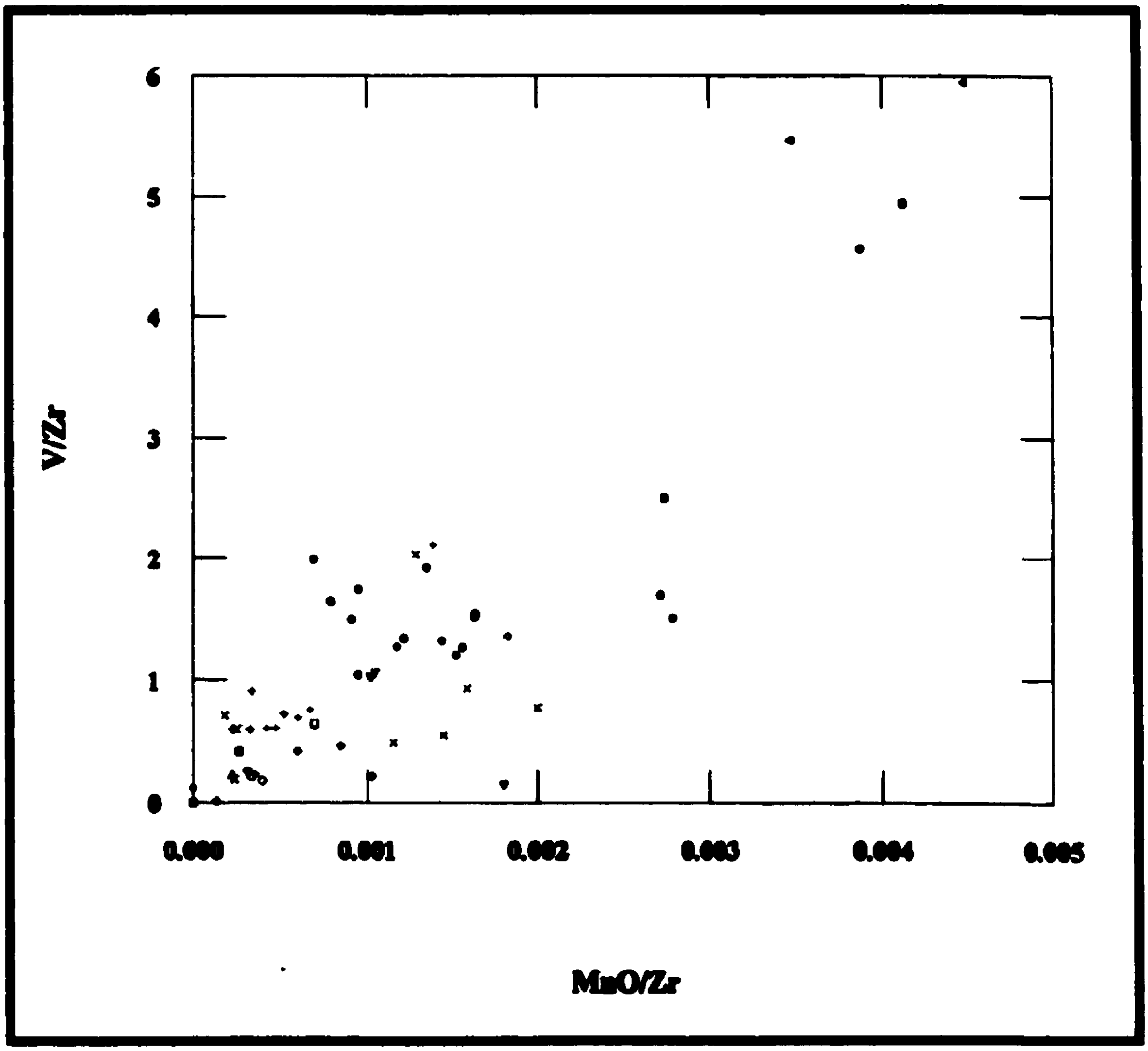


Figure 6.21 MnO-Zr-V ratio diagram for all samples collected within the Finlayson and Lumby Lake Greenstone Belts. Refer to Figure 6.03b for symbol legend.

rocks which have had a significant mafic source. As can be seen, The Little Falls Lake, Lower Finlayson Lake, and the Lower Lumby Lake sedimentary rock samples form a cluster as do the Main Finlayson Lake and the Norway Lake samples (Fig. 6.21). The samples taken from the Main Finlayson Lake area and those from Norway Lake have higher V/Zr and MnO/Zr ratios, suggesting a greater mafic contribution to the sedimentary rocks within these two areas.

The lapilli tuff samples from Little Falls Lake, and a conglomerate clast from Norway Lake plot very near the Steep Rock Lower Mafic unit (SRLM). The Steep Rock Upper Felsic (SRUF) plots within the cluster defined by the Little Falls Lake samples and the Lower Lumby Lake samples. The Norway Lake samples and the main Finlayson Lake samples lie between these points and may represent a mixing of two sources with compositions similar to the Steep Rock Lower Mafic unit and the Steep Rock Upper Felsic unit or Old Tonalite. It is interesting that the Steep Rock Lake sandstone samples tend to separate into two clusters, one grouping of similar composition to the more felsic rocks of the Little Falls Lake, Lower Finlayson and Lower Lumby Lakes, and the remaining samples tend toward the main group of Finlayson Lake and Norway Lake samples.

Cr/Al2O3 vs. Ni/Al2O3 (Figure 6.22) shows a similar trend as MnO/Zr vs. V/Zr . Again, the Little Falls Lake, Lower Finlayson Lake and Lower Lumby Lake samples cluster, while the Main Finlayson Lake and the Norway Lake samples show higher Cr/Al2O3 and Ni/Al2O3 ratios (Fig. 6.22). These ratios should be

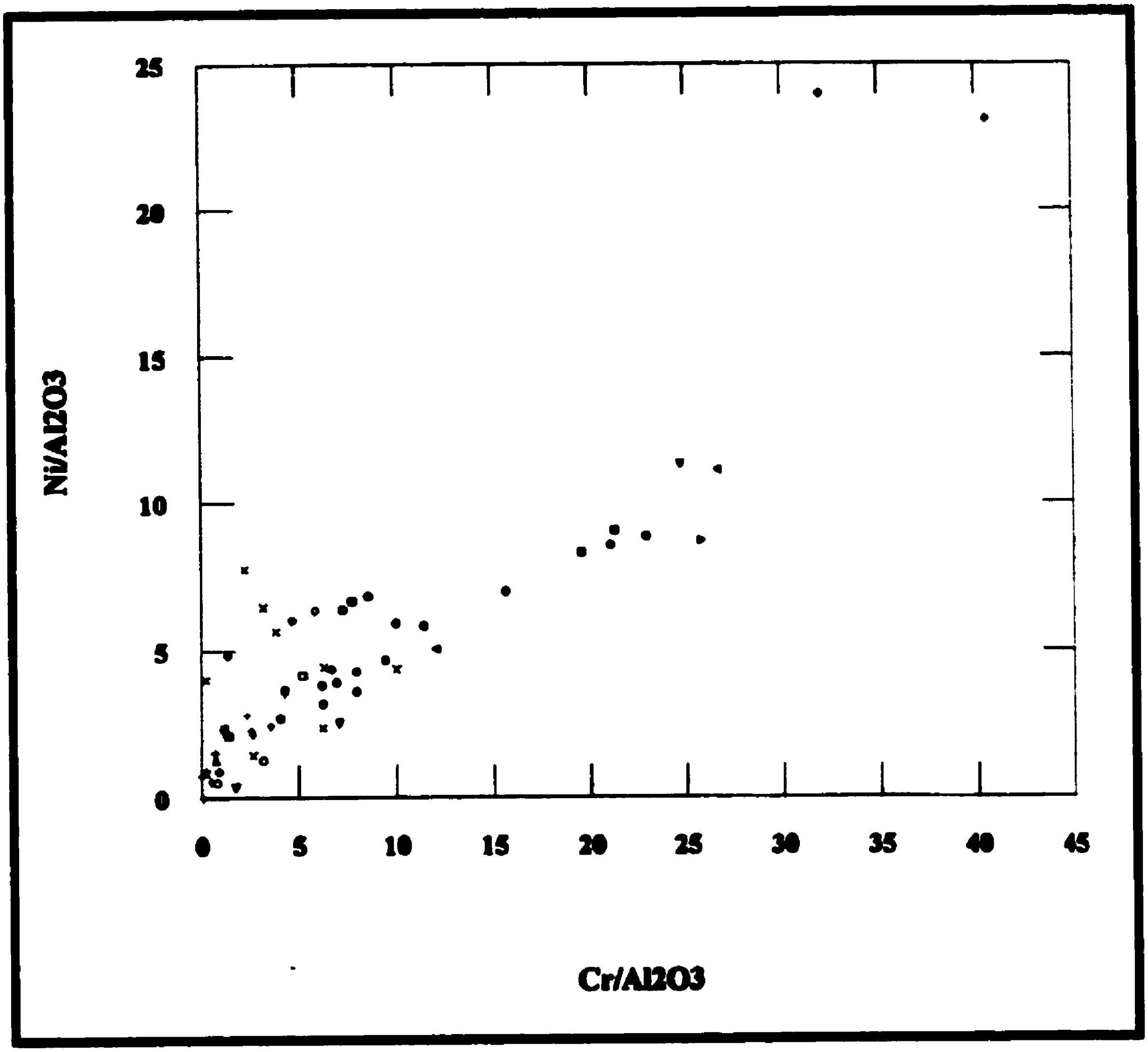


Figure 6.22 Al2O3-Cr-Ni ratio diagram for samples collected within the Finlayson and Lumby Lake Greenstone Belts. Refer to Figure 6.03b for symbol legend.

higher within rocks with a significant mafic volcanic component. The Norway Lake samples have significantly higher ratios than all other samples, which suggests a substantial mafic, or possibly ultra-mafic contribution.

A diagram developed by Winchester and Floyd, (1977) that is normally used to discern the composition of volcanic rocks may also be useful here to identify the type of source material for sediments as the provenance for all units is first cycle igneous (Figure 6.23). Again a fairly tight grouping of the sedimentary rocks into separate fields can be observed. The sedimentary rocks taken from the Little Falls Lake, lower Finlayson and lower Lumby sequences all plot in the same field, with a suggested rhyodacite-rhyolite source composition. As indicated by previous diagrams, a felsic source is likely to have shed these sedimentary rocks. All of these sedimentary rocks have a source area of similar composition. Sediments of the main group within Finlayson Lake and the Norway Lake sediments all plot towards the more mafic end of the diagram in the alkaline to subalkaline basalt fields. These sedimentary rocks have a much more mafic sediment contribution than the Little Falls and lower Finlayson Lake sequences. Several Steep Rock Lake samples plot near the Finlayson Lake sedimentary rocks, while others are distinct, plotting within the andesite field. This may reflect possible alteration of some of these samples.

The volcanic samples taken from Little Falls Lake and Finlayson Lake both plot very near to the Steep Rock Lower Mafic unit and the Metagabbro points. The Little Falls Lake and lower sequence sedimentary rocks plot very near the

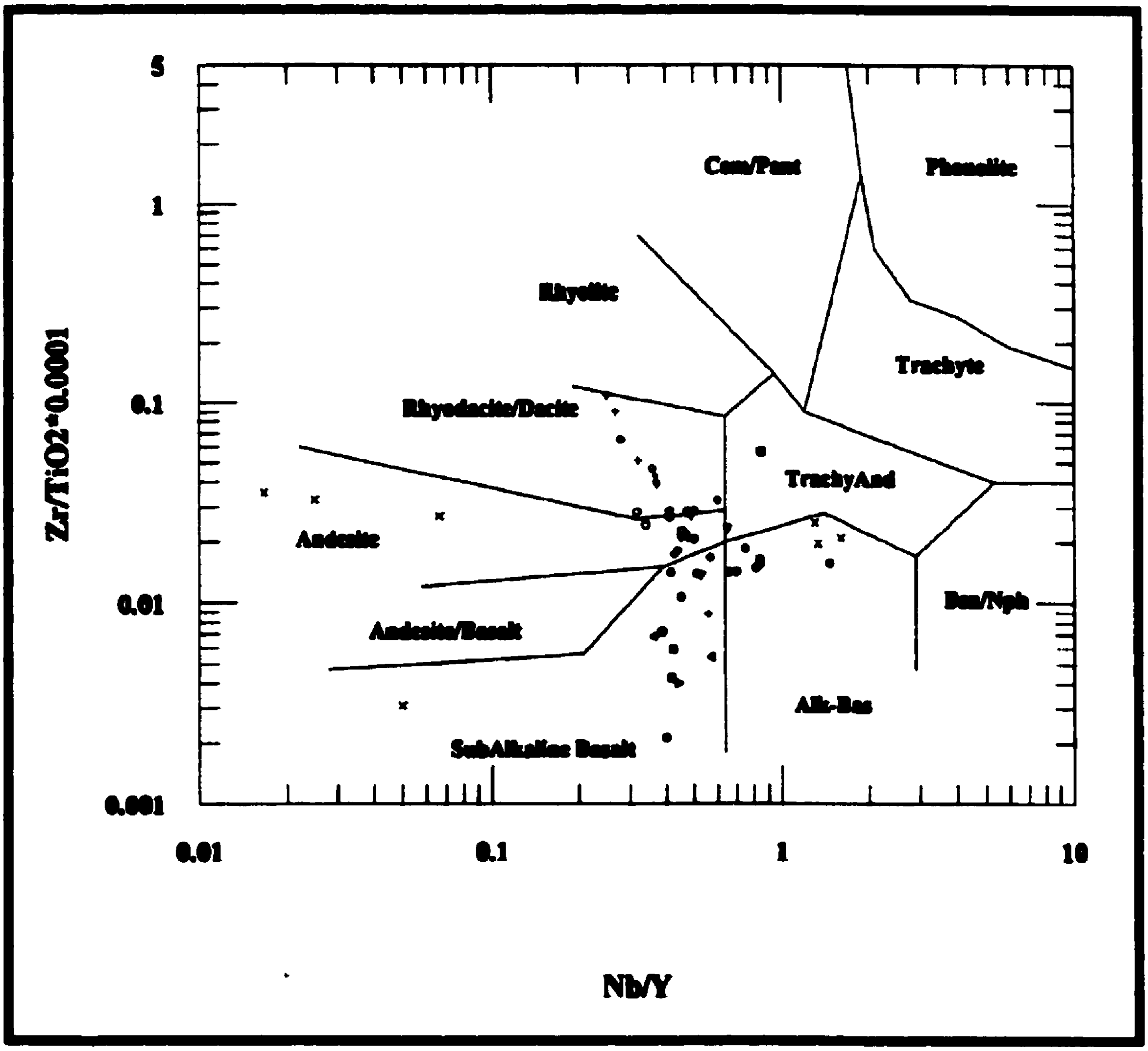


Figure 6.23 Winchester and Floyd (1977) discrimination diagram for samples Collected within the Finlayson and Lumby Lake Greenstone Belts. Refer to Figure 6.03b for symbol legend.

Steep Rock Upper Felsic unit and the Old Tonalite. As in previous diagrams, a felsic source, which resembles the Steep Rock Upper Felsic Unit and the Old Tonalite, is suggested for the Little Falls area and the Lower Lumby sequence. The Finlayson main group and Norway Lake sedimentary rocks lie on a mixing line between the felsic source material and the Steep Rock Lower Mafic Unit.

All graphs show a fairly clear grouping of the sedimentary rocks from the Finlayson and Lumby Lake belts. The sedimentary rocks can be divided into two fairly broad categories, one having a more felsic source than the other. It is quite clear that the main group of Finlayson sediments and the Norway Lake sediments had a geochemically similar source; with the only difference being a possible ultramafic contribution to the Norway Lake sequence. It is interesting to note that komattiites have been mapped in the volcanic succession near Norway Lake (Jackson, 1985). These were likely the result of erosion and mixing of sediment from a felsic volcanic and the Steep Rock Lower Mafic unit, or the Finlayson Lake mafic volcanics. It is also interesting to note that the lower Finlayson Lake sedimentary rocks always plot in the same field as the Little Falls Lake sedimentary rocks rather than the main body of Finlayson Lake sedimentary rocks. It is likely that the Little Falls Lake sedimentary rocks are lateral equivalents of the lower Finlayson sequence rather than the main Finlayson sequence, as previously correlated (Stone et al., 1992). The lower sequence in the Lumby Lake belt is also similar in composition to Little Falls

Lake sedimentary rocks and likely had a similar source. A much more felsic source, such as the Steep Rock Upper Felsic unit, which is possibly a more evolved extrusive equivalent of the Old Tonalite, is suggested for these sedimentary rocks.

In comparison the sandstone samples from the Steep Rock Group tend to split into two fractions, one group similar to the main Finlayson Lake and Norway Lake sedimentary rocks and the other has a more felsic composition. It is possible that this bimodality may reflect sediment from two different source areas being transported through incised fluvial channel systems, represented by the Wagita Formation (Wilks and Nisbett, 1986), before mixing.

6.3 U-Pb Geochronology

Radiogenic isotopes can be used as a powerful geochronological tool as well as in petrogenetic studies, where they are helpful in identifying geological processes and sources. Several U/Pb age determinations were done on rocks within the Finlayson and Lumby Lakes Greenstone Belts. Also, recently, studies using the Sm/Nd system have been attempted.

The U/Pb system involves two independent geochronometers. Data may be evaluated using a concordia diagram. The ratio of concentrations of isotopes of U and Pb in a rock form a discordant line where the upper intersect with the concordia curve is significant in determining the possible age of crystallization. For a more detailed explanation of the U/Pb system refer to Faure (1986) and

Rollinson (1993).

Several of the felsic volcanic and plutonic rocks in both the Finlayson Lake and the Lumby Lake area have had U/Pb age determinations done in the past. Stone et al. (1992) discusses the geochronology of several of the igneous rocks surrounding the Finlayson Lake Greenstone Belt. The U/Pb ages referred to in the discussion are taken from D. Davis (pers. comm.).

One of the oldest ages within the area was obtained from a sample of the Mafic Tonalite collected along the transmission line north of Wagita Bay on Steep Rock Lake. This sample yielded a U/Pb age of 3001 Ma. Mafic Tonalite is found as scattered xenoliths within the Old Tonalite unit of the Marmion Batholith, which is believed to be approximately synchronous in places (P Pufhal, pers. comm.) and as isolated intrusions north-east of Finlayson Lake.

Old Tonalite is a major phase of the Marmion Batholith, the Lefteye Stock and several small intrusive bodies south-west of the Bow Pluton, along the eastern edge of the Finlayson Lake Greenstone Belt (Fig 1.03). A U/Pb titanite age of 2953 Ma was obtained from a sample of Old Tonalite which was taken near the Ontario Hydro generating station (Stone et al., 1992).

Tonalite Gneiss comprises most of the Dashwa gneiss complex to the north-west of Finlayson Lake (Stone et al, 1992). A sample of Tonalite Gneiss from the Dashwa Gneiss Complex north-east of the Eye Dashwa Pluton (Fig. 1.03) yielded a U/Pb zircon age of 2928 Ma (Stone et al., 1992).

All three of these plutonic intrusive units adjacent to the Finlayson Lake

Greenstone belt, yield U/Pb ages within a span from ≈2928 to ≈3001 Ma. This indicates that the plutonic rocks intruding the lower portions of the Finlayson Lake Greenstone Belt were intruded and cooled during this ≈70 Ma period.

Six sedimentary zircons taken from the Wagita Formation gave an age of approximately 2999 Ma (D. Davis pers. comm., Fralick and King, 1996). The Wagita Formation is a fluvial unit consisting of conglomerate and sandstone, and is found at the base of the Steep Rock Group (Wilks and Nisbet, 1988).

Felsic tuffs near the top of the volcanic sequence within the Northern Finlayson Lake area yield a zircon age of approximately 2930 Ma (D. Davis pers. comm., Fralick and King, 1996).

Some of the rocks within the Lumby Lake Greenstone Belt yield ages consistent with those acquired within the Finlayson Lake area. A geochronological study was done by Davis and Jackson (1988). Two felsic tuff samples and a rhyolite, collected from the same stratigraphic level, and laterally continuous with the Lower Lumby Lake Sedimentary Belt, were analysed. These samples gave a U/Pb zircon age of 2999 ± 1 Ma. In addition to these a tonalite gneiss sample from the Marmion Lake Batholith was also analysed in the study, and yielded an age of 3003 ± 5 Ma.

A sandstone sample collected for the present study from the Southern Finlayson Lake area was also analysed by D. Davis of the Royal Ontario Museum. Six zircons from the sample gave ages of 3002 ±0.9, 2997 ± 2.5, 3001 ± 0.8, 2999 ±

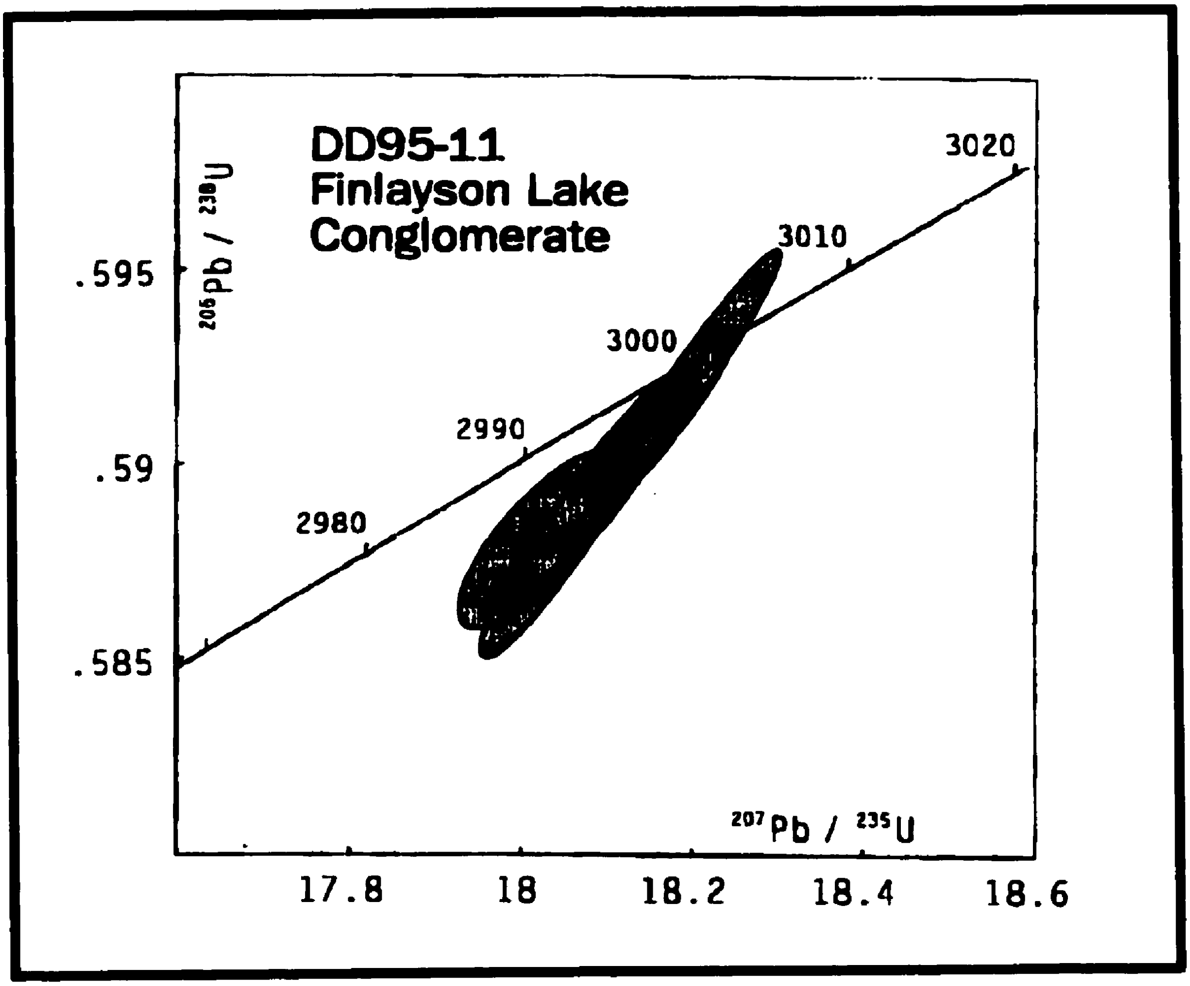


Figure 6.24 Concordia diagram for a sandstone sample collected from the southern Finlayson Lake area.

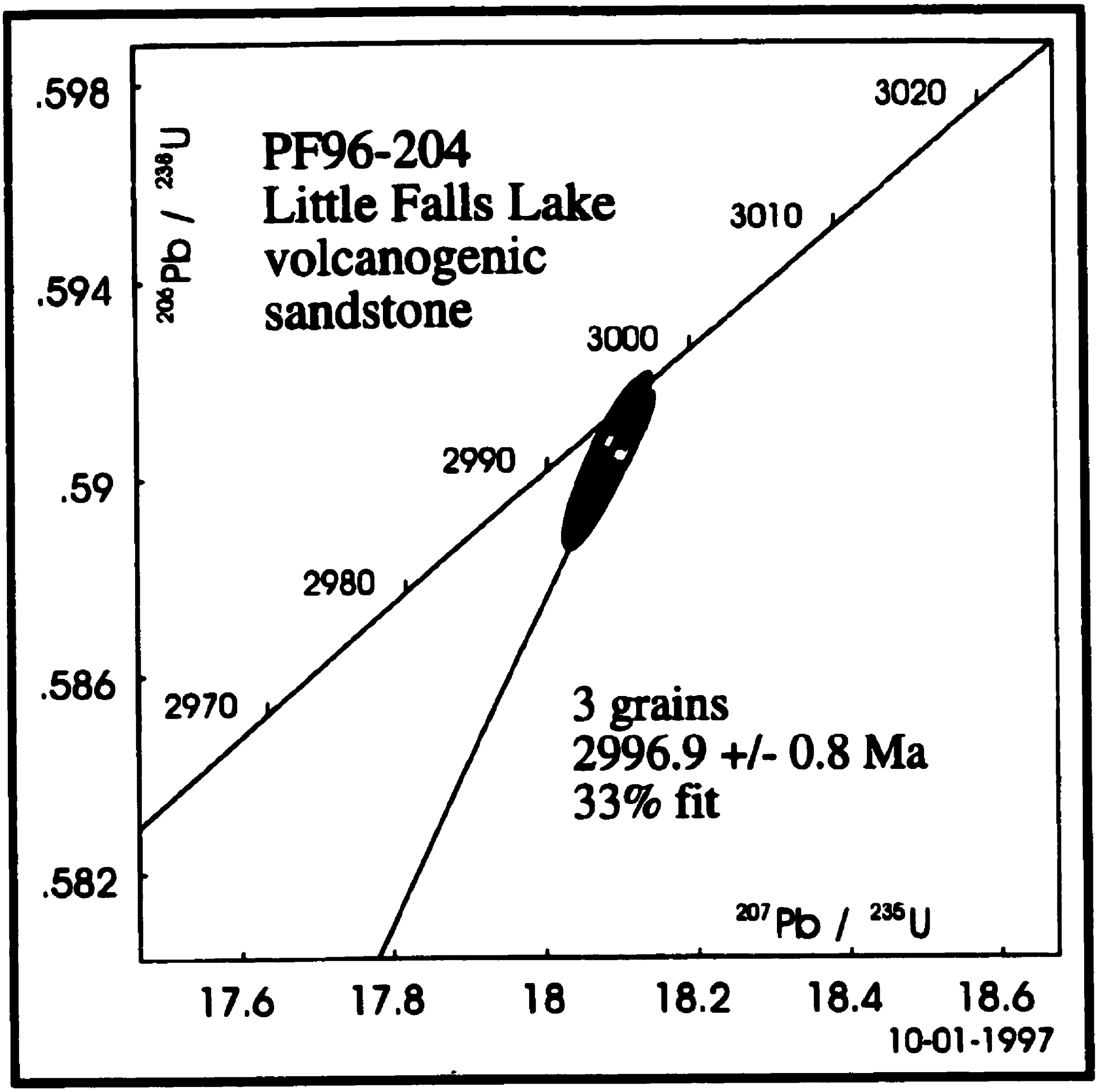


Figure 6.25 **Concordia diagram for a sandstone sample collected from the Little Falls Lake area.**

1.2, 2999 ± 1.5, and 3001 ± 1.2 Ma . The concordia plot shows that the average of these data is approximately 3001 Ma, but gives a fairly low probability of fit (Fig. 6.24). This suggests that the sandstone had sources with slightly different ages, although it would appear that they may be different phases of the Marmion Batholith (D. Davis pers. comm.).

A second sandstone sample, collected from the Little Falls Lake area, was also analysed by D. Davis. This sample yielded three zircons which gave an age of 2996.9 +/- 0.8 Ma (Figure 6.25). All three grains have very similar ages, suggesting a single provenance rock for the sandstones within the Little Falls Lake area.

6.4 Sm/Nd Geochronology

The elements Sm and Nd are less mobile than some of the other radiogenic isotopes used in geochronological studies. For this reason they may be able to "see through" younger events which have effected the rocks. Model ages, a measure of the length of time a sample has been separated from the mantle, can give valuable insight into the age of igneous rock, and therefore possible ages of source rocks for sedimentary sequences. Model age calculations make an assumption about the isotopic composition of the mantle source from which the sample was derived, and must be interpreted with care. The two frequently quoted models for the Nd isotope composition of the mantle reservoir involve either the Chondrite Uniform Reservoir (CHUR) or the Depleted

Mantle (DM). For a more detailed discussion of Sm/Nd systematics refer to Faure (1986) and Rollinson (1993).

Detrital sedimentary rocks are composed of rock particles from the igneous and metamorphic rocks from which they were derived, therefore their Sm/Nd compositions will depend on those of the rock particles of which they are composed. Studies have shown that the Sm/Nd ratio of sedimentary rock is not effected by the processes of chemical weathering, transport, and deposition and are similar to the ratios of there provenance rocks (McCulloch and Wasserburg, 1978).

Several sedimentary rock samples collected for this study were analysed for Sm/Nd and compared to gneisses of the Marmion Lake Batholith by Dr. Ross Stevenson of the University of Montreal.

The data, provided by Dr. R. Stevenson, from four sedimentary rock samples and four samples of tonalitic gneiss from the Marmion Lake Batholith, are given in Table 6.1.

As can be seen in table 6.1, the model age of all but one of the samples, with respect to DM, is very close to 3.0 Ga, generally between 2.9 and 3.1. The TDM value is an estimate of the time at which the samples separated from their mantle source region. The sedimentary rock samples consistently give a model age slightly greater than that of the gneiss samples, 3.02 - 3.9 Ga for the sedimentary rocks, and 2.9 - 2.96 Ga for the tonalitic gneisses. This suggests that the sedimentary rock samples contain a component which separated from the mantle source at an earlier time than the tonalitic gneiss.

95 TB 4	gneiss	3,000	1.39	8.51	0.0989	0.51083	11	2.5	0.6	2.80	2.96	-1.3
95 TB 5a	gneiss	3,000	3.06	13.46	0.1373	0.51158	14	2.4	0.7	2.69	2.95	0.1
95 TB 6a	gneiss	3,000	3.21	14.53	0.1335	0.51151	15	2.4	0.7	2.71	2.96	-0.1
95 TB 8	gneiss	3,000	3.22	18.46	0.1051	0.51099	14	3.2	0.6	2.73	2.90	-0.4
N-1	sedts	3,000	4.62	14.32	0.1949	0.51269	10	1.8	0.8	-4.69	3.90	1.7
DF-14	sedts	3,000	2.01	10.66	0.1135	0.51106	9	1.8	0.6	2.84	3.02	-1.5
2-34	sedts	3,000	3.01	14.18	0.1282	0.51135	9	1.3	0.6	2.85	3.07	-1.3
DP-02	sedts	3,000	4.39	19.69	0.1348	0.51147	9	1.1	0.7	2.86	3.11	-1.3

Table 6.1 Sm/Nd data taken from Dr. R. Stevenson, University of Montreal

The εNd value is a measure of the deviation of a sample from the expected value in a uniform reservoir, in most cases CHUR, and is useful as a comparison value for samples of similar age. The εNdT value represents the initial εNd value at the time of crystallization, and can provide information about the source magma of the sample. The εNdT values for all the samples are positive at the age of crystallization. These values suggest that all the samples had a depleted mantle source, with Sm/Nd ratio greater than CHUR. At the age of crystallization the εNdT value for the gneiss samples ranges from 2.5 - 3.2 and the εNdT value for the sedimentary rock samples ranges from 1.1 - 1.8. This suggests, that while the gneisses may be derived from tonalitic melts generated from basalts, the sedimentary basins also received detritus that was older than the 3.0 Ga tonalites, but not greatly older (Stevenson , pers. comm.). This older detrital component in the sedimentary rocks was not recognised by the U/Pb zircon geochronology which pointed to the tonalite of the Marmion Batholith as the source of detritus for the Upper Finlayson Lake sedimentary rocks. This implies that the older component of the sedimentary rocks may have been of a mafic composition, which would not have shed detrital zircons into the basin.

In summary, from the geochronological data for the Finlayson and Lumby Lakes area several conclusions can be drawn. First, the sedimentary rocks of the Lower Lumby Lake Sedimentary Belt are laterally continuous with felsic volcanic rocks that have a U/Pb zircon age of approximately 2999 Ma, and likely were deposited during the same time period. Secondly, it can also be concluded that the sedimentary rocks within the Little Falls Lake area had a single source rock, with an age of approximately 2997 Ma. Thirdly, the sedimentary rocks within the Upper Finlayson and Lumby Lakes Sedimentary Belts had more than one source rock, an approximately 3.0 Ga source such as the tonalite of the Marmion Batholith, and a component with a mafic composition, that is older than the 3.0 Ga tonalite.

CHAPTER 7

DEPOSITIONAL ENVIRONMENTS WITHIN THE FINLAYSON LAKE GREENSTONE BELT

7.1 Introduction

This chapter discusses the processes which are most likely to have deposited the sedimentary rocks now exposed in the area, and suggests possible depositional environments in which these processes acted. For simplicity the discussion is in the same format as previous chapters, dividing the Finlayson Lake Greenstone Belt into separate geographical areas. The Little Falls Lake area and the Lower Finlayson Lake Sedimentary belt contain similar sedimentary assemblages, just as the Southern and Northern Finlayson Lake areas contain sedimentary rocks which are very similar.

Many of the clastic sedimentary facies found within both the Finlayson Lake Greenstone Belt and the Lumby Lake Greenstone Belt can be attributed to gravity flow processes such as grain flows, debris flows and turbidity currents. Sediments deposited by turbidity currents were described by Bouma (1962). Low-density turbidity currents result in the deposition of sediments with a regular pattern, now recognised as the classical Bouma sequence (Bouma, 1962). From bottom to top the classic turbidite facies consists of five distinct divisions: T_a - massive to graded sand; T_b - parallel laminated sand; T_c - cross-laminated, ripple laminated, and/or convolute bedded sand; T_d - parallel laminated fine-grained sand and silt; T_e - pelagic mud. All five divisions are rarely found in a single turbidite bed, instead partial sequences are usually developed.

Two types of turbidity currents, high- and low-density, are recognised (Lowe, 1982). A single turbidity current may evolve, initially depositing coarse-grained sediment as a high-density turbidity current, then gradually transforming into a low-density turbidity current (Lowe, 1982). The classic Bouma Sequence described above is attributed to deposition by low-density turbidity currents, from which fine-grained sand and silt are deposited by traction and suspension. High-density turbidity currents occur as sandy or gravely flows. According to Lowe (1982) deposition by high-density turbidity currents can be traced through three stages: I) a traction sedimentation stage; II) a traction carpet stage; and III) a suspension-sedimentation stage. This sequence is attributed to increasing flow unsteadiness and the collapse of the high-density turbidity current. The first stage, traction sedimentation, will deposit sand beds which may produce bedforms similar to those developed in low density flows, including plane beds and dune-like features, resulting in flat lamination and oblique or cross-stratification (Lowe, 1982). During the second stage of deposition, traction carpet, the concentration of coarse particle near the bed rises, and transport in the bed-load layer becomes dominated by grain collisions, resulting in deposition of successions of inversely graded coarse-grained sand. The third stage of deposition involves direct suspension sedimentation. At higher suspended load fallout rates it is not possible to develop a bed-load layer or an organised traction carpet and sedimentation is by direct suspension sedimentation (Walker, 1978). The resultant deposit is grain supported and lacks traction structures and may be equivalent to the T_A of the classic Bouma sequence (Lowe, 1982).

Grain-flows are quasi-visco-elastic flows which are characterized by grain to grain collisions which result in a dispersive pressure that keeps the grains in suspension (Reading, 1996). Grain-flows generally require steep slopes, near the angle of repose, between 18°-28° and deposition from grain-flows is due to frictional freezing. Grain-flows generally deposit thick, massive sandstones, that lack internal structure and have sharp upper and lower contacts with few interbeds (Stauffer, 1967; Reineck and Singh, 1980). Lowe (1982) suggests that true grain-flows rarely exceed 5 centimetres in thickness because the grains at the base cannot produce dispersive pressures great enough to support against the gravity of a thicker layer (Lowe, 1982). Thicker beds, in excess of 0.4 meters are the result of density modified grain flows, depositing poorly sorted sand, with little grading.

Rapid deposition of thick massive sands may occur due to non-uniformity in prolonged, quasi-steady high-density turbidity currents (Kneller and Branney, 1995). These authors suggested that the thickness of the resulting sandstone beds has no relationship to the thickness of the current which deposited them. They proposed a method of gradual aggradation of sand beneath a migrating flow boundary. Deposition of sediment is not instantaneous and gradual deposition of sands can occur for as long as it takes the current body to pass. This process will result in massive sandstones, which can be up to several meters in thickness. In some cases normally graded tops or mud drapes may be present separating successive sandstone beds.

7.2 Little Falls Lake Area

The Little Falls Lake area is dominated by massive, coarse-grained sandstones, with lesser monomictic conglomerate near the top of the sequence. Mafic detritus-rich sandstones and lapilli tuff layers are randomly interbedded within the lower part of the section.

The coarse-grained sandstones are generally massive, with rare grading near the top of the beds, and rare mud drapes. Bed thickness ranges between a few tens of centimetres and a few meters. These sands are only consistent with deposition from either, or both, of two processes discussed previously, grain-flows or prolonged high density turbidity currents as described by Kneller and Branney, (1995). Lowe (1982) suggests true grain flows rarely exceed 5 centimetres in thickness, therefore dispersive pressure was not the primary force. Rapid aggradation beneath a migrating flow boundary is more likely.

The conglomerate facies found within the Little Falls Lake area were deposited from similar mass-flow processes or high density turbidity currents. Conglomerates are interpreted as either more proximal deposits, or the result of larger, more powerful flows, which brought coarser-grained sediment further into the basin. The conglomeratic units are more commonly found within the upper part of the sequence in the Little Falls Lake area, which suggests that they are the result of progradation of the depositional system.

The sedimentary rocks in the Little Falls Lake area are laterally transitional with the sequence of felsic volcanic rocks to the south. Sedimentary rocks form a wedge, thickening to the south, towards the felsic volcanic rocks (Stone et al, 1992). This suggests that they resulted from the redeposition of felsic tuffs. The sedimentary sequence developed here represents the subaqueous portion of a clastic debris apron, which received clastic material from the volcanic centre to the south. Massive sandstones and conglomerates higher in the succession are a result of progradation of the debris apron. Periodic, local, mafic volcanism deposited isolated lapilli tuffs and resedimentation led to the development of mafic detritus-rich sandstones, intermittently interbedded within the lower portion of the sequence. The lack of reworked tops of beds suggests that the sediments were deposited at water depths below the wave base.

7.3 Lower Finlayson Lake Area

The Lower Finlayson Lake sedimentary belt consists of a series of isolated pods of sedimentary rock which are similar in character to the Little Falls Lake sedimentary unit. Similar poorly sorted massive sandstones are present, although the grainsize is reduced. Conglomerate and mafic detritus-rich sandstones are not found in the area.

Grain flow and high-density turbidity current processes are again suggested as the dominant depositional processes acting in the area. This interpretation is the only one consistent with the poorly sorted, massive sandstones present here. The sandstones are more distal deposits of the same subaqueous clastic debris apron and were fed from the same volcanic centre which delivered sediment to the Little Falls Lake area (see Provenance chapter).

7.4 Southern Finlayson Lake

Both the southern and northern Finlayson Lake areas are part of the Upper Finlayson Lake Sedimentary Belt, and share many of the same sedimentary rock facies. Clastic facies present in the southern Finlayson Lake area consist of fine-grained sandstone and siltstone couplets, medium-grained sandstone, coarse-grained scoured sandstone, and conglomerate. Chemical sediments include chert as well as oxide and sulphide facies iron formation.

Fine-grained sandstone and siltstone couplets are the product of low density turbidity currents (Lowe, 1982). Most beds within this facies contain parallel laminated fine-grained sandstone and siltstone at their base which is interpreted as the T_d division of a low-density turbidity current deposit (Bouma, 1962). The fine silt to clay rich tops of this facies are the T_e division.

The medium-grained sandstone facies is also consistent with deposition from turbidity currents and may represent the natural evolution from high-density to low-density turbidity current deposition. The beds consist of a basal, normally graded, medium-grained sandstone division, commonly showing a gradation upward into a parallel laminated fine-grained sandstone and siltstone division, which may or may not be capped by mudstone. This facies is interpreted as $T_{a,d,e}$ and $T_{a,d}$ turbidity current deposits. Although Bouma (1962) interprets the T_a division as the result of low-density turbidity current deposition, Lowe (1982) suggests that the T_a division may in fact be the final stage of deposition by a high-density turbidity current. There is no evidence that the T_a division of these beds forms by high-velocity traction sedimentation, and experimental results produced similar beds that were deposited by direct sedimentation from high-density flows (Middleton, 1967). Therefore, this facies may represent the transition from high- to low-density turbidity current deposition during a single turbidity current event. The basal, normally graded sands are consistent with the S_3 facies of Lowe (1982), deposited from high-density turbidity currents and the parallel laminated fine-grained sandstone and siltstone division is consistent with deposition from a low-density turbidity current.

Coarse-grained scoured sandstones are also the result of turbidity current deposition. However, this facies is more consistent with sediments deposited from high-density turbidity currents. These sandstones are composed of coarse-grained sand, containing small pebbles. This facies is the result of traction sedimentation, the first stage of deposition from a high-density turbidity current, and are interpreted as the S_1 division described by Lowe (1982). The internal scours contained within these units are filled with coarser-grained sandstone and pebbles, and are the same scale as those described by Lowe (1982). This facies is also similar to the tabular-bedded pebbly sandstone facies described by Balance (1988). Minor scouring and deposition of coarser-grained fill occurred locally, due to fluctuations in flow-velocity, during deposition from a single turbidity current event.

Conglomerate in the Southern Finlayson Lake area is commonly lenticular bedded and often contains sandstone lenses within the conglomerate. The lenticular bedding in this facies suggests that the conglomerates represent more structured, possibly channelled flow. The massive, non-graded and poorly sorted nature of these conglomerates is consistent with deposition within a channel.

Chemical sedimentary facies in the southern Finlayson Lake area include chert, magnetite, and sulphides, all of which are commonly intimately associated with one another. Alternating chert and magnetite-rich iron formation as well as alternating chert and sulphide-rich iron formation are common within units up to tens of meters thick. The deposition of Archean iron formations is not fully understood, although there have been several studies into possible processes which may have led to their accumulation. The source of the iron and silica is one factor which is under debate. Possible sources of iron include subaerial and subaqueous weathering of iron bearing minerals (Garrels et al, 1973: Hubber, 1959), or a hydrothermal source (Gross, 1965: Fralick, 1987). A hydrothermal source has been proposed for the silica as well (Gross, 1965; Fralick, 1987). Changes in the Eh, Ph, or temperature of the solutions are needed to cause precipitation of the constituents forming iron formations. These changes can be achieved by the venting of hot, saturated, acidic, hydrothermal fluids (Fralick, 1987: Barret et al., 1989). It is suggested that hydrothermal venting may cause a zone of oversaturation and precipitation which extends tens to hundreds of kilometres away from the vent (Fralick, 1987). Proximity to the source of venting fluids may play a role in the facies of iron formation which is precipitated. Sulphide facies are likely precipitated nearer to source vents, where conditions may be more reducing, while oxide facies iron formation may occur further from the vent (Goodwin, 1973). The rhythmic alternations between chert and iron-rich layers cannot be sufficiently explained, other than it is likely due to fluctuations in the Eh, Ph, temperature and dissolved concentrations in the surrounding environment and venting hydrothermal fluids. One possibility is an alternation between black and white smoker fields. Black smoker may develop over the shallowest points of magma chambers, emitting iron rich fluids (Ballard et al., 1981,; McConachy et al., 1986). As the magma chambers cool the black smokers are transformed into white smoker fields, and with recharge of the magma chamber, may revert back to black smoker activity.

Iron formations are common in the Finlayson Lake area and are the result of venting of hydrothermal fluids. The small-scale stockwork system present in the northern Finlayson Lake area is evidence of hydrothermal activity in the area.

The network of veins, which are now filled with iron carbonate, provided a conduit system, allowing fluids to travel up through the volcanic pile. Venting of the fluids caused precipitation of chert and iron formation, as evidenced by the chert capping the stockwork system which is present in the area.

The sedimentary assemblage present in the southern Finlayson Lake area is deposited on a thick unit of subaqueous mafic volcanic flows. A well developed coarsening upward sequence is present within the metasedimentary rocks, from iron formation and fine-grained hemipelagic sediments at the base to conglomerate and coarse-grained sandstones at the top.

Chert and iron formation is present at points in the sedimentary succession when there was little clastic input or within areas that were sheltered clastic influx. Iron formation is commonly present between the volcanic flows beneath the sedimentary succession and at the base of the sedimentary sequence.

With cessation of volcanic activity, continued venting of hydrothermal fluids deposited iron formation and chert in areas isolated from significant clastic influx. Chert and sulphide facies iron formation were deposited in areas more proximal to venting fluids, while chert and magnetite iron formation accumulated in more distal sites.

It is clear that turbidity currents played a major role in the deposition of the clastic sedimentary rocks which are found in the Finlayson Lake area. A sequence dominated by turbidity currents, as are the Finlayson Lake sedimentary rocks, immediately brings to mind deposition within a submarine fan system. However, there are significant differences between the Finlayson Lake sequence and typical deep sea submarine fan sequences. Deep sea submarine sequences have been described in the literature by many authors including Walker, (1978), Mutti, (1985) and Normark (1970). Although submarine fans display many similar sedimentary rock facies, and coarsening-upward sequences as those found in the Finlayson Lake area, the scale of the depositional system on a submarine fan is typically much greater. Submarine fan coarsening-upward sequences are often several hundreds to thousands of meters thick, with superimposed smaller scale fining- and thinning-upward sequences on the order of 10-50m thick (Walker, 1978). Also, submarine fan sequences can usually be divided into separate sub-environments, such as the lower fan, suprafan lobe, distributary channels and upper fan main channel, in which distinct sedimentary facies associations are present. The clastic sequence found in the Finlayson Lake area is on a much smaller scale, with little evidence of common submarine fan environments such as the upper fan which contains large scale fan channel sequences.

Another environment in which turbidity currents play a major role in deposition, and which may be a better analogue, is a deltaic system. Many turbidite dominated systems such as the 450m thick sequence of classic turbidites present in the Susquehanna Valley area of Pennsylvania have been interpreted as prodeltaic turbidites (Walker, 1971).

The clastic sequence in the Finlayson Lake area can best be interpreted as a deeper water, possibly, fan deltaic sequence. Facies present are similar to those described in the coarse-grained, volcanic-fed Huriwai delta of New Zealand (Ballance, 1988). Deeper water, at least below average storm wave base, is suggested due to the lack of reworking features such as wave-induced scour, and ripples. Initially prodelta sediments were deposited over the iron formation in the form of low-density turbidity current deposits, as clastic sediments began to enter the basin. With progradation of the delta, clastic sedimentation dominated proximal to the delta and, with time, coarser-grained sediments reached further out into the basin developing a coarsening-upward sequence. More channelled flow, possibly on the surface of lobes, is represented by the conglomerates. There are several smaller scale, coarsening-upward sequences developed near the top of the sedimentary succession which probably represent lobe switching.

The sedimentary succession is capped by siltstone and shale couplets which suggests that sediment input to the area decreased and a return to deeper water sedimentation occurred as relative sea level rose.

7.5 Northern Finlayson Lake

Sedimentary rocks in the northern Finlayson Lake area are similar to, and continuous with, the sedimentary succession exposed in the southern Finlayson Lake area. In the most northern parts of the area the sedimentary succession was deposited on a thick sequence of agglomerate rather than mafic flows and pillow lavas.

Clastic sedimentary rocks in the area include fine-grained sandstone and siltstone couplets, medium- to coarse-grained sandstone, and conglomerate. These metasedimentary units are laterally continuous with those found in the southern Finlayson Lake area and similar processes are responsible for there deposition. Turbidity currents were the main depositional mechanism operating in the northern Finlayson Lake area. Section F-2, (see Fig. 4.0) begins with a unit of polymictic conglomerate which is succeeded upward by iron formation. The conglomerate contains felsic volcanic, mafic volcanic and rare iron formation clasts. This conglomerate, which is found near the base of the sequence, is similar to conglomerate present in the Norway Lake area and may represent a localized source entering the basin in the Northern Finlayson Lake area. The presence of iron formation clasts suggests that during emplacement, the conglomerate may have eroded and incorporated some of the basal iron formation unit or the iron formation found between volcanic flows. Gravity flow processes are suggested as the sediment delivery system operating during emplacement of the conglomerate. Fluidized flows, initiated by slumping, are the most likely mechanism for the deposition of this conglomeratic unit. Slumping tends to deposit massive, highly disorganized beds, with a fine-grained matrix, as seen in the northern Finlayson Lake area.

The thick agglomerate sequence present in some areas suggests that the northern Finlayson Lake area may have been near a volcanic centre. Iron formation is common within the area, as thin units between successive agglomerate beds and at the base of the sedimentary sequence. Venting of hot hydrothermal fluids during and after cessation of explosive volcanic activity is again suggested as a source for the chemical sedimentary units found in the area. An increase in the amount of sulphide facies iron formation in this area may be further evidence that a volcanic centre was nearby.

During periods of volcanic quiescence, and after cessation of the mafic volcanism, chemical precipitates were deposited in areas with little clastic input. A small localized sediment source in the northern Finlayson Lake area is suggested by the deposition of conglomerate near the base of the sequence in a position correlated with units of iron formation elsewhere in the basin. As the deltaic system to the south grew, clastic sediments entered the basin and the distal deposits of turbidity currents reached the area. These are represented by the fine-grained sandstones and siltstones, which were deposited above the chemical sediments. As the delta to the south prograded, coarser-grained sediment arrived in the northern Finlayson Lake area, in the form of turbidity currents, and was deposited as the medium- to coarse-grained sandstone facies.

CHAPTER 8

DEPOSITIONAL ENVIRONMENTS WITHIN THE LUMBY LAKE GREENSTONE BELT

8.1 Hook Lake Area

The metasedimentary sequence exposed in the Hook Lake Area is part of the Lower Lumby Lake Sedimentary Belt. The sedimentary rocks here are on strike with felsic volcanic rocks exposed to the east. Clastic sedimentary rocks present in the area include medium-grained sandstone, fine-grained sandstone and slate. Chemical sediments include carbonate and chert.

The sandstones are thin to medium bedded, somewhat quartz-rich, and may be graded or non-graded. Sediment gravity flows, such as turbidity currents are the main mechanism of deposition. Interbedded fine-grained slates are background sedimentation deposited during periods of quiescence between the successive turbidity current events.

Chert units in the area are laminated on a centimetre and sub-centimetre scale. Venting hydrothermal fluids may have been the source of the silica. Carbonates are massive and recrystallized, however fractures with decimetre scale spacing may reflect original bedding. Precipitation from sea-water is the interpreted mechanism for the deposition of the carbonate units.

The sedimentary succession in the Hook Lake area represents the distal, reworked, debris shed from a volcanic centre situated to the east (see

Provenance chapter). Sediment was transported into the area via turbidity currents. Carbonate and chert were precipitated in areas which were sheltered from clastic input, while mud deposition in other areas was interrupted by turbidity current events. The sequence gives very little information as to the depth of water in which the sediments were deposited.

8.2 Norway Lake Area

The sedimentary rocks exposed in the Norway Lake area are the western extension of the Upper Sedimentary Belt within the Lumby Lake Greenstone Belt. This sedimentary succession is similar to the Upper Sedimentary Belt within the Finlayson Lake Greenstone Belt and is likely the fault offset equivalent of it (Stone and Pufhal, 1995).

Clastic sedimentary rocks consist of fine- to medium-grained sandstone and conglomerate. Conglomerates are massive and poorly sorted, within a fine-grained matrix. Gravity flow processes such as debris flows are most likely responsible for their deposition. Cohesive debris flows tend to deposit poorly sorted, massive beds rarely showing grading, with a fine-grained matrix (Reading, 1996).

The fine- to medium-grained sandstone facies present in the Norway Lake area are interpreted as $T_{a,d,e}$ and $T_{d,e}$ turbidity current deposits. Low-density turbidity currents are responsible for their deposition.

Chemical sedimentary rocks present in the area are dominated by carbonates, chert and magnetite. Chemical precipitation from saturated seawater is a likely mechanism for deposition of the carbonates. Layering within the carbonates is possibly evidence of stromatolites. The chert and magnetite are again considered chemical precipitates with a hydrothermal source. Venting fluids from black and white smoker fields caused oversaturation of iron and silica in the water column (Fralick, 1987), resulting in precipitation of chert and magnetite.

The clastic sedimentary sequence in the Norway Lake area may be considered an extension of the depositional system operating in the Finlayson Lake area. It is clear that turbidity currents were the major depositional process operating in this area, as well as the Finlayson Lake area. Clastic sedimentary rocks were deposited within the basin by low density-turbidity currents possibly originating from the deltaic system active in the Finlayson Lake area. Conglomerate may reflect debris flows, coming from a localised source entering the basin in the Norway Lake area. Alternatively, the deltaic system operating in the Finlayson Lake area may not have been sufficiently large, and may not have supplied sediment into the Norway lake area. It is possible that a second deltaic system, centred in the Norway Lake area could have been operating, and was the sediment source for the clastic sedimentary sequence found in the area.

As the basin subsided clastic sedimentation waned and chemical precipitates became dominant. Carbonates were deposited within the shallow regions while chert and magnetite were deposited in the deeper waters of the basins.

8.3 Pinecone and Cryderman Lakes Area

The sedimentary rocks within the Cryderman and Pinecone Lakes area are part of the Upper Sedimentary Belt within the Lumby Lake Greenstone Belt and are laterally continuous with those in the Norway Lake area to the west, and those present in the Keewatin and Hematite Lakes area to the east. Clastic sedimentary rocks are finer-grained than those in the Norway Lake area, and consist of fine-grained sandstone and siltstone couplets. Chemical precipitates are abundant, including chert, magnetite, sulphides, and graphitic slate.

Turbidity currents are the sedimentary process which deposited the fine-grained sandstone and siltstone couplets. These sedimentary rocks are interpreted as T_d and T_e divisions of low-density turbidity current deposits, originating to the west.

Chemical sedimentary rocks are again interpreted as chemical precipitates settling from the water column, with a hydrothermal vent source for the silica and iron (Fralick, 1987). Alternation between black smoker and white smoker activity dominating in the area at any one time may cause the alternation between chert domination and magnetite domination.

Isolated lenses of graphitic slate and sulphides are considered to be more vent proximal. Local areas with Eh's low enough that sulphides were the preferred Fe-bearing phase may have existed near vents, or within topographic lows nearby on the ocean floor.

8.4 Keewatin and Hematite Lakes Area

The sedimentary succession exposed in the Keewatin and Hematite Lakes area is laterally continuous with the sedimentary rocks found in the Pinecone and Cryderman Lakes area to the east. These sedimentary rocks are the eastern-most extension of the Upper Sedimentary Belt, within the Lumby Lake Greenstone Belt. Clastic sedimentary rocks are secondary in abundance to the chemical sedimentary rocks, and consist dominantly of fine-grained sandstone to siltstone and shale couplets. Chemical precipitates, which dominate the sequence, are highly variable and consist of chert, magnetite, carbonate, sulphides and graphitic slates.

The fine-grained sandstone to siltstone and shale couplets are the product of the same low-density turbidity currents which deposited sediment within the other areas to the west, in the Upper Sedimentary Belt. These sedimentary rocks are interpreted as T_d and T_e turbidity current deposits. The thin bedded and fine-grained nature of these sedimentary rocks suggest that they are distal to the source area and were possibly deposited in deeper water.

A green, chloritic mudstone facies is found interbedded with the chemical sedimentary rocks in the area. This is interpreted as hemipelagic sediment, possibly fine-grained, wind blown ash from distant mafic volcanic eruptions.

Many facies of chemical precipitates are present in the area. A hydrothermal source is again suggested. Fluctuations in Eh, pH and ionic concentrations are responsible for the variation in the facies which were deposited. Sulphides were likely deposited in more reducing environments, possibly nearer to the sources of venting fluids, while magnetite, chert and carbonates were deposited more distal to the venting fluids. The carbonates in the Keewatin and Hematite Lakes area have a higher Fe/Ca ratio than those found to the west in the Norway Lake and Pinecone-Cryderman Lakes area which may reflect deposition from water with a higher Fe^{+2} / Ca^{+2} ratio. Alternatively this may be a diagenetic effect. The lack of stromatolitic structures in the rocks and their higher Fe/Ca ratio may suggest deeper water deposition.

The influx of clastic sediments marks the end of the long-lived period of mafic volcanism, whose products underlie the sedimentary sequence in the Lumby Lake Greenstone Belt. With active delivery of sediment into the basin, originating from a centre to the south of Finlayson Lake, fine-grained sandstone to siltstone and shale were deposited in this area by the final stages of turbidity currents. The shale tops of these beds are the product of hemipelagic sedimentation between successive turbidity currents. With time clastic influx waned and chemical precipitates became the dominant sediments deposited in the area.

CHAPTER 9

DISCUSSION & CONCLUSIONS

Several conclusions can be drawn regarding the Finlayson and Lumby Lake Greenstone Belts.

Within both the Finlayson Lake and Lumby Lake Greenstone belts two distinct sedimentary sequences are present. Each of the belts contain an upper and lower sedimentary sequence which differ in age and geochemical composition.

The lower sequence in the Finlayson Lake Greenstone Belt is represented by the Little Falls Lake sedimentary rocks and the laterally equivalent lower Finlayson Lake sedimentary rocks. These sedimentary rocks consists of coarse-grained sandstones, conglomerates, and lesser interbedded mafic detritus rich sedimentary rocks, and are laterally continuous with felsic volcanic rocks to the south. Deposition of these sedimentary rocks was by high-density turbidity current processes. Their geochemical signature is distinct from that of the main Finlayson Lake sedimentary rocks and suggests a single felsic volcanic source with composition similar to that of the Steep Rock Upper Felsic unit and the Old Tonalite unit. U/Pb geochronology again supports a single source rock, with an age of 2996.9 +/- 0.8 Ma.

The main band of Finlayson Lake sedimentary rocks are distinct from the sedimentary rocks present in the Little Falls Lake area and lower Finlayson Lake areas. Their geochemistry suggests that the upper Finlayson Lake sedimentary rocks are similar and likely continuous with, although fault offset from the upper Lumby Lake sedimentary rocks (Stone and Pufahl, 1995). U/Pb data from conglomerate in the southern Finlayson Lake area give zircon ages ranging from 2997.0 ± 2.5 to 3002.3 ± 0.9 Ma. Sm/Nd data suggests that the basin received detritus derived from tonalitic rocks as well as a slightly older mafic volcanic component. This data agrees with geochemical data which suggests that the composition of the upper Finlayson Lake sedimentary rocks lies on a mixing line between the Old Tonalite and the Steep Rock Upper Mafic unit or the Finlayson Lake mafic volcanics.

A well developed coarsening upward sequence is developed within the upper Finlayson Lake sedimentary rocks. The sequence consists of iron formation and chemical sedimentary rocks at the base, overlain by DE turbidites which coarsen to pebbly sandstones and conglomerates near the top of the sequence. There is also some lateral facies variation with coarsest-grained sedimentary rocks found in the southern part of Finlayson Lake. These sediments are consistent with deposition from both high- and low-density turbidity currents and were likely deposited by a prograding delta system that was centred south of the area.

As in the Finlayson Lake Belt, the Lumby Lake Belt also contains two stratigraphically distinct sedimentary bands. The lower sedimentary band is represented by the sedimentary rocks present within the Hook Lake area, while the upper sedimentary band is represented by the sedimentary rocks near Norway Lake and west to the Keewatin-Hematite Lakes area.

The lower Lumby Lake sedimentary rocks are laterally continuous with 2999 Ma old felsic volcanic rocks to the east (Jackson, 1985), and are likely the resedimented equivalent of them. Their geochemical signature is similar to that of the Little Falls Lake sedimentary rocks, and a geochemically similar source is suggested.

The upper Lumby Lake sedimentary sequence is similar to the upper Finlayson Lake sequence and is likely the fault offset equivalent (Stone and Pufhal, 1995). The geochemical signature of the sedimentary rocks suggests that the upper Lumby Lake sedimentary rocks had source rocks of the same composition as the source of the upper Finlayson Lake sedimentary rocks. The upper Lumby Lake sequence is dominated by iron formations and chemical precipitates, with lesser fine-grained clastic sedimentary rocks. There are lateral facies variations from more clastic dominance in the west to chemical precipitate dominance in the east. The predominance of chemical precipitates is evidence of widespread hydrothermal activity throughout the area. It is possible that the upper Lumby Lake clastic sedimentary rocks represent the distal equivalent of the turbidite system developed in the Finlayson Lake area. Alternatively the upper Lumby Lake portion of the basin may have been fed by a localized source centred in the Norway Lake area. A dominance of clastic sedimentary rocks and the presence of debris flow conglomerates may be evidence of this.

A possible paleogeographic reconstruction can be seen in Figure 9.01. After an initial period of mafic volcanism, which produced the lower portion of the sequence developed in the Finlayson and Lumby Lake Greenstone Belts, localized felsic volcanism occurred between 2999 and 2997 Ma. Subaqueous volcani-clastic sediment wedges developed around volcanic centres, and iron formation and carbonates were precipitated in areas sheltered from clastic input in both the Little Falls and Hook Lake areas. Mafic volcanism continued after this short lived period of felsic volcanism and tonalitic melts intruded the upper portions of the volcanic pile. Iron formations were deposited during periods of volcanic quiescence. Another period of tonalitic intrusion occurred at around 2930 Ma and uplifted the area. Volcanism ended at approximately this age and a period of quiescence allowed thick accumulations of chert and iron formations and carbonates to develop. Erosion of the tonalitic-mafic source terrain began to feed sediment into the basin that had developed in the Finlayson and Lumby Lakes areas. A prograding deltaic sequence developed, overlying the iron formation and carbonates previously deposited in the area. After an initial influx of sediment the basin subsided, and a return to deeper water sedimentation is marked by the fine-grained sedimentary rocks capping the deltaic sequence.

The deltaic system which deposited the upper sedimentary sequence in the Finlayson and Lumby Lake Greenstone Belts is consistent with the sedimentary sequence of the Steep Rock Lake Group of similar age. The conglomerates and sandstones of the basal Wagita Formation have been

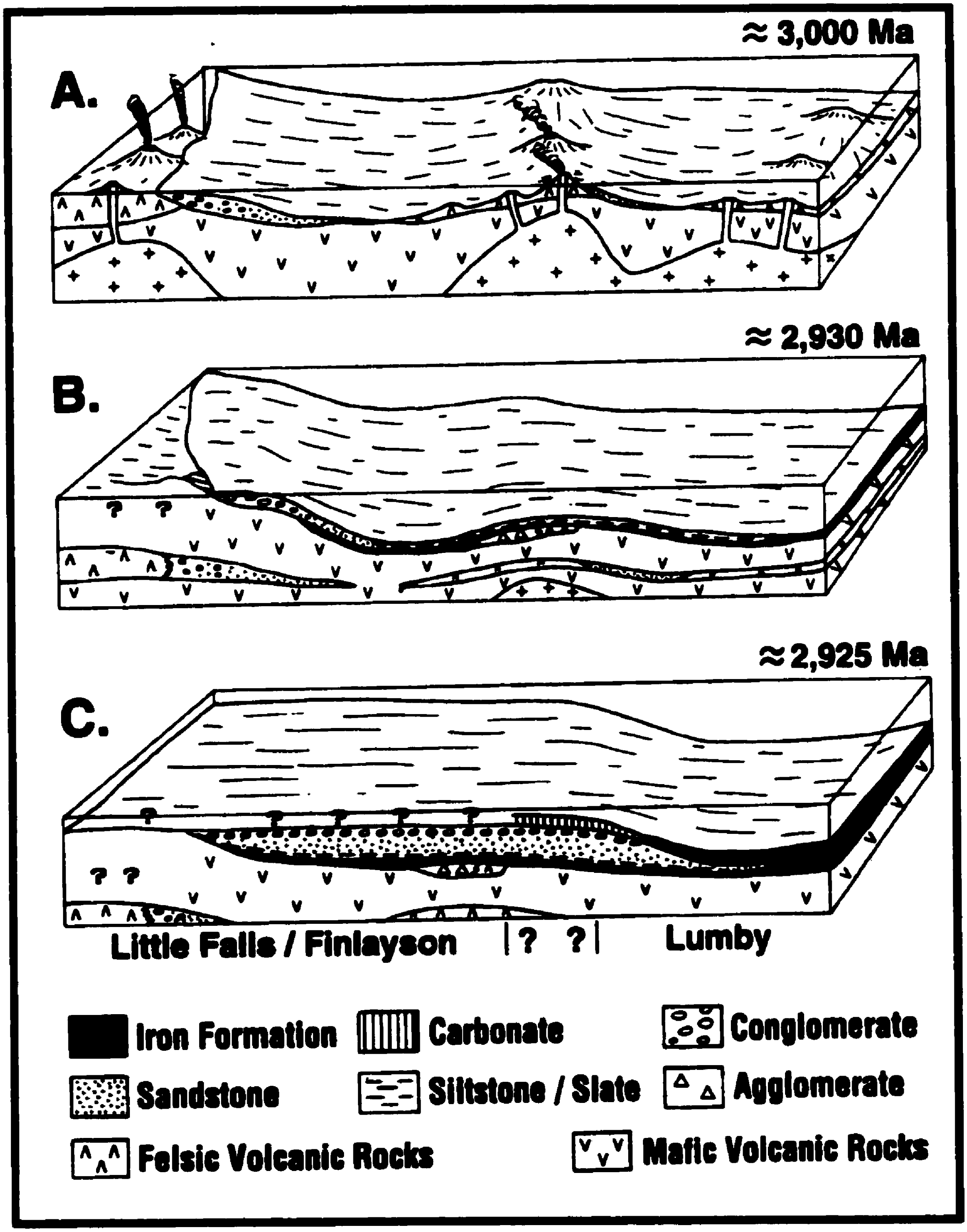


Figure 9.01 Possible paleogeographic reconstruction of the Finlayson and Lumby Lake Greenstone belts through time. Refer to text for explanation.

interpreted as erosive channels, cut into the underlying Marmion Complex (Wilks and Nisbet, 1988). The clasts within the conglomerate are of similar composition to those present within conglomerates of the Finlayson Lake area (see Provenance chapter). The Wagita Formation may represent the more proximal deposits of the sedimentary sequence developed in the Finlayson Lake area. The conglomerates represent the feeder channels, which carried sediment to the delta forming near Finlayson Lake. As the basin subsided, relative sea level rose, and the channel system back filled. Sediment supply to the Finlayson Lake area was cut off, and fine-grained sandstones and siltstones were deposited above the coarse-grained sandstones and conglomerates representing the delta sequence. With further subsidence of the basin, carbonate and iron formation secceeded fluvial deposition in the Steep Rock Lake area and formed reef platforms (Wilks and Nisbet, 1988).

This interpretation is consistent with the sedimentary rocks found in the Finlayson and Lumby Lake Greenstone Belts, as well as those in the Steep Rock Lake sequence, and relates all of the coeval sequences to a single depositional systems tract.

REFERENCES

Argast, A., Donnelly, T.W., 1987. The chemical discrimination of clastic sedimentary components. J. Sediment. Petrol. 57, 813-823

Ayres, L. D., 1977 Importance of stratigraphy in early Precambrian volcanic terrains: cyclic volcanism at Setting Net lake, northwestern Ontario. In W.R.A. Barager, L.C. Coleman, and J. M. Hall (eds) Volcanic regimes in Canada, Geol. Assoc. Canada Spec. Paper 17, p243-264.

Ballance, P. F., 1988. The Huriwai braidplain delta of New Zealand: a late Jurassic, coarse-grained, volcanic fed depositional system in a Gondwana forearc. In: Fan Deltas: Sedimentology and Tectonic Settings. (Ed. By W. Nemec and R. J. Steel), pp 430-444. Blackie and Son, London.

Ballard, R. D., Francheteau, G., Juteau, T., Rangan, C., and Normark, W., 1981. East Pacific Rise at 21°N: the volcanic, tectonic, and hydrothermal processes of the central axis. Earth and Planetary Science Letters, 55, p.1-10.

Barrett, T.J., Cattalani, S., Chartrand, F., Jones., P., 1991. Massive sulphide deposits of the Noranda Area, Quebec. II. The Aldermac Mine. Can. J. Earth Sci. 28, 1301-1327

Barrett, TJ., MacLean, W.H., 1991. Chemical, mass and oxygen isotope changes during extreme Hydrothermal alteration of an Archean rhyolite, Noranda, Quebec. Econ. Geol. 86, 406-414

Bhatia, M. R. 1983. Plate Tectonics and geochemical composition of sandstones; Journal of Geology, v. 91, p 611-627

Bickle, M. J., Bettenay, L. F., Boulter, C. A., Groves, D. I. And Morant, P. (1980). Horizontal tectonic interactions of an Archean gneiss belt and greenstones, Pilbara block, Western Australia. Geology, 8, p525-529.

Bickle, M.J., Morant, P., Bettenay, L. F., Boulter, C. A., Blake, T. And Groves, D. I. 1985. Archean Tectonics of the Shaw batholith, Pilbara block, western Australia: structural and metamorphic Tests of the batholith concept. In L. D. Ayres, P. C. Thurston, K. D. Card, and W. Weber (eds), Evolution of Archean supracrustal sequences, Geol. Assoc. Canada Spec. Paper, 28, p325-341.

Bouma, A. H. 1962 Sedimentology of some flysch deposits: a graphic approach to facies interpretation. Elsevier, Amsterdam, 168p.

Collins, W. J., 1989. Polydiapirism of the Archean Mount Edgar Batholith, Pilbara Block, Western Australia. Precambrian Research, 43, p41-62.

Corfu, F. and Aryes, L. D., 1991 Unscrambling the stratigraphy of an Archean greenstone belt: a U-Pb Geochronological study of the Favourable Lake belt, northwestern Ontario, Canada. Prec. Res. 50, p201-220.

Davis, D. W., and Jackson, M. C., 1988 Geochronology of the Lumby Lake greenstone belt: A 3 Ga complex Within the Wabigoon subprovince, northwest Ontario. Geol. Soc. of America Bull. V. 100, p 818-824.

Deer, W. A., Howie, R. A. and Zussman, J., 1967 An Introduction to the Rock-formining Minerals. Longman and Green, London, 528pp.

De Witt, M. J., and Ashwal, L. D. 1997 eds. Greenstone Belts, Clarendon Press, Oxford

Cullers, R. L., Chaudhuri, S., Arnold, B., Moon, L. and Wolf, C. W., 1975 Geochim cosmochim Acta v. 39, 1691-1703

Fenwick, K. G., 1976. Geology of the Finlayson Lake area, District of Rainy River; Ontario Division Of Mines, Geoscience Report 145, 86p.

Fralick, P., 1987. Depositional environment of Archean iron formation: inferences from layering in sediment and volcanic hosed end members; in Precambrian Iron Formations, P. Appel and G. LaBerge, editors, Theophrastus Publications, S. A., Athens, Greece, p. 251-266.

Fralick, P.W., Kronberg, B.I., Geochemical discrimination of clastic sedimentary rock sources. Sedimentary Geology 113(1997) 111-124

Garrels, R.M., E.A. Jr. & MacKenzie, F.T., (1973) Genesis of Precambrian iron-formations and the Development of atmospheric oxygen. Econ. Geol., 68, 1173-1179

Goodwin, A.M., 1973: Archean iron- formations and tectonic basins of the Canadian Shield. Econ. Geol., vol. 68, p. 915-933

Gross, G.A., 1965, Geology of Iron Deposits in Canada: General Geology and Evaluation of Iron Deposits: Geol. Surv. Can. Econ. Geol., Rep. 22: v.1, p.181

Hollings, P., Wyman, D., Polat, A., and Kerrich, R., 1996. Trace Element Geochemistry of the Lumby Lake Greenstone Belt, Western Wabigoon Subprovince, N. Ontario. Abstract, GAC MAC, 1996.

Huber, N.K., (1959) Some aspects of the origin of the Ironwood iron-formations of Michigan and Wisconsin. Econ. Geol., 54, 82-118

Jackson, M. C., 1985. Geology of the Lumby Lake area, western part, Districts of Kenora and Rainy River; Ontario Geological Survey, Open File Report 5534, 151p.

Jolliffe, A. W., 1955: Geology and iron ores of Steep Rock Lake; Economic Geology, v50, no. 4, p373-398. 1966 Stratigraphy of the Steep Rock Group, Steep Rock Lake, Ontario; in The relationship of mineralization to Precambrian Stratigraphy in certain mining areas of Ontario and Quebec, Precambrian Symposium, edited by A.M. Goodwin, Geological Association of Canada, Special paper No.3, p. 75-98.

King, D. and Fralick, P. 1995 Depositional systems associated with the 3.0 Ga Finlayson and Lumby Lake Greenstone belts, Northwestern Ontario. Abstract, Lake Superior Institute Meeting, May 1995

Kneller, B. C. and Branney, M. J., 1995. Sustained high-density turbidity currents and the deposition of thick massive sands. Sedimentology, v 42, p 607-616.

Kusky, T. M., 1989 Accretion of the Archean Slave Province, Geology, 17, p63-67.

Kusky, T. M., 1990 Tectonics of a late Archean arc/continent collision in the Slave Province, Northwest Territories, Canada. In Third International Archean Symposium, edited by J.E. Glover and S. E. Ho, etended abs. Vol., Geoconferences (W.A.) Perth, p453-456.

Kusky, T. M., and Vearncombe, J. R. 1997, Structural aspects: Greenstone Belts Dewitt, M. J. and Ashwall L. W. eds. Clarendon Press, Oxford

Lawson, A. C., 1912. The Geology of Steeprock Lake, Ontario; Geological Survey of Canada. Memoir 28, p 7-15.

Lowe, D. R. 1982. Sediment gravity flows: II. Depositional models with special reference to the deposits of high density turbidity currents; Journal of Sedimentary Petrology, v. 52, no. 1, p. 279-297.

MacLean, W.H., 1990. Mass change calculations in altered rock series. Miner. Deposita 25, 44-49

MacLean, W.H., Hoy, L.D., 1991. Geochemistry of hydrothermally altered rocks at the Horne Mine, Noranda, Quebec. Econ. Geol. 86, 506-528

MacLean, W.H., Kranidiotis, P., 1987. Immobile elements as monitors of mass transfer in hydrothermal alteration: Phelps Dodge massive sulphide deposit, Matagami, Quebec. Econ. Geol. 82, 951-962

Marshack, S., Alkmim, F., and Jordt-Evangelista, H., 1992 Proterozoic crustal estension and the generation Of dome- and-keel structure in an Archean granite-greenstone terrane . Nature, 357, 491-493.

Middleton, G.V., 1967, Experiments on density and turbidity currents, III. Deposition of sediment: Canadian Jour. Earth Sci., v.4, p. 475-505

McConachy, T. F., Ballard, R. D., Motte, M. J., and von Herzen, R. P., 1986. Geologic form and setting of a hydrothermal vent field at lat. 10°56'N, East Pacific Rise: a detailed study using Angus and Alvin. Geology, 14, p. 295-298.

Nesbitt, H.W., 1979. Mobility and franctionation of rare earth elements during weathering of a Granodiorite. Nature 279, 206-210

Nesbitt, H. W., Mackovics, G. and Price, R. C., 1980 Cemical processes affecting alkalis and alkaline Earths during contental weatherin. Geochim. Cosmochim Acta, v. 44, p 1659-1666.

Normark, W.R., 1970, Growth patterns of deep-sea fans: Am. Assoc. Petroleum Geologists Bull., v. 54, p. 2170-2195

Parkinson, R. N., 1962. Operation Overthrust; p. 90-101 in the Tectonics of the Canadian Shield, Roy, Soc. of Canada, Special Publications, No. 4, 180p.

Reading, H. G. 1986, 1996. Sedimentary Environments and Facies; Blackwell Science Ltd. Osney Mead, Oxford, 688p.

Roser, B. P. and Korsch, R. J. 1986. Determination of tectonic setting of sandstone mudstone suites using SiO2 content and K2O/Na2O ratio; Journal of Geology, v. 94, p. 635-650.

1988. Povenance signatures of sandstone-mudstone suites determined using discriminant function analysis of major element data; Chemical Geology, v. 67, p. 119-139

Sastri, G. G. K. and Sastry, G. S., 1982. Chemical characteristics and evolution of the laterite profile in Hasaridadar Bauxite Plateau, Madhya Pradesh, India; Economic geology, v. 77, p. 154-161.

Sawyer, E.W., 1986: The influence of source rock type, chemical weathering and sorting on the geochemistry of elastic sediments from the Quetico metasedimentary belt, Superior Province, Canada – Chem. Geol. 55: 77-95

Schau, M. and Henderson, J. B., 1983. Archean chemical weathering at three localities on the Canadian Shield; Precambrian Research, v. 20, p. 189-224.

Shklanka, R., 1972 Geology of the Steep Rock Lake area, District of Rainy River; Ontario Department of Mines and Northern Affairs, Geological Report 93, 114p.

Smyth, H. L., 1891. Structural geology of Steep Rock Lake, Ontario; American Journal of Science, v. 42, p 317-331.

Stauffer, P.H., (1967) Grain-flow deposits and their implications, Santa Ynez Mountains, California. J. sedim. Petrol., 37, 487-508

Stone, D., Kamineni, D. C., and Jackson M. C., 1992. Precambrian Geology of the Atikokan Area, Northwestern Ontario; Geological Survey of Canada Bulletin 405, 106p.

Stone, D., and Pufhal, P., 1995. Geology of the Atikokan-Sapawe Area: Regional Controls on Gold Mineralization in the Marmion Batholith. In: Summary of Field Work and Other Activities 1995, Ontario Geological Survey, Misc. Paper 164, p. 45-47.

Swager, C. and Griffin, T. J., 1990 An early thrust duplex in the Kalgoorlie-Kambalda greenstone belt, Eastern Goldfields Province, Western Australia. Precambrian Research, 48. P63-73.

Thurston, P. C., and Chivers, K. M., 1990. Secular variation in greenstone sequence development Emphasizing Superior Province, Canada; Precambrian Research; v. 46, p. 21-58

Tomlinson, K. Y., Hughes, D. J., and Thurston, P. C., 1996. Metavolcanic Rocks of the Central Wabigoon Subprovince: 1) The Lumby Lake Greenstone Belt., in: Summary of Field Work and Other Activities 1996, Ontario Geological Survey Misc. Paper 166, p. 60-63.

Uglow, W. L., 1913. Geology of the Vicinity of Steep Rock Lake; in: Guide Book No. 8, pt 1, 12th International Geological Congress, issued by the Geological Society of Canada, p. 46-53.

Walker, G.P.L., 1971. Grain size characteristics of pyroclastic deposits. Jour. Geol., Earth Sc., Vol. 79, p. 696-714

Walker, R.G. (1978) Deep water sanstone facies and ancient submarine fans: models for exploration for stratigraphic traps. Bull. Am. Ass. Petrol. Geol., 62, 932-966

Wilks, M. E., 1986. The geology of the Steep Rock Group, northwestern Ontario: a major Archean Unconformity and Archean stromatolites; M.Sc. thesis, University of Saskatchewan, Saskatoon Saskatchewan, 206p.

Wilks, M. E. and Nisbet, E. G., 1988, Stratigraphy of the Steep Rock Group, northwest Ontario: a major Archean unconformity and Archean stromatolites. Can. J. Earth Sci., v 25, p. 370-391.

Woolverton, R. S., 1960. Geology of the Lumby Lake Area. Ontario Department of Mines Vol LXIX, Part 5, 1960.

APPENDIX I

Geochemical Analysis

Sample #	2	7	8	9	10	11	16	20	23
Location	Little Falls	Little Falls	Little Falls	Little Falls	Little Falls	Little Falls	Little Falls	Little Falls	Little Falls
	m-c sst	m sst	m-c sst	m-c sst	m-c sst	m-c sst	m-c sst	c sst	c sst
%									
SiO2	67.35	68.67	65.70	75.69	66.64	74.22	66.67	65.31	67.21
TiO2	0.31	0.40	0.41	0.25	0.56	0.20	0.25	0.28	0.33
Al2O3	12.16	13.53	14.78	12.15	16.58	11.72	14.20	14.57	11.75
FeO T	4.05	3.98	4.09	3.08	5.30	2.75	4.28	4.25	3.90
MnO	0.02	0.05	0.05	0.03	0.05	0.03	0.05	0.05	0.03
MgO	1.59	2.24	2.15	1.01	2.11	0.41	2.21	1.63	1.37
CaO	0.25	3.01	3.30	1.74	1.41	1.15	2.48	3.34	1.84
Na2O	3.12	2.81	3.95	2.81	4.43	4.20	4.72	2.58	4.66
K2O	2.50	2.26	1.40	1.88	2.27	2.03	1.19	2.69	0.56
P2O5	0.07	0.07	0.08	0.05	0.10	0.02	0.07	0.07	0.08
H2O	1.98	1.62	1.53	1.53	2.07	0.63	1.80	2.16	1.35
CO2	0.07	2.24	0.81	1.36	0.77	0.84	2.05	2.60	1.54
Total	93.47	100.88	98.25	101.56	102.29	98.21	99.98	99.53	94.62
ppm									
Cr	16	58	38	b.d.	4	1	33	10	15
Ni	26	48	34	9	15	b.d.	40	22	25
Co	19	21	54	33	21	9	72	17	17
V	82	65	76	4	92	2	70	73	79
Cu	42	3	23	32	b.d.	5	36	15	6
Pb	66	66	77	67	89	62	67	74	60
Zn	50	63	64	55	59	37	65	53	41
Mo	1	3	2	2	1	3	b.d.	1	b.d.
S	74	15	68	178	64	56	91	b.d.	78
Ba	554	340	342	207	627	402	190	550	70
Sr	101	142	248	82	109	70	142	88	80
Nb	3.0	3.9	3.9	5.1	6.5	5.0	2.6	3.3	3.7
Zr	90	107	36	226	154	218	97	120	132
Y	5.9	7.9	7.4	18.9	16.3	20.1	6.5	8.9	9.5
B	13	26	23	14	21	17	24	15	23

Sample #	27	29	30	12	31	21	32
Location	Little Falls	Little Falls	Little Falls	Little Falls	Little Falls	Little Falls	Little Falls
	m-c sst	c sst	c sst	clast	clast	L. tuff	L. tuff
%							
SiO2	72.19	66.54	64.88	63.62	62.11	45.63	48.81
TiO2	0.22	0.43	0.39	0.53	0.56	0.91	0.56
Al2O3	12.80	15.01	14.78	13.66	13.78	15.23	13.34
FeO T	3.05	4.67	3.90	3.71	3.28	14.51	11.04
MnO	0.04	0.06	0.05	0.07	0.09	0.17	0.17
MgO	1.20	0.49	0.32	b.d.	b.d.	3.83	5.49
CaO	1.73	1.88	2.89	5.21	5.22	5.89	10.61
Na2O	3.40	5.44	6.06	5.15	5.44	2.72	1.31
K2O	2.12	1.01	0.59	0.09	0.10	0.46	0.03
P2O5	0.05	0.07	0.07	0.10	0.09	0.12	0.04
H2O	1.35	1.26	1.44	1.17	1.26	4.59	3.06
CO2	1.47	1.06	2.35	1.80	2.57	5.02	1.10
Total	99.62	97.93	97.72	95.11	94.50	99.09	95.57
ppm							
Cr	6	54	39	59	56	183	355
Ni	7	37	32	50	37	77	148
Co	10	46	39	61	44	73	90
V	26	68	58	77	71	268	226
Cu	2	-3	129	138	32	150	110
Pb	66	68	75	79	65	84	64
Zn	39	59	50	30	38	141	84
Mo	4	b.d.	b.d.	b.d.	b.d.	b.d.	b.d.
S	b.d.	232	141	1020	b.d.	884	1220
Ba	[illegible]	[illegible]	200	[illegible]	97	[illegible]	b.d.
Sr	44	152	237	319	274	127	b.d.
Nb	3.2	3.8	3.6	5.8	5.4	7.4	5.4
Zr	113	90	84	74	59	49	38
Y	10.1	7.9	7.6	14.0	12.4	13.0	14.6
B	13	52	64	48	59	92	118

Sample #	62	64	73	63a	63b	70	61
Location	Finlayson	Finlayson	Finlayson	Finlayson	Finlayson	Finlayson	Finlayson
	m sst	m sst	m-c sst	m sst	m sst	clast	m volcanic
%							
SiO_2	69.35	81.15	65.96	65.44	66.86	53.64	51.31
TiO_2	0.35	0.43	0.60	0.65	0.58	0.66	0.70
Al_2O_3	10.70	8.52	12.34	13.39	11.50	13.91	13.23
FeO T	5.31	6.67	6.87	7.51	7.38	9.37	10.89
MnO	0.20	0.10	0.13	0.09	0.07	0.20	0.16
MgO	0.42	2.42	1.26	3.25	2.47	2.06	5.51
CaO	6.55	0.85	2.76	2.23	1.39	7.87	10.31
Na_2O	2.20	1.46	2.25	1.88	1.46	2.37	1.71
K_2O	1.63	0.57	1.11	2.40	2.00	0.31	0.31
P_2O_5	0.05	0.03	0.10	0.09	0.08	0.04	0.05
H_2O	2.07	2.97	2.70	2.79	2.70	2.97	2.52
CO_2	5.10	1.17	2.68	3.08	1.43	4.51	0.18
Total	103.94	106.34	98.76	102.79	97.92	97.90	96.88
ppm							
Cr	67	73	193	153	115	319	341
Ni	34	58	86	78	68	123	115
Co	29	38	56	57	65	68	61
V	108	94	105	165	146	234	223
Cu	22	57	63	53	259	86	93
Pb	52	46	57	67	62	61	61
Zn	76	100	127	95	94	96	82
Mo	2	1	b.d.	4	3	b.d.	b.d.
S	16	968	3449	57	88	2680	b.d.
Ba	175	91	220	297	241	55	51
Sr	103	31	80	60	39	97	106
Nb	3.0	3.5	5.1	4.9	4.2	6.4	6.2
Zr	72	61	83	95	89	14	28
Y	6.1	5.0	9.5	6.1	4.8	16.2	13.5
B	19	16	62	27	19	44	92

Sample #	78	79	38	39	40	41	46	47
Location	lower Finlayson	lower Finlayson	Finlayson	Finlayson	Finlayson	Finlayson	Finlayson	Finlayson
	m sst	m sst	m-c sst	m-c sst	m-c sst	m-c sst	c sst	c sst
%								
SiO_2	72.25	74.12	78.81	60.93	59.86	69.41	70.24	72.38
TiO_2	0.37	0.36	0.39	0.65	0.34	0.40	0.26	0.54
Al_2O_3	13.86	13.68	8.05	14.10	10.78	12.33	8.95	11.70
FeO T	3.20	3.21	4.85	8.53	6.82	6.46	11.09	8.52
MnO	0.03	0.04	0.09	0.07	0.20	0.10	0.09	0.08
MgO	b.d.	b.d.	1.46	1.83	3.54	2.07	2.06	1.92
CaO	2.52	2.04	0.49	0.59	6.57	1.27	1.30	0.75
Na_2O	4.17	4.41	1.69	1.05	1.57	1.10	1.47	2.25
K_2O	1.21	1.63	0.90	2.45	1.67	2.12	0.22	0.96
P_2O_5	0.06	0.06	0.05	0.09	0.04	0.04	0.05	0.05
H_2O	0.63	0.45	0.90	3.33	1.17	2.16	3.15	2.70
CO_2	0.22	0.37	0.37	0.84	11.18	3.92	0.95	0.51
Total	98.53	100.36	98.04	94.46	103.73	101.38	99.83	102.35
ppm								
Cr	11	44	64	133	67	86	65	93
Ni	7	17	29	66	41	48	57	50
Co	45	31	31	38	40	29	67	46
V	19	18	83	203	125	142	99	131
Cu	81	5	20	29	17	26	41	2
Pb	66	63	41	72	54	66	49	64
Zn	29	43	51	78	79	79	62	49
Mo	b.d.	b.d.	1	2	4	b.d.	1	4
S	729	-6	280	635	136	415	1288	73
Ba	290	377	147	109	100	201	20	127
Sr	219	171	44	40	164	80	38	53
Nb	5.1	5.4	3.3	4.4	2.8	3.0	1.9	4.2
Zr	90	102	55	102	74	74	74	88
Y	15.0	16.8	5.2	3.1	5.9	4.3	4.0	5.4
B	17	24	17	18	25	12	27	18

Sample #	141	142	143	144	145	148
Location	lower Lumby m sst	Lower Lumby m sst	Lower Lumby m-f sst	Lower Lumby m-c sst	Lower Lumby m-c sst	Lower Lumby m-f sst
%						
SiO_2	65.85	78.05	46.82	81.43	74.40	47.91
TiO_2	0.30	0.41	0.45	0.14	0.23	0.72
Al_2O_3	14.64	8.52	9.57	5.33	6.88	13.98
FeO T	2.30	6.87	7.94	7.45	6.89	6.56
MnO	0.03	0.07	0.15	b.d.	0.11	0.16
MgO	b.d.	b.d.	9.76	b.d.	b.d.	2.43
CaO	2.73	0.11	7.13	b.d.	2.14	7.86
Na_2O	4.10	0.50	1.54	0.38	0.60	2.53
K_2O	2.25	1.00	0.04	1.08	1.18	1.58
P_2O_5	0.12	0.04	0.22	0.01	0.03	0.67
H_2O	1.26	2.34	3.87	0.72	1.17	2.16
CO_2	3.59	1.03	11.92	0.37	4.80	11.81
Total	97.16	98.95	99.41	96.91	98.41	98.38
ppm						
Cr	13	50	656	25	46	19
Ni	13	54	380	32	30	68
Co	30	44	56	75	38	32
V	25	50	111	11	23	88
Cu	19	60	35	35	35	53
Pb	67	38	49	32	35	69
Zn	32	100	70	14	38	113
Mo	b.d.	b.d.	2	b.d.	b.d.	b.d.
S	684	8397	238	55736	22532	3598
Ba	920	109	91	241	222	710
Sr	391	30	316	26	61	720
Nb	2.4	4.1	3.5	1.4	2.5	7.8
Zr	97	118	82	91	107	190
Y	4.4	10.4	7.8	4.6	7.4	19.4
B	49	25	80	4	11	70

Sample #	161	162	163	92	96	165
Location	Hematite	Hematite	Hematite	Norway	Norway	Steep Rock
	m-c sst	m-c sst	m-c sst	m sst	m sst	m-c sst
%						
SiO2	50.75	64.41	55.28	55.91	55.89	90.50
TiO2	0.87	0.31	0.64	0.52	0.61	0.20
Al2O3	13.84	16.62	15.02	12.75	9.71	4.81
FeO T	7.67	3.24	7.65	7.74	7.68	2.04
MnO	0.12	0.13	0.15	0.13	0.12	0.01
MgO	2.15	b.d.	5.35	8.02	5.05	b.d.
CaO	7.37	2.31	7.35	8.22	5.24	b.d.
Na2O	5.52	1.48	3.70	2.88	2.09	0.10
K2O	2.61	4.39	0.74	1.74	1.04	0.51
P2O5	0.47	0.10	0.25	0.19	0.20	0.04
H2O	1.26	1.98	1.35	0.99	2.16	1.35
CO2	6.23	1.76	1.10	0.59	1.83	0.37
Total	98.87	96.73	98.58	99.69	91.61	99.94
ppm						
Cr	98	29	371	408	394	48
Ni	35	6	170	305	224	21
Co	64	28	51	46	59	32
V	120	10	142	119	130	8
Cu	38	b.d.	27	24	27	30
Pb	68	74	68	67	45	19
Zn	98	26	89	90	61	11
Mo	b.d.	b.d.	b.d.	5	3	b.d.
S	4089	374	568	23	87	84
Ba	720	[illegible]	171	209	120	b.d.
Sr	1547	163	205	346	190	b.d.
Nb	7.9	2.6	6.3	5.1	5.1	1.6
Zr	118	72	134	90	102	42
Y	15.2	3.9	14.0	11.5	9.2	1.1
B	86	15	61	22	9	6