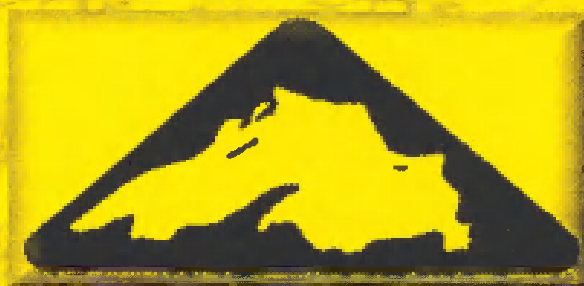


**FORTY-SIXTH ANNUAL MEETING
INSTITUTE ON LAKE SUPERIOR
GEOLOGY**

Thunder Bay, Ontario May 08-13, 2000



**PROCEEDINGS VOLUME 46
PART 2: FIELD TRIP GUIDE BOOKS**

46th I.L.S.G.

Field Guide

May 8-13, 2000



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**INSTITUTE ON
LAKE SUPERIOR
GEOLOGY**

PROCEEDINGS

Volume 46

Field Trip Guidebook

Compiled by: Philip Fralick

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GEOLOGY OF THE MESOPROTEROZOIC SIBLEY GROUP

**Philip Fralick, Lakehead University
Mark Smyk, Ontario Geological Survey**

This trip will examine the depositional architecture of a playa-alluvial succession developed in an intracratonic basin. The pristine sequence records the effects of both climate and tectonics on sediment character.

Geology and Stratigraphy of the Mesoproterozoic Sibley Group

by

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Amethystine stromatolite, Rossport Formation, Dorion, Ontario

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FOREWORD

This field guide is intended to provide a comprehensive overview of the Sibley Group sedimentary rocks, their stratigraphy and associated mineral deposits. Trip stops have been selected on the basis of representivity and accessibility; many other key and noteworthy sites cannot be easily accessed but are described in the text.

This guide follows in the footsteps of a number of previous field trips that have focused on the Proterozoic geology of the Thunder Bay area. Field stop descriptions have incorporated and built upon those found in guide books written by Franklin and Kustra (1972), Kustra *et al.* (1977), Franklin *et al.* (1982) and Kissin (1990), as well as other, unpublished reports.

Please contact the staff of the Resident Geologist's Office, Thunder Bay South District, (807) 475-1331, for current information regarding permitted access to private and staked properties.

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INTRODUCTION

Sedimentary rocks of the Sibley Group discontinuously outcrop on the north shore of Lake Superior and around Lake Nipigon (Figure 1). The flat-lying to gently dipping, clastic-carbonate succession occupies a broad oval area disconformably to unconformably overlying Mesoproterozoic, Paleoproterozoic and Neoarchean rocks.

Sibley Group rocks are located in the Southern Province of northwestern Ontario. The first documentation of these red beds was by Logan (1863). Hunt (1873) divided Logan's "Upper Copper-Bearing Series" into the Animikie and Keweenaw groups; the Sibley was assigned to the latter. Other descriptions of the Sibley Group were provided by Bell (1870), Parks (1901) and Wilson (1910) who compared the Sibley (then called "Nipigon Series") to similar Keweenawan and Paleozoic rocks elsewhere. Despite similarities to these younger rocks, the Sibley Group rocks were different enough to preclude assigning them a definite Precambrian age.

Subsequently, they were given the name "Sibley Group" by Tanton (1931) after the Sibley Peninsula on which they are well-exposed. The Sibley Group rocks were studied by Hawley (1930) and Moorhouse (1960), but they were not mapped comprehensively until Coates (1972). McIlwaine (1971a,b; 1975) also studied the Sibley Group rocks. Franklin (1970) pursued a doctoral thesis regarding the metallogeny of Proterozoic rocks in the Thunder Bay District, including the Sibley Group. Thereafter, he described the stratigraphy and depositional setting of the Sibley Group (Franklin *et al.* 1980). The tectonic setting, source areas, and environments of deposition of Keweenawan strata in the Lake Superior region were published by Morey and Ojakangas (1982). A doctoral thesis by Cheadle (1985) was executed on the "lower Keweenawan" Sibley Group. In all of these studies the Sibley Group rocks are presented as being genetically related to the Keweenawan Rift. This notion was refuted by Fralick and Kissin (1995).

Outcrop exposure and drill core lithologies provide evidence that the Sibley Group sedimentary rocks are continuous, near-horizontal strata there were deposited in an extensive, low-lying area. For this reason, the structure in which the sediments accumulated will be called the Sibley Basin. The Sibley Basin is elongate in a northwesterly direction, extending from the north shore of Lake Superior in the south, along the western margin of Lake Nipigon, to Armstrong in the north. It crops out discontinuously within a 15 000 km² area (Franklin *et al.* 1980). The exposed basin is now 175 km wide and 400 km long.

The fluvial and lacustrine strata of the Sibley Group are divided into three subhorizontal formations: 1) the lowermost is the Pass Lake Formation; 2) the Rosspoint Formation overlies it, followed by 3) the Kama Hill Formation.

The Pass Lake Formation varies from 0 to 50 m in thickness and consists of basal conglomerate and upward-thinning beds of quartz arenite. It was deposited in a shallow lacustrine environment (Franklin *et al.* 1980). The 135 m thick Rosspoint Formation is composed of a lower, arenaceous, red dolomite unit, a central chert-carbonate and stromatolitic unit, and an upper, argillaceous, red dolomite unit. These beds were laid down in a shallow, saline basin of fluctuating size (Franklin *et al.* 1980). Fifty metres of purple shale constitute the Kama Hill

Formation. Desiccation cracks, evaporite clasts, and mud chip conglomerates in this purple shale were formed on mudflats that were subaerially exposed and dried out periodically (ibid).

The lack of fossils in the Precambrian presents problems in age determination and in interpreting the environment of deposition. The stromatolites that are present in the Sibley Group rocks (Hoffman 1969) can occur in both marine and fresh water, but the red beds are characteristic of a continental setting (Morey and Ojakangas 1982). The thick dolomite and gypsum sequences formed during intermittent, arid periods.

Historically, it has been postulated that the Sibley Basin (1339 $\pm$ 33 Ma, to 1537 Ma; Franklin 1978a; Davis and Sutcliffe 1985, respectively) developed because of crustal attenuation caused by the Keweenaw Midcontinent Rift System. Franklin *et al.* (1980) described it as a failed arm, radial to a palaeoplume in relation the Keweenaw Rift. Morey and Ojakangas (1982) described a sedimentary-tectonic framework for the Sibley Basin that is related temporally and spatially to the Keweenaw Rift; the Sibley Basin developed as renewed subsidence of the Animikie Fault Block Basin (2200-2000 Ma). Conversely, Cheadle (1986) described the tectonic framework of the Sibley Basin as broad, subsiding sag where significant fault control is lacking at the margins. It was thought that the Sibley Group red beds are representative of the earliest evidence of the Keweenaw Rift Event. However, this interpretation is problematic.

The oldest igneous activity that is related to rifting is dated at $\sim$ 1109 Ma (Davis and Sutcliffe 1985) a minimum of $\sim$ 230 My after the Sibley Group was deposited. Conceptually, a thermal dome precedes rifting, and erosion will occur. It is reasonable to presume that volcanism can be preceded by sedimentation on this time scale, but the palaeodrainage system will develop away from the dome. This conflicts with the palaeocurrent work done by Cheadle (1986) in which the palaeodrainage system flowed toward the thermal welt. It is unreasonable to hypothesize that the pre-rift basin was overlying the thermal welt. Thus, the presumed relationship of the Sibley Basin with the Keweenaw Rift must be reappraised. The two events are not related directly (Fralick and Kissin 1995). Alternatively, the Sibley Basin may be genetically linked to the heating event represented by the igneous complex formed at Redstone Point, Lake Nipigon (1537 Ma; Davis and Sutcliffe 1984) as these igneous rocks lay directly under the Sibley Group Basin (Fralick and Kissin 1995). In this scenario, mantle upwelling would cause partial melting of the lower crust, forming the Redstone Point felsic magmas. The higher heat flow would cause thermal doming with erosive stripping of the dome. Upon relaxation of the crust an intracratonic basin, the Sibley Basin, would be created. This basin would be saucer-shaped without pronounced boundary faults. This corresponds to the actual basin geometry.

REGIONAL GEOLOGY

The Sibley Group rocks rest unconformably on the Paleoproterozoic Animikie Group (Rove and Gunflint Formations) and Neoarchean granitic and volcanic rocks (Morey and Ojakangas 1982). The Sibley Group possibly correlates with the Puckwunge Formation, a series of trough cross-stratified, medium-grained, lentic sandstones occurring in northeastern Minnesota (Mattis 1972). This formation unconformably overlies Paleoproterozoic Rove Formation shales and

represents South Saskatchewan River-type, braided fluvial channel deposition, exhibiting strong, unimodal flow to the southeast.

The Sibley Group consists of a lower basal conglomerate and sandstone unit, the Pass Lake Formation, a middle unit of red and buff dolomitic mudstone, the Rossport Formation, and an upper unit, the Kama Hill Formation, consisting of purple shale and buff-coloured siltstone to fine sandstone. Cheadle (1986) has subdivided the Sibley Group according to the different sedimentary facies. From bottom to top:

Pass Lake Formation	Loon Member (basal conglomerate facies)
	Fork Bay Member (plane-bedded facies and cross-bedded facies)
Rossport Formation	Channel Island Member (cyclic facies and mudstone facies)
	Middlebrun Bay Member (stromatolite facies)
	Fire Hill Member (chalcedonic mudstone facies)

Kama Hill Formation is constituted wholly by a laminated shale facies and thus, is not subdivided.

PASS LAKE FORMATION

The *Loon Lake Member* is best exposed at Pass Lake in a cliff, consisting of the Pass Lake Formation, which runs parallel to the Canadian National Railway. At the north end of the outcrop, the Paleoproterozoic Rove Formation is visible where the lowermost beds of the Pass Lake Formation disconformably overlie it. The Pass Lake Formation ranges in thickness from 18 to 50 m. The Loon Member, the basal conglomerate facies, consists of thin, lensoid conglomeratic sediment. It is thickest (~15 m) near the basin margin and it thins down to about one metre in the basin center (Cheadle 1986). Animikie taconite, jasper, and shale clasts predominate in the conglomerate deposited over Animikie rocks whereas vein quartz vein, granite, and metavolcanic rock clasts dominate where an Archean basement is present (Cheadle 1986). The clasts are ≤ 30 cm in diameter (Morey and Ojakangas 1982) and are in matrix-support (Cheadle 1986). The polymictic matrix is brick-red (Franklin *et al.* 1980) and it is composed of medium to coarse grains of quartz, hematite, and mudstone (Cheadle 1986). The conglomerate has calcareous cement. The lenses have scour and fill structures with deeper and narrower channels cut proximal to the source. They are cross-bedded with rare dolomite mudstone layers. At Silver Islet, the lensoids are graded with the lower half of the bed in clast support that changes into matrix support in the upper half (Cheadle 1986).

The *Fork Bay Member* is dominated by medium-grained sandstones which are subdivided into a lower plane-bedded facies and an upper cross-bedded facies. It overlies the Loon Member.

Plane Bedded Facies: 20 to 80 m of plane- to cross-bedded quartz arenite overlie the basal conglomerate (Cheadle 1986). This buff-coloured sandstone consists of well-sorted, rounded to

well-rounded, fine- to coarse-grained, mature quartz arenite to subarkose to sublitharenite (Morey and Ojakangas 1982). Minor chert and polycrystalline quartz fragments are present. As well, there are thinly interbedded dolomitic mudstone and siltstone layers that separate the 5 cm to 1 m thick sandstone beds. The upper 5 to 10 cm of some sandstone beds contain mud chips. The plane-bedded facies hosts an array of sedimentary structures including horizontal laminations and parting lineations, isolated planar cross-beds, straight- to sinuous-crested oscillation ripples, rare current ripple laminations, rare sand volcanoes and dewatering tubes, and desiccation cracks and halite clasts in thinly interbedded mudstone (Cheadle 1986).

Cross Bedded Facies: The cross-bedded facies is 5 to 10 m thick, with medium- to large-scale, cross-bedded arenite intercalated with 1 m thick planar, cross-stratified sandstone beds at Rossport, Red Rock, Channel Island, and Pass Lake. This facies contains scoured bases and pebbly infill (Cheadle 1986). The ripple-marked and mud-cracked quartz arenite beds have horizontal to wavy stratification with minor trough cross-beds. The palaeocurrents of the cross-beds at Pass Lake and the Enterprise Mine are scattered, with predominantly south-southeast and east-southeast directions, respectively (Morey and Ojakangas 1982).

Franklin *et al.* (1980) logged the 50 m cliff exposure at Pass Lake. Twenty metres of buff arenite rest on the basal lensoid conglomerate. This buff arenite thins upward from 1 m thick beds at the base to 1 to 2 cm thick beds at the top. The sequence alternates upwards from dolomite to arenite. They recorded 1 m of red dolomitic arenite to 13 m of buff arenite; 1 m of thinly bedded red dolomite and dolomitic arenite; 10 m of buff arenite; 1 m of red dolomitic arenite; and 7 m of thickly bedded buff arenite. A sharp increase in the dolomite content, a pigmentation change, and a gradual decrease in clastic content mark the upper boundary of the Pass Lake Formation (Franklin *et al.* 1980).

At Redstone Point, along the Lake Nipigon shoreline, the Pass Lake Formation overlies an anorogenic, felsic intrusive-extrusive granitoid suite (Fralick and Kissin 1996, unpublished data) dated at 1537 Ma (Davis and Sutcliffe 1985). One metre below the volcanic-sedimentary contact, the igneous rocks are weathered and the feldspar exhibits a chalky, cream to salmon colouration. There is a 30 cm thick volcanic breccia with caliche overlying the weathered igneous rocks. The Pass Lake Formation is lacking the basal conglomerate at Redstone Point. Instead, there are fine-grained, red sublitharenite to subarkosic sandstones with massive or parallel-laminated beds. Trough cross-stratification is rare. These strata contain very angular, granular to pebble to small cobble-sized, felsic volcanic clasts with oxidized rims and weathered feldspars. There are also less abundant clasts of hematitic mudstone, chert, and caliche pebbles. The basal pebble sandstone is poorly sorted with subrounded to rounded sand and small angular pebbles. It has carbonate cement with clay filling between the grains. The beds range from 1 to 80 cm thick, with a mean thickness of 15 cm. They are laterally continuous over several metres with rare, 1 m wide channels. Tightly packed quartz arenite overlies the basal pebbly sandstone. The quartz arenite is well-sorted, subangular to subrounded, contains minor clay in the matrix, and has quartz cement. The arenites are folded slightly and plunge at $<15^\circ$ to the east (Fralick and Kissin 1996, unpublished data).

ROSSPORT FORMATION

The Rossport Formation consists of red, fine-grained, arenaceous and clayey dolomite. It lies disconformably on the Pass Lake arenite (Franklin *et al.* 1980) and it has a maximum vertical outcrop of 134 m at its type locality where it is lacking the stromatolite marker bed (Morey and Ojakangas 1982). North of Nipigon and north of Dorion there have been 300 m and 180 m of the Rossport Formation drilled, respectively. Franklin *et al.* (1980) logged and subdivided the Rossport Formation, starting with a lower, argillaceous red dolomite unit. They recorded 1 m of red conglomerate at the base. Upwards through the sequence, there is 96 m of well-bedded, brick red dolomite and arenaceous dolomite intercalated with buff to white arenite, 1 to 2 m of black chert, gray chert, and buff calcite that alternate from 0.5 to 1 cm thick laminae with a low-amplitude crinkle structure. The sequence is topped with 37 m of massive red dolomite and arenaceous dolomite that fractures conchoidally. Spherical and irregular reduction spots are noted throughout the upper and lower units. The top of the Rossport Formation is marked by a change in the dolomite content as it decreases gradually to zero. The Rossport Formation rests unconformably on Archean rocks at Kama Bay. At this location, as well as at Pass Lake, Red Rock, and Moseau Mountain, the Rossport's basal, red conglomerate is not present.

Cheadle (1986) has divided the Rossport Formation into four facies from base to top:

Channel Island Member	(cyclic and mudstone facies)
Middlebrun Bay Member	(stromatolite facies)
Fire Hill Member	(chalcedonic mudstone facies)

Cyclic Facies: The cyclic facies of the Channel Island Member is 7 to 30 m thick. It has rhythmic layering of mudstone and dolomite couplets or mudstone and sandstone couplets. The couplets range from 5 to 160 cm in thickness. The sandstone and dolomite are end members that have various degrees of mud interlayering throughout the sequence. The dolomitic mudstone is red, massive, friable, and tabular. It is common to both cycles. The mudstone often contains silt or sand and rounded, reworked pebble-sized carbonate clasts (Cheadle 1986).

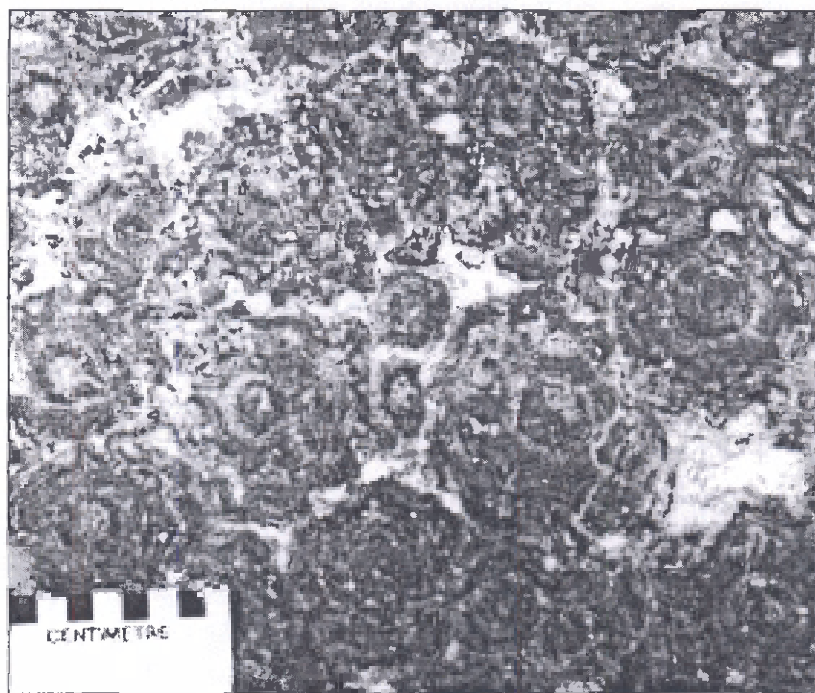


Red Rossport Formation mudstones with diagenetic evaporite minerals formed by precipitation in a clastic-dominated, sabkha setting

Mudstone Facies: The mudstone facies has

red and buff, dolomicritic mudstone and claystone that are dominantly massive and friable, but have local desiccation cracks and adhesion ripples, and small, scattered gypsum and anhydrite nodules. One to two metre thick fine- to medium-grained, poorly sorted, subarkose and sublitharenite with 10% matrix are sporadically introduced into the dolomicritic mudstone. These subarkose and sublitharenite beds have powering-down sequences: scoured bases, plane-bedded laminations, ripple cross-laminations, and horizontal, suspended load laminations (Cheadle 1986).

Stromatolite Facies: The stromatolite facies, or the Middlebrun Bay Member, which is located in the middle of the Rosspoint Formation, is 80 to 140 cm thick. The gray chert-carbonate, cryptalgal laminate is laterally extensive, but commonly brecciated. The base of this marker bed has alternating layers of light gray dolomite, dark gray chert, and crinkly laminae. Moving upwards, there is a thin chert-carbonate breccia, a layer of laterally linked and domical single parabolic stromatolites, and a dark gray chert band (Cheadle 1986).



*Plan view section of
stromatolitic marble
(Conophyton garganicus),
Wolatko Lake*

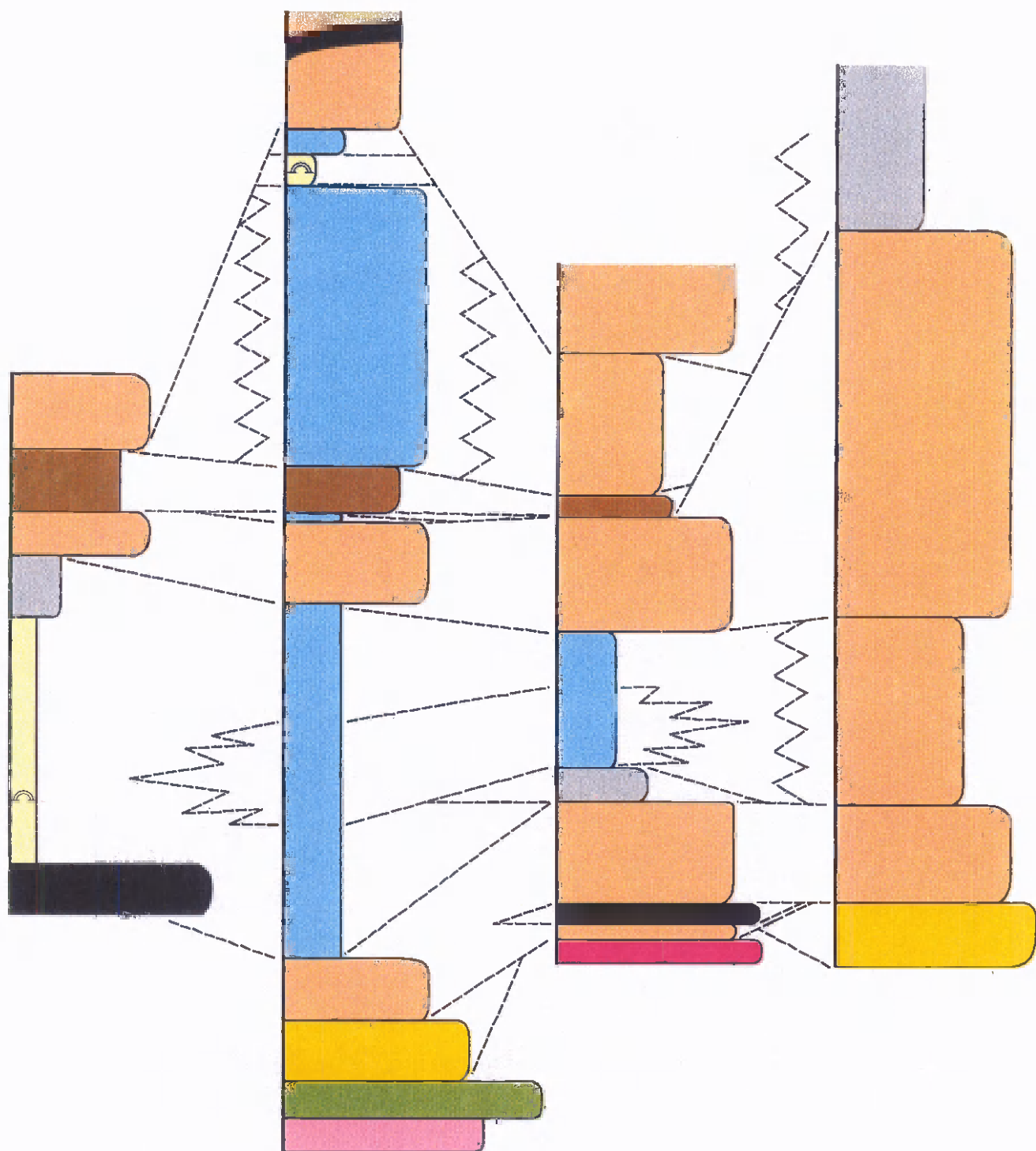
Chalcedonic Mudstone Facies: The upper Rosspoint Formation is composed of the chalcedonic mudstone facies. At the base, there are local lenticular, intraformational, paraconglomerate wedges that contain subangular mudstone clasts up to 2 m in diameter. Reduction spheres are common. Red and buff, dolomicritic mudstone and claystone are the dominant constituents of this sequence. They lack the coarse silt and sand that is present in the Channel Island mudstone and claystone. Dolomite partially replaces 2 cm diameter chalcedony nodules that are found in resistant weathering horizons.

KAMA HILL FORMATION

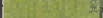
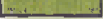
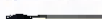
The Kama Hill Formation consists of a single (laminated shale) facies, and thus is not subdivided into constituent members. The original thickness of the Kama Hill Formation is unknown as the top is truncated by the Logan diabase sills. The Kama Hill laminated shale has a minimum thickness of 46 m (exposed at Kama Hill), and this varies, depending on the location. At Red Rock, Channel Island, and Stewart Lake it is 20 m thick, and at Albert Lake it is 45 m thick (Franklin *et al.* 1980). The horizontally laminated, buff arkosic siltstone to fine sandstone are exposed in a cliff face along the shore of Kama Bay, east of Nipigon (Franklin *et al.* 1980; Cheadle 1986). The base is transitional over 1 to 2 m and contains very little carbonate. Reduction spots are common, as are mud cracks and ripple marks on bedding surfaces (Franklin *et al.* 1980). Breccia dikes are present and mud-chip conglomerates are associated with the silt-dominated layers (Morey and Ojakangas 1982; Cheadle 1986). The siltstone layers contain the following sedimentary structures: erosional bases, ripple cross-laminations, low amplitude ripple marks, centimetre to decimetre fining-upward sequences, normal grading, and suspensate caps (Cheadle 1986) evaporite casts and raindrop impressions. This formation locally contains carbonate algal stromatolites at the base of the facies in the north half the basin (Morey and Ojakangas 1982; Cheadle 1986).



Enigmatic mudcracks(?) in distal bar sequences of the Kama Hill Formation



LEGEND

	Basal Conglomerate		Dolomite		Mud Chip Conglomerate
	Sandstone		Gypsum		Magnetic Granite
	Siltstone		Stromatolites		Gneiss
	Shale		Diabase		

DEPOSITIONAL SYSTEM

The Sibley Group sediments were laid down in a broadly subsiding, ovoid basin during the Mesoproterozoic. There is no evidence of high relief along any exposed segments of the basin margin. At the time of deposition, the Sibley Basin was situated very close to the equator (unpublished paleomagnetic results, Dr G. Borradaile) and was subjected to an arid to semiarid climatic regime. These factors combined to dictate the types of depositional systems which developed during the basin's active lifetime.

The large-scale depositional patterns are quite clear and will be discussed in the following pages. However, superimposed on this are complex, and sometimes conflicting, data provided by small-scale structures. This complexity is a result of the excellent preservation on minute structures in both outcrop and core. During the field trip we will investigate possible solutions to some of the conflicting data, and hopefully gain further insight into how a Mesoproterozoic, low-relief, interior drainage depositional system operated.

Pass Lake Formation Disagreement exists as to the depositional setting of the Pass Lake Formation. Franklin *et al.* (1980) believed that the sandstones and conglomerates were deposited in a shallow lacustrine environment, whereas Cheadle (1986) put forward an alluvial fan-fluvial model. This controversy may be due to different depositional processes operating in different places. The sequences of powering-down beds, such as at Pass Lake, appear to represent lacustrine sheet sandstones laid down during the waning phases of storm events in a shallow lacustrine environment. The lensoid, large-scale planar and trough cross-stratified sandstones, present on Channel Island and neighbouring islands, probably represent transverse bar sand flats in a braided fluvial system. However, they may be aeolian. The rapid lateral variations in thickness of the unit, not related to position relative to the basin margin, reflect infilling of basement depressions rather than clastic wedges advancing from sediment entry points. A picture arises of the Pass Lake Formation representing first gravels than sands evening the topography by infilling basement valleys through braided stream transport to a shallow lacustrine environment.

In the upper portion of the Pass Lake Formation, the sheet sandstones become interbedded with red, silty shales. This represents a more offshore depositional setting where muds deposited between storm events are not eroded by the next storm, as is the case in the nearshore multi-storey sandstones. The upward decrease in the sand-to-mud ratio indicates deepening water and culminates in the silt- and mud-dominated Rosspport Formation.

Rosspport Formation The contact between the Pass Lake Formation and the Rosspport Formation is gradational, with the name change occurring where the mudstone begins to dominate. Upward through the sequence, indications of dessication begin to appear (e.g. mudcracks and early, authigenic gypsum growth). Commonly at this point, interlayering, with millimetre- to centimetre-scale dolomite laminae, begins to occur. In drill cores from areas closer to the basin center, an interlaminated, decimetres thick, succession of dolomite and gypsum occurs at this stratigraphic position. The dessicated mudstones probably represent periodically exposed mudflats surrounding the lake. This necessitates a regressive phase in the lower Rosspport Formation, for which there is some evidence (e.g. brick-red, friable siltstone at

horizon). The dolomite-mudstone interlayered sequence represents a shallow offshore environment. Muds were washed in during rainy periods and dolomite precipitated during dry periods as the saline lake shrank. The deeper offshore only received clastic input during the latter stages of this depositional episode. It was probably dominated by chemical precipitation of dolomite during the wetter periods when the lake salinity decreased, and by rainout of gypsum crystals on the bottom during dry intervals with increasing salinity. Alternatively, both may have precipitated during dry intervals with hiatuses representing wet periods. The control in this model is the Mg/Ca ratio. The presence of Ca^{2+} inhibits dolomite precipitation and a period of removal of Ca^{2+} via gypsum precipitation is needed before dolomite will begin to precipitate. Thus, first gypsum then dolomite would precipitate during the dry intervals in this scenario. Oxygen and carbon stable isotope analysis is needed to choose between the two alternatives.

Sandstone beds appear interbedded with the upper dolomite-mudstone sequence or overlie it. These are interbedded with red mudstone and the unit is laterally extensive throughout the basin. Closer to the southern edge of the basin, repetitive sandstone assemblages occur (metre- to decametre-scale thicknesses) with intervening decametre-scale, dolomite-mudstone assemblages. Here, trough cross-stratification and channelling are much more evident in the sandstones. Paleocurrents in this unit show higher variability than in the Pass Lake Formation. They appear to have a southern source as opposed to the northern source for the Pass Lake. The sandstones appear to represent renewed tectonism and basin tilting, resulting in sediment influx from the south, with sand dunes migrating through scour channels during high-discharge events. Whether this was subaerial or shallow subaqueous deposition is difficult to determine as strandline position is obscure.

Metres to decametres of mud-chip conglomerates overly the sandstones, either interbedded with the upper sandstones or interbedded with siltstones overlying the upper sandstones. Mudstone fragments are up to metres in diameter, although this is rare, and have variable compositions. This unit appears to thicken towards the basin center and may represent upraising and recycling (cannibalization) of basin margin lithofacies; perhaps a further stage of the north-down basin tilting, represented by the sandstone wedges, upraising the southern basin margin.

The mudblock conglomerate is succeeded in some areas by red mudstones with mudcracks and abundant gypsum veins and nodules representing clastic sabkhas. In other areas, a brecciated and stromatolitic carbonate occurs at this horizon. This represents subaqueous, shallow lacustrine deposition, with teepee structures reflecting periodic drying. These units are overlain by the Kama Hill Formation.

Kama Hill Formation: The Kama Hill Formation consists of red to purple mudstones with interlayered, ripple-laminated, coarse-grained siltstones to fine-grained sandstones. The abundance of rippled units increases upward until they are replaced by sandstones, which continue the coarsening- and thickening-upward trend. The sandstones and all stratigraphically higher units are only present in core. The Kama Hill mudstones represent subaqueous deposition in a non-saline lake. The coarsening- and thickening-upward sequence reflects delta progradation and eventual establishment of a fluvial delta top environment, forming the base of the next formation.

Nipigon Bay Formation: This is a new unit which has not been formally named. It is present only in drill core from under Nipigon Bay. At some localities, the coarsening-upward delta sequence of the Kama Hill Formation is overlain by large-scale, planar cross-stratified, medium-grained sandstones of the Nipigon Bay Formation. These reach approximately 500 m in thickness. Mudstone and siltstone interlayers are almost totally absent with the entire unit consisting of multi-storey sandstones. Colour varies from red to white, with the white sandstones containing gypsum grains and a gypsum cement. Research continues on this formation, with the present data indicating an aeolian origin with possibly some braided fluvial reworking.

At other locations, a different assemblage separates the one described above from the Kama Hill Formation. This assemblage overlies the coarsening-upward sequence of the delta and is composed of fining- and thinning-upward, multi-storey sandstones, metres to decametres thick, separated by friable red mudstones and fissile green mudstones. In some places, minor carbonate layers are associated with the green mudstones. The sandstones are fluvial channels; the intervening mudstones represent a rare example of a preserved Precambrian floodplain succession with sun-baked, poorly developed soils, friable units, and small, open-water ponds. The possibly aeolian sandstones lie sharply on top of this unit.

FIELD TRIP ROAD LOG

<u>Stop Number</u>	<u>Locality</u>	<u>Road Log (km)</u>
<i>DAY ONE STOPS</i>		
1-1	Remote Drill Core Storage Facility (Generalized Stratigraphy)	
1-2	Thunder Bay North (Downtown) (Architectural Stone)	
	Junction Highways 11-17 and 587	0.0 km (Reset)
1-3	Pass Lake (Railway) (Animikie/Sibley disconformity; Pass Lake Fm. Type section)	5.9 to 6.1 km
1-4	Pass Lake (Highway) (Lacustrine Sheet Sandstones, Rossport Fm.)	8.6 km
1-5	Pass Lake (Highway) (Lacustrine Channel Sandstones, Rossport Fm.)	9.9 km
	Pass Lake Crossroad	10.3 km
1-6	Pass Lake (Highway) (Subaerial Siltstone-Caliche, Rossport Fm.)	10.7 km
	Junction Highways 11-17 and 587	0.0 km (Reset)
1-7	Turn-off (Pipeline road) to Enterprise Pb-Zn Mine	14.5 km

DAY TWO STOPS

<u>Stop Number</u>	<u>Locality</u>	<u>Road Log (km)</u>
	Junction of Highways 11 and 17, east of Nipigon	0.0 km (Reset)
2-1	Gurney (Highway) (<i>Archean-Sibley Unconformity, Debris Flows, Pass Lake Fm.</i>)	39.9 km (east of junction)
2-2	Kama Hill (Highway) junction) (<i>Interbedded Dolostone-Siltstone, Rossport Fm.</i>)	21.4 km (east of junction)
2-3	Ruby Lake Marble Quarry (Turn-off from Highway 17; 4.1 km into site)	6.8 km (east of junction)
	Junction of Highways 11 and 17, east of Nipigon	0.0 km (Reset)
	Junction of Highways 11-17 and 585	4.0 km
	Turn-off to Moseau Mountain	5.0 km
2-4	Moseau Mountain (<i>Pass Lake Fm. & Evaporite Breccias, Rossport Fm.</i>)	17.5 - 18.5 km
	Junction of Highways 11-17 and 628	0.0 km (Reset)
	Pull-off parking lot; start of hiking trail	6.3 km
2-5	Lloyd's Lookout (<i>Shales & Rippled Siltstones, Kama Hill Fm.</i>)	(walk trail 0.8 km)
	Junction of Highways 11-17 and 628	0.0 km (Reset)
2-6	Big Squaw Creek Road Cut (<i>Hornfelsed Rossport Fm.</i>)	10.2 km

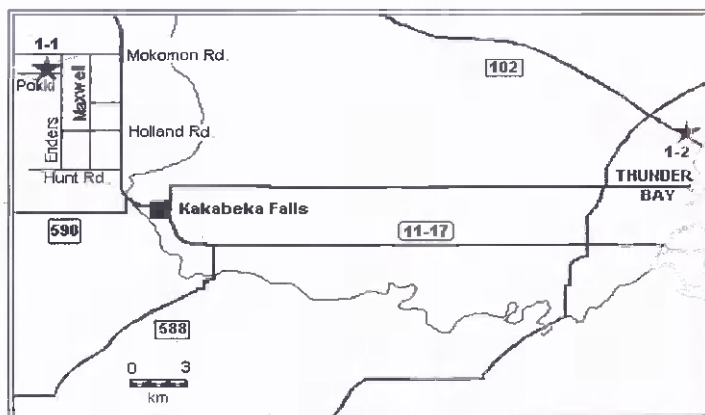
Thunder Bay, Ontario



FIELD TRIP STOP DESCRIPTIONS

STOP 1-1: THUNDER BAY REMOTE DRILL CORE STORAGE SITE, CONMEE TOWNSHIP

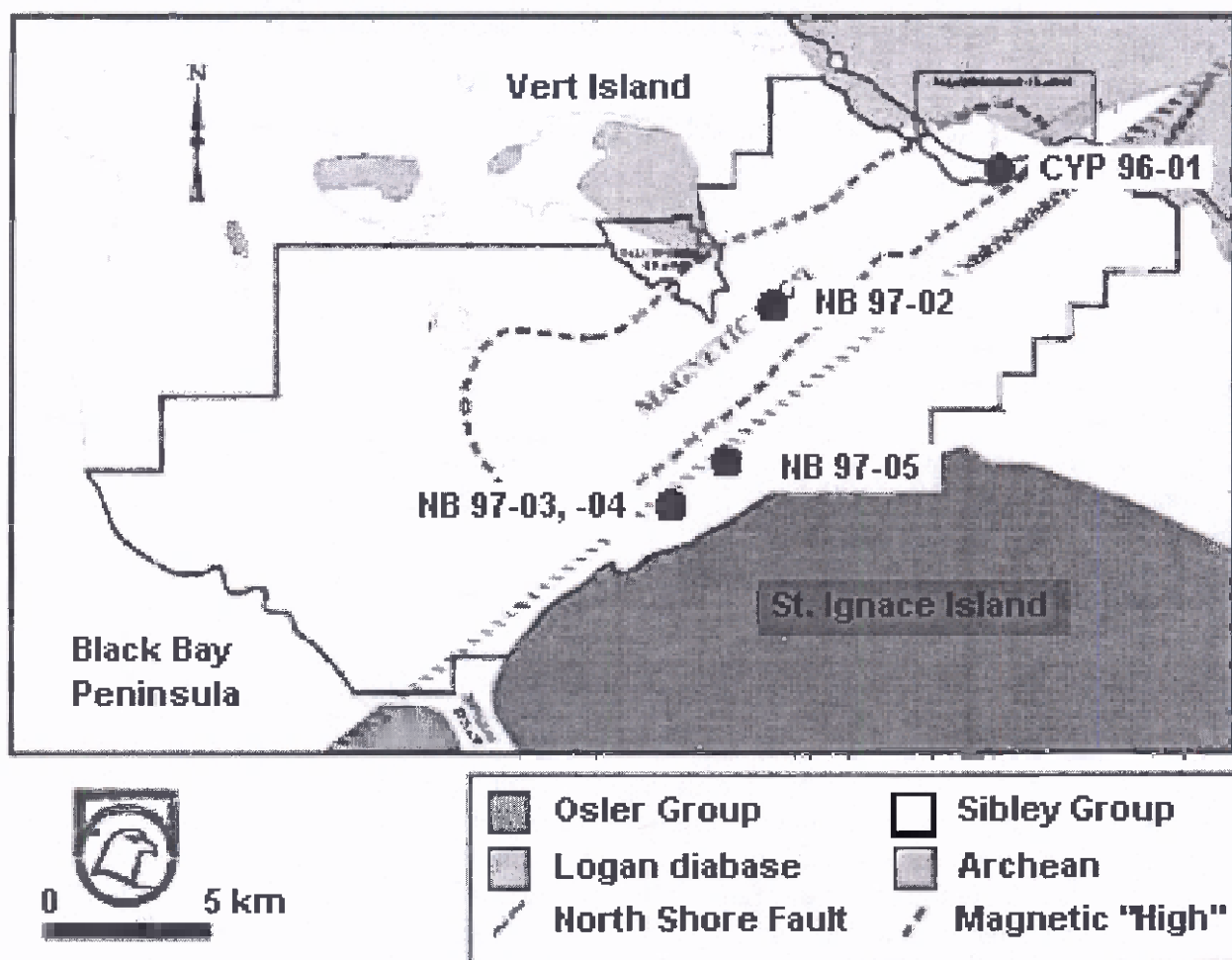
This stop presents an opportunity to view stratigraphic sections of the Sibley Group in diamond drill core donated to, or collected by, the MNM from private sector mineral exploration projects. Core from three drill holes has been selected from different parts of the Sibley depositional basin to provide geographically and stratigraphically distinct sections of the constituent formations and their characteristic lithologies.



The latest drilling was completed by Falconbridge Limited and joint-venture partners, Canmine Resources Corporation and Red Engine Resources, in 1997 from the ice of Nipigon Bay as part of an exploration program designed to evaluate the area's nickel potential (Assessment files, Thunder Bay South District, Thunder Bay). Falconbridge became interested in the area in 1991 and started a compilation and regional reconnaissance program which delineated a 23 km long magnetic anomaly adjacent to the North Shore fault (see map below). Airborne and ground geophysical surveys, geological mapping, soil sampling ensued, culminating in the deep drilling program (ibid). No significant nickel nor copper values were returned from the drilling and the Exploration Licence of Occupation on Lake Superior was allowed to lapse in 1999. Despite the disappointing results, the drilling provided a plethora of new stratigraphic information on the Sibley Group in an area where no drilling had previously taken place. Most remarkable of these results were the unexpectedly thick sections (up to 850 m) that dwarfed previously measured sections and estimated total thickness on the mainland, as well as the presence of upper units of the Sibley Group which had not previously been recognized (i.e. proposed Nipigon Bay Formation).

Drill Hole	Overburden (m)	Sibley Group (m)	Total Length (m)
CYP 96-01	2.0	62.56	464.5
NB 97-02	22.7	452.5	677.8
NB 97-03	30.0	75.0*	105.0
NB 97-04	28.6	858.2	1015.6
NB 97-05	73.2	635.1*	744.2

*Table showing thicknesses of Sibley Group intersected by Falconbridge Limited's drilling.
(*hole did not reach Archean basement)*



Map of Nipigon Bay, showing location of Falconbridge Limited's drilling

The sedimentary rocks unconformably overlie a red granite which has been subjected to Mesoproterozoic, equatorial weathering.

Two drill holes, completed by Noranda Inc., will also be examined at this stop. These holes were drilled north of Dorion on a north-south line, with one lying approximately 15 km closer to the basin center than the other. Thus, the three holes examined will show basin margin (Falconbridge), medial (Noranda) and basin center (Noranda) lithofacies assemblages.

The basin-center drill hole bottoms out in diabase which is overlain by millimetre- to centimetre-scale layers of dolomite and gypsum. This assemblage is approximately 40 m thick and contains pink to red, clay-rich interlayers near its top. This is overlain by a shale-siltstone sequence containing abundant layers of intraformational conglomerate (mud-chip conglomerates).

The medial drill hole overlies a gneissic basement. An upward-fining, conglomerate to sandstone to siltstone, trend is well developed in the Pass Lake Formation present in this hole. Approximately 50 m of inter-laminated dolomite and red mudstone overlie the Pass Lake Formation. This, in turn, is succeeded by a siltstone containing layers of mud-block conglomerate. A brecciated and stromatolitic, white carbonate unit overlies the clastic rocks and this is succeeded upwards by siltstones of the Kama Hill Formation.

The basin-marginal drill hole is similar to the medial hole except: 1) the basal Pass Lake is almost entirely missing; 2) the upper carbonate unit is not present; 3) it contains approximately 550m of overlying Sibley strata not previously discovered elsewhere in the basin. The upper strata consist of a coarsening-upward, deltaic sequence at the top of the Kama Hill Formation. This is overlain by a well-developed fluvial system with multi-storey, powering-down, channel sandstone assemblages imbedded with floodplain fines. Overlying this is a sandstone succession hundreds of metres thick which was probably deposited in a dry aeolian system.

We will spend the morning in the core yard examining the various lithofacies which are present in each of these lithofacies assemblages. For a more detailed description of the sequence see the section on "Depositional Environment".

STOP(S) 1-2 A,B,C: THUNDER BAY NORTH (DOWNTOWN)

This series of stops is designed to show examples of rough and cut dimension stone that was quarried from three different sites around the turn of the century. Sibley Group stone quarries are listed below (summarized from Hinz *et al.* 1994):

Quarry	Rock Type	Years of Operation
Cooke Point	Marble	1931-1940(?)
George Point	Gray sandstone	Late 1800's(?)
La Grange Island	Red sandstone	1882-1883(?)
Nipigon River	Variegated marble	1883-1910(?)
Quarry Island	White sandstone	Late 1800's(?)
Ruby Lake	Variegated marble	1998-present
Simpson Island	Buff sandstone	1904-1912
Vert Island	Red sandstone	1881-1912
Wolf River	Buff sandstone	1913-15; 1921-31



These three architectural stops along one of Thunder Bay's major arteries provide an opportunity to see a variety of Sibley Group stones and note the effects of almost a century of exposure. The architectural descriptions are taken from a walking tour website developed and maintained by the Local Architectural Conservation Advisory Committee (L.A.C.A.C.) (<http://www.city.thunder-bay.on.ca/tbwalkingtour/>).

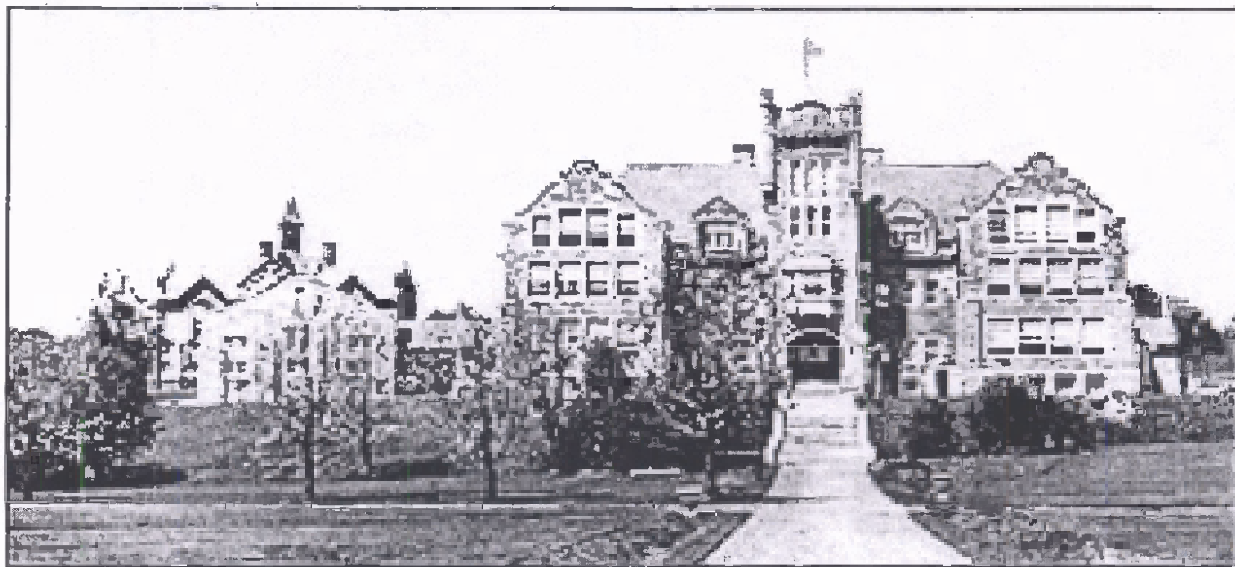
STOP 1-2A: 432 RED RIVER ROAD

This retaining wall is constructed with distinctive, brick-red Vert Island sandstone (Nipigon Bay Formation?). This was a popular stone used in the construction of many grand homes and other buildings in Fort William (now Thunder Bay South).

STOP 1-2B: PORT ARTHUR COLLEGIATE INSTITUTE (401 Red River Road)

This building, constructed in 1909 of Simpson Island buff sandstone (Nipigon Bay Formation?), is an example of the Queen Anne style that was in common use from the 1880s to the 1910s. When the school board made the initial

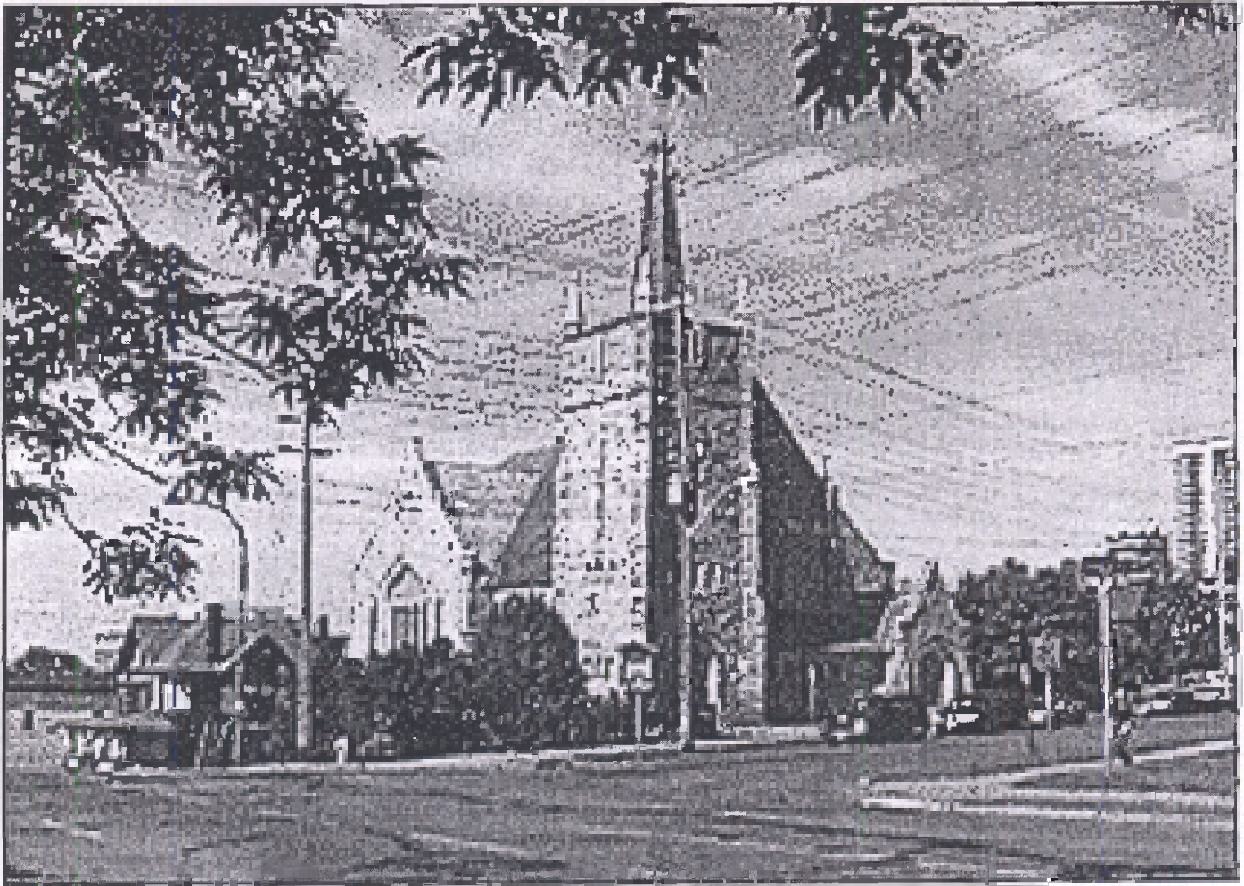
planning for the building, it was decided that it should be "erected for posterity, and not be of the 'shack' order", so they chose the stately Queen Anne style. Alterations in 1925 resulted in the addition of four more classrooms, and more renovations to the north and south in 1953 and 1962 created other rooms. These alterations used stone in an attempt to blend into the original building, but a gymnasium planned in 1964 and completed in 1974, provoked controversy as its design was incompatible with the rest of the school.



Port Arthur Collegiate Institute, shortly after its construction

STOP 1-2C: TRINITY UNITED CHURCH (30 Algoma Street South)

This building, constructed in 1906, was formerly known as the Trinity Methodist Church, and became the Trinity United Church after the United Church of Canada was formed in 1925. Constructed of rough cut, Simpson Island buff sandstone (Nipigon Bay Formation?), this structure is an example of the Late Gothic Revival style that was popular from the 1890s to the 1940s. The unusual tower features very narrow windows, four buttresses, each capped with a pyramid shaped finial, and an extremely sharp hexagonal spire. The rest of the building also features very steeply pitched roofs, and arched windows in the Gothic style.



Trinity United Church

STOP 1-2D: MASONIC HALL (262-270 Red River Road)

Built in 1910 and also known as Shuniah Lodge, this stone, brick and concrete building replaced the old Masonic temple that was destroyed by fire in 1909. The first floor is made of cut Nipigon River marble (Rosspoint Formation) and the entrance features carved marble pilasters and decorative panels. Originally there was a dome on the roof over the entrance, which has since been removed. The central portion of the building has a Mansard roof of French design. The building's windows are decorated with alternating round and triangular pediments above them. Commercial space occupies the ground floor, while the lodge is located above.

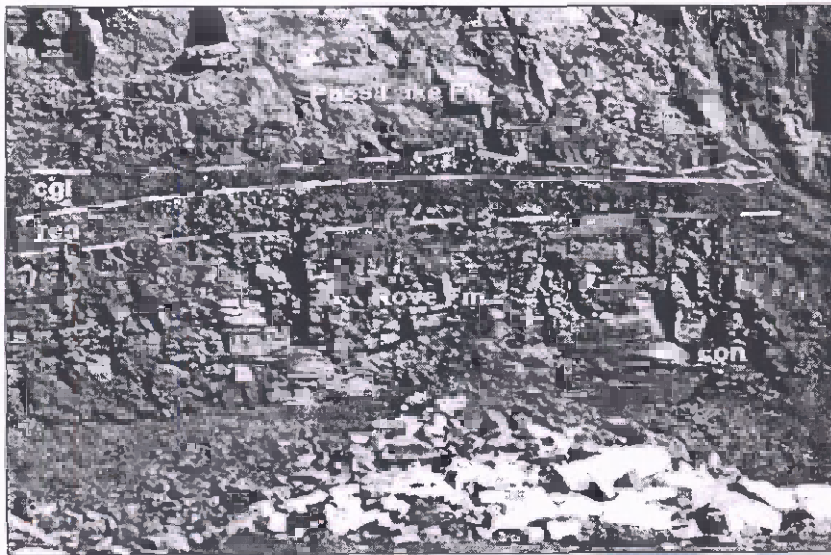
STOP 1-3: PASS LAKE (RAILROAD CUT)



A cliff on the far side of the railroad tracks contains the type section of the Pass Lake Formation. A low-angle disconformity separates the basal conglomerates of the Pass Lake Formation from the underlying, ~1.9 Ga Rove Formation shales (*lower photo*). The maroon and green shales exhibit extensive evidence of Mesoproterozoic weathering. The basal conglomerate thins and thickens laterally, pinching down to pebbly sandstone in places. Clasts are generally surrounded and dominated by local Gunflint formation lithologies. The matrix is poorly sorted. The conglomerates are overlain by a thinning-upward sequence of sandstone beds (*photo at left*). Individual beds are reasonably laterally continuous though sometimes lens out. They are dominated by upper flow regime parallel lamination with occasional ripples and small-scale dunes on their tops.

depositional environments have been proposed. While examining the lithofacies in the field we will discuss the merits of each interpretation.

The depositional environment of this sequence is continuous. Both alluvial fan-braided fluvial and shallow lacustrine (Cheadle 1986; Franklin 1980; respectively)



Disconformity between the basal conglomerate of the Pass Lake Formation (cgl) and weathered Rove Formation shales (reg) at Pass Lake. A concretion (con) is shown at lower right.

STOP 1-4: PASS LAKE (HIGHWAY)

This is a roadside exposure of the transition zone between the Pass Lake and Rosspoint Formations. Laterally continuous sandstone beds interlayer in a red silty mudstone assemblage. The sandstones are often massive, though one bed at this location contains low angle, wispy, trough cross-stratification.



Here again, two possibilities exist for the environment of deposition. The sandstones could represent crevasse splay sand sheets spread out on the floodplain fines during flood-stage failure of fluvial levee systems. Alternatively, the sand sheets may represent shallow lacustrine storm deposits caused by rainfall event-driven fluvial sand delivery to a lacustrine system and storm surge ebb plus river mouth density current transport of sand offshore into a fair weather silt-dominated area. The merits of both interpretations will be discussed in the field.

STOP 1-5: PASS LAKE (HIGHWAY)

This stop contains a variety of lithofacies. Large sandstone lenses cap the outcrop, overlying siltstones, a fairly continuous, non-silty, purple shale and a massive, lensoid dolomite.



Layer geometry and contact relations in this exposure are perplexing at first glance. Take some time to trace out the geometries of the various lithologies. Hopefully we will be able to answer the following questions while at this stop:

- Is the dolomite really lensoid?
- Is this caused by structural disturbance?
- How does the dolomite correlate across the road?
- Are any intraformational conglomerates present?
- What is their geometry?
- How could have this small sequence of various lithologies have formed?

STOP 1-6: PASS LAKE (HIGHWAY)

This is a small, crumbly outcrop of red, friable silty-mudstone. Although it is not impressive, its brick-red colour and friable nature are important. Commonly, outcrops such as this one are interpreted as highly sunbaked soils / caliche, and that is probably what this outcrop represents.

STOP 1-7: ENTERPRISE MINE

Peter McKellar discovered a lead-copper-mineralized vein at this location in 1865 that was subsequently developed as the Enterprise Mine (Tanton 1931). Between 1870 and 1876, two shafts were sunk. No. 1 shaft reached a depth of 180 feet (55m); levels were run easterly 76 feet and westerly 66 feet at a depth of 60 feet. A stope was made in the west drift and a winze was sunk in the east drift about 50 feet from the shaft. A crosscut was driven from the bottom of the shaft 15 feet south. Shaft No. 2 was sunk 60 feet at a point approximately 300 feet west of No. 1 shaft (ibid).

The total recorded production from the mine was a single shipment of 167 tons made to Swansea, Wales, in 1875 (ibid) that averaged 37.4% Pb. Although no other metal credits are listed, appreciable copper, silver and gold were reported in some samples. Analyses conducted by Dr. E.J. Chapman in 1874 include one of a large, average sample of the vein near surface that returned 41.84% Pb, 5.40% Cu, 3.2 ounce Ag per ton and 0.33 ounce Au per ton (Tanton 1931). Hawley (1930) reported Chapman's estimate of part of the ore body near surface as 47.5% Pb, 10.0% Cu, 17 dwt. 12 grammes (sic) Au and 2 ounces 2 dwt. Ag per ton.

Mining operations ceased in 1876; fire destroyed the mine buildings a few years later (Tanton 1931). New owners dewatered and sampled the workings in 1884 but no ore was produced. Dewatering, examination and subsequent suspension took place between 1926 and 1927 (ibid). The mine site was initially rehabilitated in 1926 and 1927 and is now monitored and maintained by the Canadian National Railway.

Hawley (1930) and Tanton (1931) have described the local geology. Ore-bearing veins occupy a breccia zone in dolomitic siltstones of the Rossport Formation that extends down into the Archean granitic basement. This breccia zone strikes 065° and dips 70° SE to vertically. It has been traced over 60 m and has an average width of 1.2 m. Fracture-controlled offshoots and branch veinlets may occur in a metre-wide envelope or "shatter zone" around the main vein. The veins tend to pinch-and-swell; mineralized zones over 4 m wide have been noted (ibid).

The veins consist of the ore minerals galena, low-Fe sphalerite pyrite and chalcopyrite in a gangue assemblage of quartz / amethystine quartz, calcite and pink barite. Malachite has precipitated on chalcopyrite-rich, weathered dump material.

First synopsized by Hawley (1930) and further described by Tanton (1931), "lead-zinc-barite veins" (LZBV), as they are locally termed, were reported as early as 1866. Of the approximately 30 deposits, only the Dorion, Ogema and Enterprise deposits were mined to any degree. As noted by Franklin and Mitchell (1977), these small deposits comprise a distinct metallogenetic group that has a consistent, simple vein mineralogy and structure, and is spatially confined to the Sibley Group depositional basin. Although temporally similar, they are clearly distinguishable from silver-bearing veins that occur near Thunder Bay in Paleoproterozoic Animikie Group sedimentary rocks and Logan diabase.

Franklin and Mitchell (1977) subdivided these LZBV deposits into three classes:

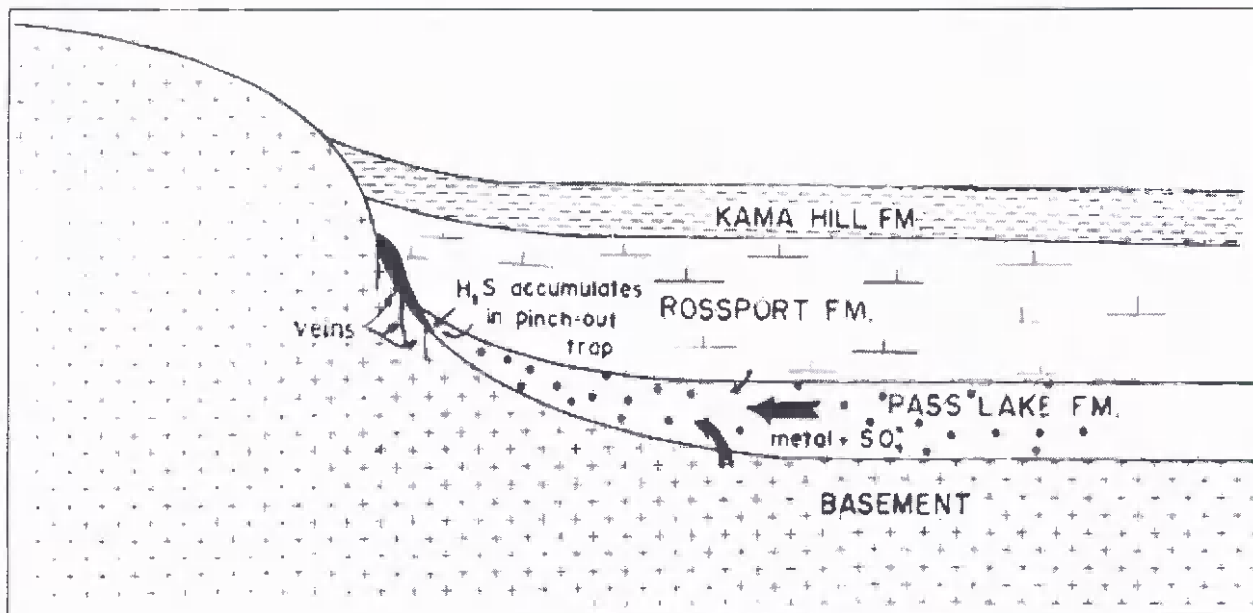
- (1) those that have the Sibley Group as at least one wall rock (e.g. Dorion Mine);
- (2) those that are entirely enclosed within Sibley Group rocks (e.g. Enterprise Mine); and
- (3) those lie within the Archean basement but near the unconformity with basal sandstones of the Sibley Group (e.g. Ogema Mine).

Wall rock alteration, typically best manifested in dolomitic rocks, is tan to grey in colour and is characterized by enrichment in SiO_2 and CaO , slight enrichment in FeO and MnO and depletion of MgO , CO_2 and Al_2O_3 . (ibid). This indicates silicification, the conversion / replacement of dolomite by calcite, removal of feldspar and mica and pyritization. Metal enrichment is noted in altered wall rock.

Archean wall rocks have returned Archean K-Ar ages, indicating that vein-forming fluids were not hot enough to re-equilibrate the argon in the analysed mica. Sulphur-isotopic data indicate equilibrium between galena and sphalerite yielding a depositional range of $35^\circ\text{--}135^\circ\text{C}$, and disequilibrium between sulphide-sulphate pairs (ibid). Haynes (1988) noted that fluid inclusions in quartz and sphalerite indicate deposition from Na-Ca chloride brines (26-33 wt.% $\text{NaCl} + \text{CaCl}_2$ equiv. salinity) at temperatures of $90^\circ\text{--}200^\circ\text{C}$. He also suggested that two isotopically distinct brines were involved in vein formation. The inclusion fluids contain $\text{NaCl} > \text{CaCl}_2 \gg \text{KCl}$, compositionally similar to present-day basinal brines and Mississippi Valley-type mineralizing solutions.

High $^{206}\text{Pb}/^{204}\text{Pb}$ ratios ($\cong 30.7$) from vein galena (Franklin and Mitchell 1977) likely reflect radiogenic lead associated with the radioactive, Archean, Bowker granite. As noted by Franklin (1978b), the Enterprise Mine has the distinction of being the only LZBV deposit to contain appreciable amounts of uranium. U-enriched (up to 540 ppm U) portions of the Enterprise veins (Ruzicka 1976) are apparently spatially associated with galena and likely reflect proximity to a uraniferous, basement source rock. Lead isotopes are highly anomalous, yielding a secondary isochron that likely indicates a mixed Archean-Paleoproterozoic lead source (Franklin and Mitchell 1977). These anomalous and very radiogenic leads are more uranogenic and less thorogenic than leads found in silver-bearing veins near Thunder Bay (Franklin *et al.* 1986) to which they are temporally and tectonically similar. The LZBV mineralization is most likely related to dewatering of the Sibley basin during Keweenawan rifting (ca. 1090-1110 Ma).

Franklin and Mitchell (1977) suggested that the vein metals were either leached from basement rocks and/or the breakdown of Sibley sandstone matrix. Metals and sulphate moved through permeable Pass Lake Formation sandstone as chloride complexes (Haynes 1988) in a very saline brine, and precipitated in veins at the sandstone pinch-out, at or near the Archean unconformity. Isotopically light sulphur, perhaps of organic derivation, may have been locked in a H_2S -bearing gas trap at the pinch-out, and caused the metals to precipitate.



Schematic model of the formation of LZBV deposits from Franklin and Mitchell (1977); scale is variable.

Amethyst

The largest deposits of amethyst, Ontario's provincial mineral emblem, occur in the Thunder Bay area. Nearly all of the most extensive and richest deposits occur in a 10 km-long, northeast-trending belt near the margin of the Sibley Group outcrop where it is underlain by Hilma Lake granite (Kissin 1990; Garland 1994; Vos 1976). Other occurrences, mainly hosted in granite, are also spatially related to Sibley outcrop.

These deposits consist of quartz + amethystine quartz, with minor calcite, barite and sulphide minerals (e.g. pyrite, chalcopyrite, bornite). They typically occupy faults and related fracture zones where brecciation and open-space-filling are common (*ibid*). The similarities in geologic setting between amethyst and LZBV led Kissin (1990) to suggest that these two types of vein deposits are cogenetic. Fluid inclusion studies by McArthur (1988), reported in McArthur and Kissin (1988), indicate that most amethyst formed near the surface, at temperatures ranging from 60° to 90° C. Kissin (1990) suggested that they may be low-temperature derivatives of LZBV, poorer in sulphides and richer in quartz. The presence of a uraniferous basement granite (Franklin 1978) was probably crucial in the formation of local amethyst deposits (Kissin 1990), because radioactivity is essential in causing elemental substitution and defects in the quartz crystal that impart the purple colour (Garland 1994 and references therein).

In the vicinity of the Enterprise Mine, amethyst veins and breccia zones occur in basement granitoids and in overlying clastic and stromatolitic Sibley Group rocks at or near the unconformity. Amethyst ± base metal-mineralized stromatolites (*see frontispiece, this paper*) occur at the Welch's Farm occurrence (C. Anderson, pers. comm., 1999), at the Ontario Gem Company Amethyst Mine (Garland 1994) and in other places east of Ancilff station (Tanton

1931; J. Scott, pers. comm., 1999; McIlwaine 1971a,b). They typically occur on top of prominent granite knobs which may reflect paleotopographic "highs".

The nearby rockcut produced by the railway exposes the basal contact of the Sibley Group with underlying Archean granitoids. A regolith makes identification of the contact difficult in places.



Unconformity between Archean granite (Gr) and trough crossbedded, Pass Lake Formation quartz arenite (s.s.), Enterprise Mine

Medium-grained, trough cross-stratified sandstones lie directly on the regolith. The lensoid, cross-cutting bed geometry of the sandstone units is easily discernable. Flow was to the southeast (Cheadle 1986). Again the exact depositional environment these units were laid down in is difficult to ascertain. Without fossil evidence to rely on near-shore, trough cross-stratified, sandstone sequences are difficult to differentiate from South Saskatchewan-type, braided fluvial, channel sandstones. Overall unit geometry can give some clues, as can variability of paleocurrent directions and associated minor lithofacies, but, thick, laterally persistent, monolithic successions of trough cross-stratified sandstone pose serious problems in assigning a depositional environment. Are there any clues in this assemblage which would help in deciding between the two aforementioned environments?

STOP 2-1: GURNEY (HIGHWAY)

The basal unconformity between the Sibley Group and underlying granitoids is exposed at this location. A channel is visible, eroded into basement, and containing matrix-supported conglomerates overlain by cross-stratified sandstones. This sequence differs from the other coarse-grained, basal sediments we have examined, as it represents episodes of subaerial debris flow activity interspersed with normal flash flood runoff.



Unconformity between weathered Archean granite (Gr) and Pass Lake Formation debris flows and sandstones, Highway 17 at Gurney

Archean granitic rocks are altered at the unconformity and likely represent a pre-Sibley weathered regolith. This weathered paleosol, noted by Gill (1926) and Moorhouse (1960), was locally described by Scott (1987). Friable, blotchy, red and green granite hosts quartz-carbonate veins between exfoliation blocks. Feldspars have been hematitized and/or destroyed; ferromagnesian minerals have been chloritized (ibid). Limited sampling of drill core from the Black Sturgeon Lake area cited by Scott (1987) suggests that this alteration may typically involve marked increases in Fe_2O_3 , MgO , H_2O and decreases in Na_2O , CaO and perhaps K_2O . Paleomagnetic data suggest that this weathering was equatorial (G. Borradaille, unpublished data, 1999).

As noted by Franklin (1978b), Scott (1987) and Tanton (1948), a number of uranium occurrences are associated with altered Archean granitoids and overlying sedimentary rocks within the Sibley basin, prompting comparisons with the Athabasca basin in Saskatchewan. Favourable local parameters for supergene uranium deposits include: (i) uranium-enriched basement rocks (quartz monzonites, pegmatites); (ii) onlaps of basal Sibley sandstones on Archean paleotopographic "highs"; and (iii) Keweenawan(?) faults that extend to the basement (ibid).

STOP 2-2: KAMA HILL

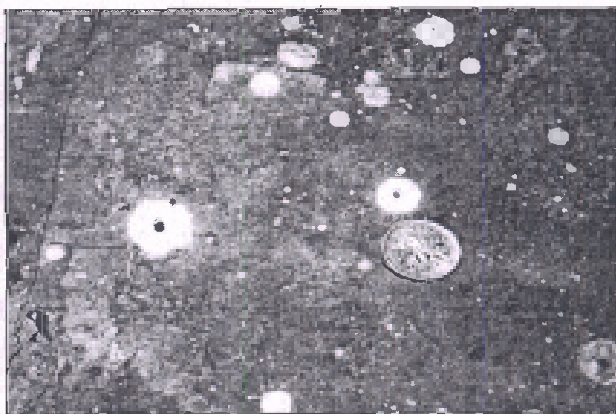
Diabase-capped Kama Hill provides an excellent roadside exposure of the Rossport Formation. The sequence is dominated by interlaminated dolostone and red shale with layer thicknesses ranging from millimeters to ten centimeters. There is some cyclicity in layer thickness variation up through the sequence, but it is not a strong trend. Some carbonate layers contain coarse- to medium-grained sand grains. Dessication cracks are present, but not common. Syndimentary deformation manifests itself as small to large slump folds and brecciation of units in places. The sequence is intruded by diabase dykes which bake immediately adjacent sediments. Some evidence has been put forward that the diabase intruded watery sediment, but we have not seen clear indications of this. Structural controls on the emplacement of local sills were suggested by Antonellini and Cambray (1992). A chert-carbonate unit (Middlebrun Bay Member; Cheadle 1986) locally hosts stromatolites and hydrocarbons.

The Kama Hill sequence, exposed below the upper diabase sill, was laid down in the shore-proximal region of a playa lake. Pieces accessible in a pile of excavated talus display ripple marks, mudcracks and rare raindrop impressions. During rainy periods, water influx into the lake brought and deposited hematite-rich clays. In dry periods, evaporation from the internal drainage system resulted in lake contraction, hypersalinity and the precipitation of dolomite. This requires a high Mg/Ca ratio in the lake water and may have been the result of the precipitation of gypsum in the central lake. However, this is unlikely as higher salinities are expected in the shallower, marginal areas than the lake center, and thus, gypsum precipitation should be initiated in the shallows, producing shale-gypsum-dolomite triplets. The lack of these means that gypsum precipitation was not necessary to increase the Mg/Ca ratio. The ambient ratio in the lake water itself must have been high enough for the precipitation of dolomite.



Interlaminated red shales and dolostone, formed by climatic cycling in a semi-arid, playa lake environment, Rossport Formation, Kama Hill.

Reduction spheres in hematitic siltstone, Kama Hill, with black, carbonaceous, locally radioactive centers.



STOP 2-3: RUBY LAKE MARBLE LTD. QUARRY

The Ruby Lake Marble Ltd. quarry is owned and operated by Don MacAlpine and Gerald Landry. Variegated, multi-coloured, banded marble is quarried for landscaping stone. Approximately 175 tonnes of marble were quarried and shipped in 1998 (D. MacAlpine, personal communication, 1999). The dimensions of the largest, transported block was 1.8 by 0.75 by 0.50 m ($\cong$ 1.8 tonnes) (ibid). Approximately 398 tonnes of marble were quarried and shipped in 1999 (D. MacAlpine, pers. comm., 2000). Additional information is provided on their World Wide Website (<http://www.rubylakemarble.com>).



This marble consists of contact metamorphosed, Mesoproterozoic, Rossport Formation (Sibley Group) dolostone and other, calcareous sedimentary rocks in the contact metamorphic aureole of Keweenawan diabase sills. It has previously been termed Nipigon River marble and was quarried from 1883 to ca. 1910 at a site on the eastern side of the Nipigon River, approximately 6 km west of the Ruby Lake quarry (Hinz *et al.* 1994). A similar, hornfelsed unit is described in more detail at Stop 2-6.

Calcite, dolomite, epidote and opaque minerals were noted in thin section by Hinz *et al.* (1994) from the Nipigon River quarry. The following chemical analyses for Nipigon River marble (wt.%) were also provided (ibid):

Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅
89MCK-09	35.21	0.20	8.20	2.04	1.53	0.05	23.56	26.74	0.00	2.38	0.09
89MCK-10	29.19	0.10	5.20	2.04	0.00	0.03	27.63	35.32	0.00	0.41	0.07

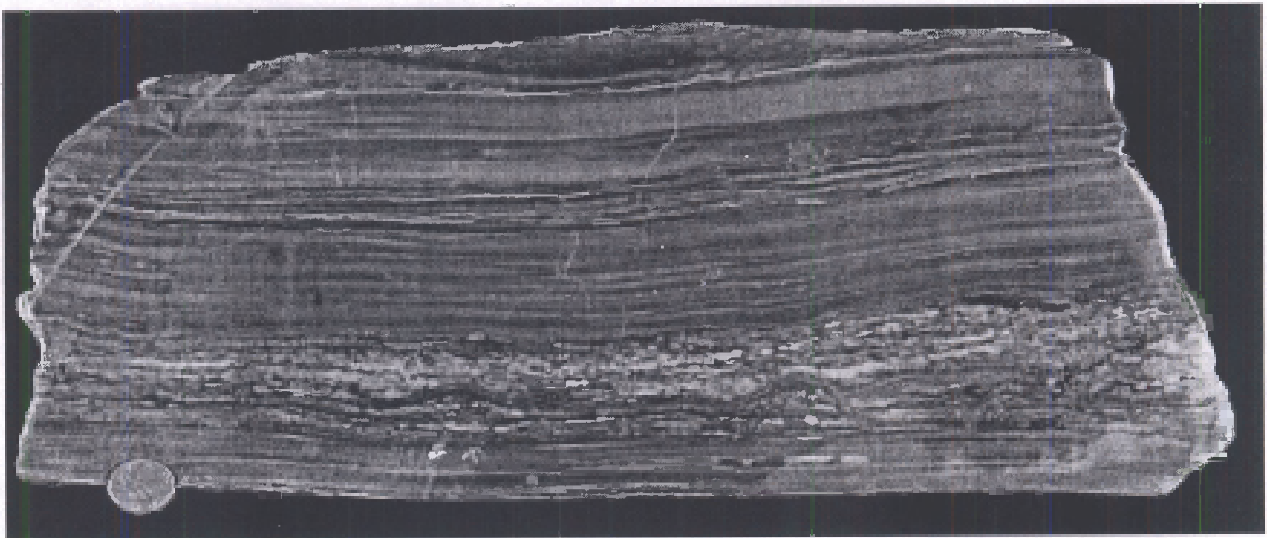
Shallow exploration trenches on the side of the road leading to the top of Ruby Mountain (i.e. top of the upper sill) have exposed copper-mineralized marble. Fine-grained, disseminated blebs of native copper (0.1 to 1.0 mm) occur along calcite-coated, hairline fractures parallel to bedding planes. Mineralized fractures are most easily recognized where secondary (supergene) malachite has formed. The top of the adjacent (middle?) sill is exposed just farther along the road. Polygonal cooling joints (seen in plan view) and the chilled, upper margin are evident.

Similar, copper-mineralized, calcareous units have been noted near a diabase sill contact at Hughes Point, 2.5 km to the south-southwest by Schnieders *et al.* (1996). At this location, 2 to 5 cm wide calcite veins contain disseminated covellite (after chalcocite?) and malachite.

The orientation of the veins are roughly parallel to joints developed in the adjacent sill. Grab samples have returned up to 3.065% Cu, nil Au and nil Ag (ibid).

Franklin (1970) described copper-mineralized stromatolitic units near Disraeli Lake. A variety of copper minerals, including digenite, cuprite, covellite, chalcopyrite, native copper and malachite, were identified as open-space fillings in the vuggy host rock. Franklin (1970) suggested that the copper was introduced epigenetically from a gabbro to peridotite plug that intruded these Sibley Group rocks. The close spatial association of copper-mineralized rocks with diabase sill contacts in the vicinity of Ruby Lake supports this theory.

The marble is stromatolitic, though this is difficult to ascertain in most places due to the alteration. The stromatolites are best seen on the tops of beds where they protrude, causing the contact to become wavy. Hints of the presence of stromatolites are visible, giving the impression that a large amount of the horizontal layering is stromatolitic S-mat. This sequence was deposited in a strandline proximal position in the playa, as denoted by the presence of teepee structures in this horizon at other locations. This infers that lake size had stabilized during this interval, eliminating the seasonal, large-scale fluctuations in shoreline positioning.



Parallel-laminated carbonate with variable amounts of hematitic clays in possible S-mat stromatolite, Ruby Lake Quarry

STOP 2-4: MOSEAU MOUNTAIN

The Moseau Mountain stop consists of a number of small outcrops extending from the granitic basement to the middle portion of the Rossport Formation. Cross-stratified sandstones of the Pass Lake Formation overlie basement. The lower sandstones are multi-storey, but as we proceed upwards mudstone interbeds begin to appear. Above this zone the sandstones are replaced by siltstones, continuing the general upward-fining trend. Higher in the siltstone succession mudblock conglomerates begin to appear. These, in turn, are upwardly transitional into evaporite block conglomerates, with gypsum and dolomite clasts. Diabase intrusions occur above this zone.

The lower portion of the sequence is similar to the one observed at Pass Lake, where a thinning and fining trend exists from the basal sandstones and conglomerates to the siltstones of the basal Rossport Formation. This again represents expansion of a playa lake system. The intraformational conglomerates which overly this sequence may have been produced by two mechanisms. Dissolution of an evaporitic layer causing collapse of the overlying units, or tectonic raising of the basin margin resulting in the erosion and resedimentation in a more basin-center environment of the upper marginal sediments. The latter explanation is preferred as this unit is regional in extent, occurring where it is unlikely that underlying evaporites ever existed.

STOP 2-5: LLOYD'S LOOKOUT

This stop, located on the Nipigon-Red Rock hiking trail, features exposures of Kama Hill Formation shales and siltstones underneath a Logan diabase sill which caps the prominent mesa and provides the vantage point over Nipigon Bay known as Lloyd's Lookout. From this vantage point, one can view the mouth of the Nipigon River and Nipigon Bay. The locations of the quarries mentioned in *STOP 1-2* are shown in the photograph below.



View from Lloyd's Lookout, looking east over Nipigon Bay

This is one of the few accessible exposures of the Kama Hill Formation. This Formation is homogeneous, consisting of brick-red mudstones with interlayered, current-rippled, coarse-grained siltstones and very fine-grained sandstones. The rippled layers are usually centimeters to a few decimeters thick. The mudstones are commonly mudcracked.

Cheadle (1986) believed that this unit was deposited as subaerial mudflats, representing the final stage of infilling of the lake basin. However, new drill core data indicate that the Kama Hill Formation is overlain by a prograding, deltaic sequence. This infers a subaqueous position for the Kama Hill Formation with the sediments representing prodelta to distal bar sequences. In core, these are overlain by distributary mouth bar and fluvial channel sandstones. The mudcracks are problematic. They may be syneresis cracks, formed subaqueously through differential swelling of clays, or they may represent large-scale, water level fluctuations. Neither of these explanations appears adequate.

The view over Nipigon Bay is towards where the drillcore examined the previous day was drilled. During Kama Hill time we would be standing in approximately 20 m of water, looking towards a horizon where a delta system would be located, building towards us.

STOP 2-6: BIG SQUAW CREEK ROAD CUT

This stop presents the opportunity to view stoped blocks of Rosspoint Formation dolostone in a diabase sill, as well as the attendant metamorphic and metasomatic effects. Pale green to buff, thinly bedded dolostones are preserved as rafted xenoliths up to several metres thick.



Rafted xenoliths of hornfelsed Rosspoint Formation in Logan diabase sill, Highway 11-17 at Big Squaw Creek

Thin section petrography reveals a recrystallized, granoblastic texture predominated by epidote. Subhedral to euhedral, epidote porphyroblasts are typically 0.05 to 0.10 mm in size, but they may reach 0.25 mm, in a fine-grained matrix of vermicularly intergrown quartz and feldspar. Large (<0.75 cm), optically continuous, decussate, acicular biotite porphyroblasts are also overgrown by epidote. Radiating, sheaf-like aggregates have been tentatively identified as chlorite. Minor sericite and calcite were also noted. Epidote persists into the fine-grained, chilled margin of the diabase. Joints and fractures surfaces may be coated with muscovite. Large, conspicuous, lime-green clots of prehnite occur in epidote-calcite veins within the diabase.

Little work has been done on contact metamorphic effects in vicinity of the diabase sills. Sutcliffe (1986) noted that such metamorphism occurs in calcareous rocks over a zone several metres thick, where bleaching is its most obvious manifestation. Fine-grained, calc-silicate mineral-bearing assemblages identified by Sutcliffe (1986) include calcite, plus (in order of increasing metamorphic grade): talc, tremolite, serpentine (after forsterite) and diopside. Calculations by Sutcliffe (1986) suggest that the diabase sills were emplaced at shallow crustal levels, near the Archean-Proterozoic unconformity, at a lithostatic load of <0.4 kbars. The sills crystallized over a temperature range of 1100° to 880°C, while adjacent wall rocks may have experienced temperatures ranging from 540° to 665°C (ibid).

The mineral assemblage in the calc-silicate hornfels at Big Squaw Creek may reflect simple recrystallization of the impure, calcareous sedimentary units without significant compositional or volumetric changes (cf. Hall and MacKevitt 1962).

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GEOLOGY OF THE LAC DES ILES PGE-Cu-Ni MINE

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North American Palladium Limited's open pit Lac des Iles Mine is hosted by the Archean Lac des Iles mafic-ultramafic intrusive complex. This trip will focus on PGE mineralization as it relates to intrusive lithology, structure and alteration.

Geology of the Lac des Iles Palladium Deposit

Regional Geology

The Lac des Iles area is underlain by the Archean age rocks (2.6 to 3.0 billion years) of the Canadian Shield. These rocks are overlain and intruded by Proterozoic age rocks (0.9 to 2.2 billion years) that formed during rifting of the continent, centered on Lake Superior. Archean age, mafic to ultramafic intrusions that host, or have the potential to host PGE mineralization, are dispersed throughout Northwestern Ontario. These intrusions are concentrated within granite-greenstone terrain, and along major faults. The Quetico Fault, which passes 20 km south of Lac des Iles, is coincident with a 180km southwesterly trend of PGE bearing intrusions. A particular high concentration of intrusions occur in the Lac des Iles area, a number of intrusions define a circular feature with a 30 km diameter. These major crustal sutures are important as they allow PGE rich magma that can only originate from the earth's mantle, to move upwards into the crust.

Proterozoic age PGE mineralization is not as common in northwestern Ontario. However, potential PGE hosting Proterozoic age mafic to ultramafic intrusions are much more voluminous. Gabbro covers or intrudes most of the Lake Superior area, no less than 20,000 km². A recent 8 g/t Pd+Pt discovery made east of Lac des Iles at Wolf Mountain in a Proterozoic age ultramafic intrusion is testament to its potential.

The known mafic to ultramafic intrusions in the Lac des Iles area consists of the following:

- Lac des Iles Complex (~30km²) (Company ownership; 100%)
- Tib Lake Intrusion (~25km²) (Company ownership; 65%)
- Legris Lake Intrusion (~12km²) (Company ownership; 0%)
- Demars Lake Intrusion (~1km²) (Company ownership; 100%)
- Wakinoo Lake Intrusion (~11km²) (Company ownership; 100%)
- Taman Lake Intrusion (<1km²) (Company ownership; 0%)
- Dog River Intrusions (<1km²) (Company ownership; 0%)
- Buck Lake Intrusion (~5km²) (Company ownership; 90%)

These intrusions are similar in that they are late tectonic and intrude tonalite gneiss, have tholeiitic affinity and contain phases ranging from ultramafic peridotitic and pyroxenitic cumulates to magnesian gabbro-norite and iron-rich gabbro. PGE mineralization is found in all except the Taman and Dog River intrusions.

The Lac des Iles Complex consists of three distinct lithological packages; to the north, ultramafic intrusions centered on Lac des Iles, complex gabbroic intrusions immediately south of the lake (Mine Complex), and a hornblende gabbro southwest of Camp Lake (Camp Lake Intrusion). These three segments are partially separated by tonalite septums. The ultramafic intrusions consist of two coalescing centres, each of which consists of several intrusive phases. The northern ultramafic centre is nearly circular in plan with a diameter of nearly 4 km. It consists of interlayered dunite and wehrlite, olivine-clinopyroxenite and clinopyroxenite, and websterite which define a

concentric inward dipping pattern. These lithologies occur in several cycles which probably represents fractional crystallization of separate magma pulses. It also represents repeated opportunities to concentrate PGE's. This part of the complex does have several PGE's bearing sulphide rich horizons, and 5 PGE occurrences (>1 g/t Pd+Pt) were discovered in 1999. The elliptical southern ultramafic centre consist of a massive wehrlite core surrounded by websterite. These two centres are partially separated by a gabbro-norite wedge on the east side of the complex. This area contains 7 PGE occurrences, 5 discovered in 1999.

The Camp Lake Intrusion, south of Camp Lake, had been considered to be a relatively homogenous hornblende gabbro and barren of PGE mineralization. Historically, it had been subjected to minimal exploration. Stripping along its northern margin in 1999 revealed unexpected complexity and sampling generated low grade PGE assays.

The Mine Complex, which is compositionally, texturally and structurally complex, contains abundant PGE mineralization. Its composition ranges from anorthosite to clinopyroxenite, gabbro-norite to melanonorite and includes magnetite rich gabbro. Textures include equigranular fine to coarse grain, porphyritic, pegmatitic, varitextured units and heterolithic-gabbro-breccia. These last three textural types coincide with all mineralized zones. Fragments in these breccias, which vary from a few centimeters to tens of meters in size, are representative of most of the lithologies in the area. The breccias were created as the result of an energetic influx of a volatile charged, PGE bearing magma, which assimilated a large volume of rock and disrupted layering. The PGE bearing magma, typically a melanogabbro, is the matrix to the breccia and contains high grade PGE values. The accompanying volatiles, migrating through the chamber, produced varitextured and pegmatitic gabbro, altered the silicates, mobilized metals and produced sulphide poor mineralization. The varitextured gabbro that defines the oval shape of the Mine Complex, is host to the Roby Zone. The occurrence of these key rock types and mineralization throughout the Mine Complex is evidence that the mineralizing process affected it in its entirety.

PGE Mineralization of the Mine Complex

Zones of PGE mineralization with delineated resources in the Mine Complex consists of the Roby Zone, Twilight Zone, Baker Zone and Moore Zone. Other mineralized areas, with values from 0.7 g/t to 23 g/t Pd have had limited amounts of surface sampling and diamond drilling and require follow-up.

Roby Zone

The Roby Zone is a bulk mineable deposit with a minimum strike length of 850 meters, width achieving 365 meters and a depth of 500 meters. The zone remains open at depth, and to the southeast and potential exist for new mineralization to the west.

The Roby Zone contains three distinct ore types; Shear Ore (7.6% of volume), North Roby Ore (5.3% of volume) and Breccia Ore (87.1% of volume).

Shear Ore

Shear Ore is confined to a portion of a 15 to 25 meter thick layer of pyroxenite that is the hanging wall in the central portion of the Roby Zone, in contact with a barren gabbro hangingwall. Only a 400 metre portion of the

pyroxenite is mineralized, the strike extensions are barren. The pyroxenite is planar, trends at 341°, is subvertical to a depth of 250 metres and shallows to the east at greater depth.

Higher PGE grades (6 to 12 g/t Pd) are achieved in those portions of the pyroxenite that are altered to a schistose assemblage of actinolite-talc-chlorite (Shear Ore). The sulphide content of the pyroxenite, including the contained shears, is 1% to 3% with local patches of net texture with up to 10% sulphide. The PGE tenor is not proportional to the sulphide content. The dominant sulphides are pyrrhotite, chalcopyrite and pentlandite. PGM are vysotskite (Pd,Ni)S, isomerticite PdIISb₂As₂, kotulskite Pd(Te,Bi), sperrylite PtAs₂, Merenskyite PdTe₂, and a Pd arsenide, either Stillwaterite or Palladoarsenide. The higher-grade segment of the pyroxenite is juxtaposed with the higher grades found in the adjacent breccia ore. The PGE tenor of the pyroxenite declines dramatically to the south where the degree of talc alteration is diminished. It also declines in the adjacent breccia ore. The pyroxenite and its contained shears pinch out to the northwest. The Shear Ore has been drilled systematically to a depth of 500 meters. The pyroxenite separates the two other major ore types.

North Roby Ore

North of the pyroxenite, the North Roby Ore is a tabular zone whose gradual boundaries are defined by assay. It is 20 to 40 meters thick, and 200 meters long. The zone strikes at 20° with its footwall dipping to the east at 45° to 60° and the hanging wall steeper. It is hosted by a wide variety of lithologies. At surface, it is dominated by coarse grain leucogabbro containing irregular masses of vari-textured gabbro and medium to coarse grain gabbro-norite. The proportion of noritic rock, and heterolithic breccia increases with depth. The sulphide content of this mineralization ranges from trace amounts to 4%, and typically is less than 0.25%. The sulphide poor mineralization cannot visually be distinguished from the barren wallrock. The dominant sulphides are pyrrhotite, pentlandite, chalcopyrite and pyrite. The PGE distribution is erratic (1 to 16 g/t Pd) and weakens in a NE direction. Systematic drilling of this zone has only attained a maximum depth of 250 metres.

Breccia Ore

Southwest of the Shear Ore, Breccia Ore is contained within a mineralized complex that measures 550 by 350 meters. This complex contains intrusive lithologies that range in composition from clinopyroxenite to anorthosite to norite. Textures include equigranular fine to coarse grain, porphyritic, pegmatitic, vari-textured and heterolithic-gabbroic-breccia. Mineralization is concentrated in these last three textural types, but it is also commonly found in clinopyroxenite, melanogabbro, medium grain norite and medium and coarse grain gabbro-norite. All of these rocks types occur as discontinuous, irregularly shaped pods with maximum dimensions of 60 meters that, when viewed on a map, are in fact a larger scale of breccia.

The east boundary to the central portion of the Breccia Ore is defined by the sub-vertical contact with the pyroxenite. The southeast extension of the Breccia Ore remains open. The southern and western boundaries of the Breccia Ore are all defined by a gradual decline in PGE content. As defined by the 0.7 g/t Pd assay cutoff, the western boundary is sub-vertical and the southern boundary dips steeply to the north. Due to the gradual nature of the PGE distribution, these boundaries are subject to changes in assay cutoff. The greatest potential for expansion of resources at the Roby Zone, assuming similar continuity of grade, is deepening the extent of the Breccia Ore.

The most definitive lithological control on PGE grade within the Roby Zone is the presence or absence of heterolithic-gabbro-breccia and accompanying varitextured gabbro. The breccia is always ore, and other lithologies are only ore grade when in proximity to breccia. However, mineralized breccia on the margin of the Roby Zone can contain less than 0.7 g/t Pd. Also, with the exception of post mineralization dykes, every lithology can either be ore or waste. Heterolithic-gabbro-breccia is composed of fragments a few centimetres to several metres in size that are representative of most of the lithologies in the area, as well as a few exotic clast. This breccia has a wide spectrum of characteristics. It can consist of sub-angular fragments crowded together with very little matrix and can grade to a breccia consisting of 10% sub-rounded fragments (digested xenoliths) and 90% matrix. A transition from matrix poor to matrix rich breccia is observed in the southern Breccia Ore. Melanogabbro is the matrix and the intrusive magma creating the breccia. Observed gabbroic composition for the matrix in many locations is interpreted as a composition change due to digestion of less mafic xenoliths. Where selectively assayed, the matrix will contain 8 g/t Pd, and the fragments 0.08 g/t Pd. This explains the observation that matrix poor breccia is usually low grade. Pegmatitic gabbro occurs on the rims of the fragments, as irregular patches and veinlets throughout.

Breccia is transitional into varitextured gabbro and lacks distinct xenoliths. Varitextured gabbroic rock range in composition from leucocratic to melanocratic, and some varieties are fine grain with medium grain patches and veinlets, medium grain with coarse and pegmatitic patches and veinlets, and coarse grain with pegmatitic patches and veinlets. Pegmatitic rock is more abundant in the central Breccia Ore. Here it occurs as convex multiple layers, dykes and lenses up to several metres in width but not laterally continuous. The common pairing of the breccia and varitextured gabbro, and its associated pegmatitic gabbro not only key to interpreting the process that created the ore, but also serves as an important association that guides exploration.

The Central Breccia Ore contain up to 5% chalcopyrite, pentlandite, pyrrhotite and pyrite. It occurs as fine grain disseminations throughout, and as coarse blebs interstitial to pegmatitic clinopyroxene and plagioclase. The fine sulphides are contained as streaks within the clinopyroxene pseudomorph that are now composed of actinolite, chlorite and talc. This late introduction of sulphide is synchronous with the alteration of pyroxene to actinolite.

Twilight Zone

The Twilight Zone is a new zone in the process of being delineated by diamond drilling. This zone is separated from the Roby Zone on its western boundary by a 50 to 70 metre thick unit of massive homogenous barren gabbro and separated from the southeastern extension of the Roby Zone by a 50 metre thick barren dyke. Twenty-two diamond drill holes to date have delineated a zone 100 metre wide, 200 metre deep and 200 metre long. The zone remains open to the east, and south. This area has the same style of lithological complexity as the Roby Zone, but is dominated by a less heterolithic gabbro-norite-breccia, and does not contain varitextured and pegmatitic gabbro. The most significant mineralization discovered in this area to date was intersected by holes 99-040 and 99-171. This mineralization, containing 0.25% to 0.5 % pyrrhotite and chalcopyrite, assayed 2.38 g/t Pd over 116 metres.

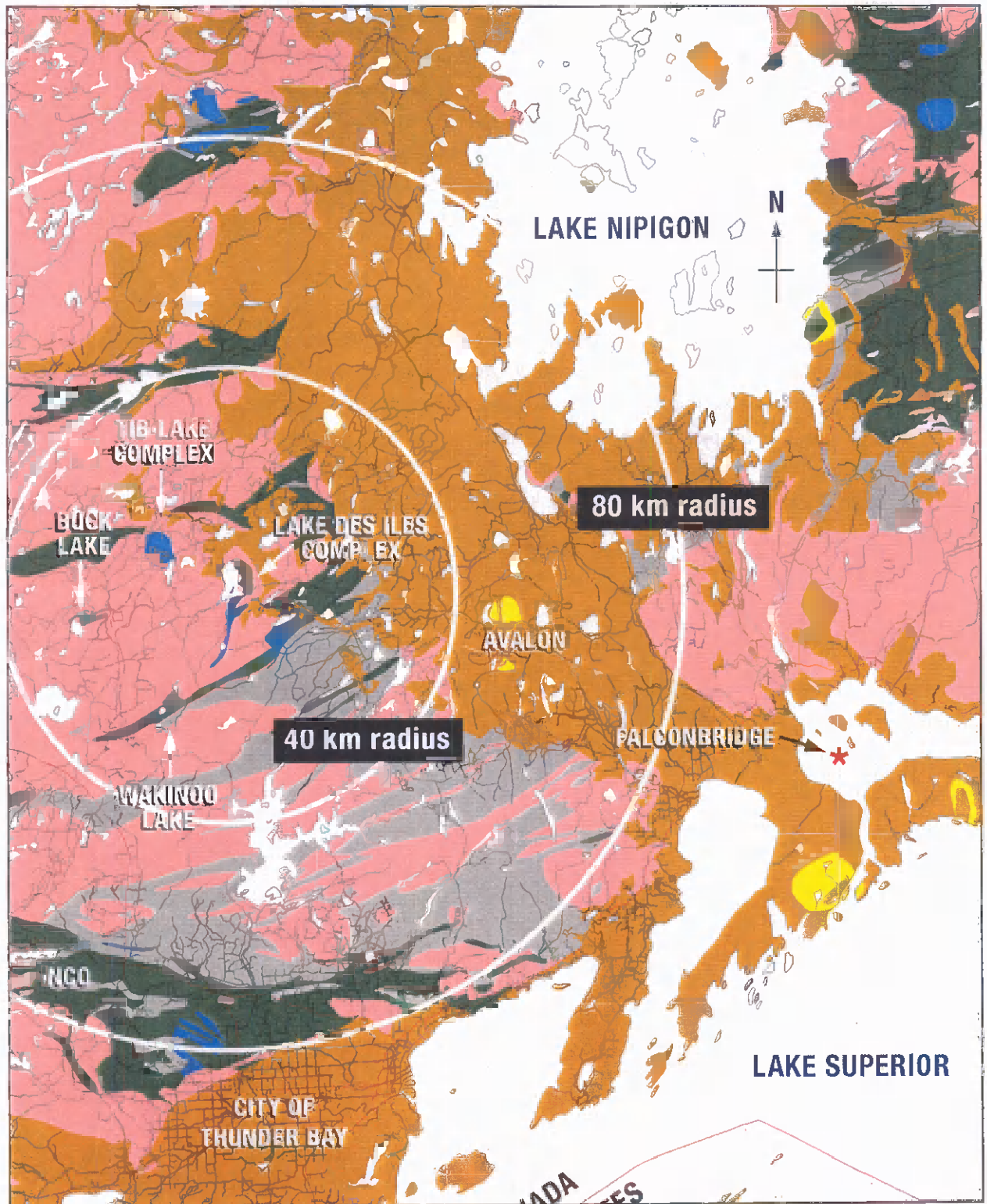
Baker Zone

The Baker Zone is one kilometer east of the Roby Zone. Detailed surface sampling has defined a mineralized zone that is 100 meters wide and traced continuously for 250 meters. This northeast-striking zone is defined by assays greater than 0.68 g/t Pd and generally contains less than 1% sulphides. The mineralized rock consists of gabbro, varitextured gabbro, pyroxenite, and heterolithic-gabbro-breccia. To date, the highest grades are associated with a pyrrhotite-chalcopryrite rich area at a breccia/gabbro contact. Channel samples over a 15 meter length assayed 4.69 g/t Pd, 0.27 g/t Au, 0.48 g/t Pt, 0.43% Cu and 0.39% Ni. A 3500 metres (20 holes) diamond drilling program conducted during the winter of 1998-99 tested the zone over a 250 metre strike length, to a maximum depth of 200 metres, at 50 metre spacing.

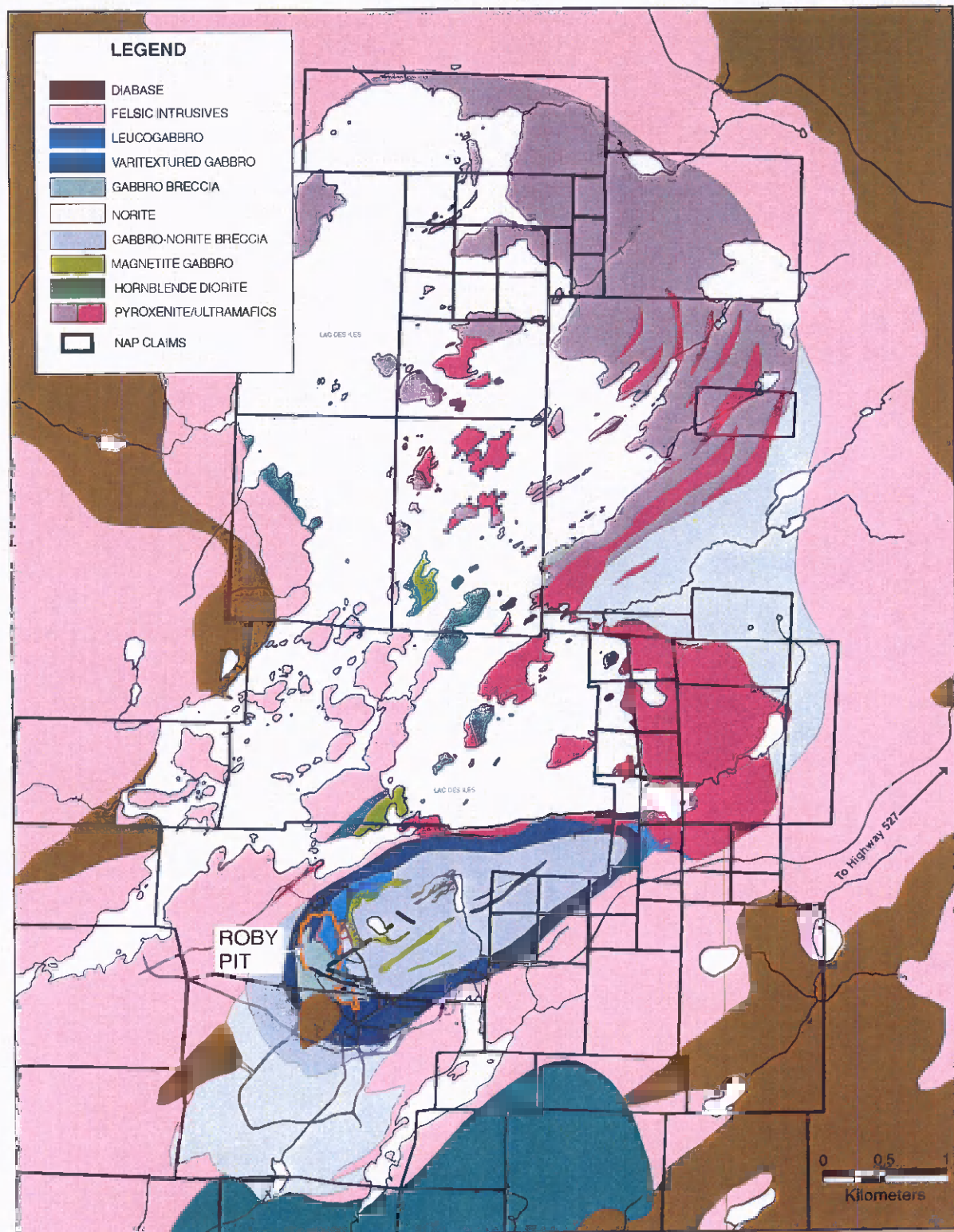
Moore Zone

The Moore Zone is a poorly exposed mineralized area in a swampy area 600 metres south of the Roby Zone boundary. Mineralization has been uncovered in an area 100 by 200 meters and appears to have a NW strike based on the alignment of a sulphide rich zone. Where best exposed the mineralization consist of heterolithic-gabbro-norite-breccia with less than 1% disseminated pyrrhotite and chalcopryrite. An interval of norite within the breccia contains 2 to 5% pyrrhotite and chalcopryrite. The best assay interval generated from diamond drilling is 24.3 meters grading 1.39 g/t Pd+Pt+Au, 0.291% Ni, and 0.312% Cu. On the basis of 17 diamond drill holes, it is interpreted that the Moore Zone is rod shaped, and has maximum dimensions of 22m x 45m x 90m deep.

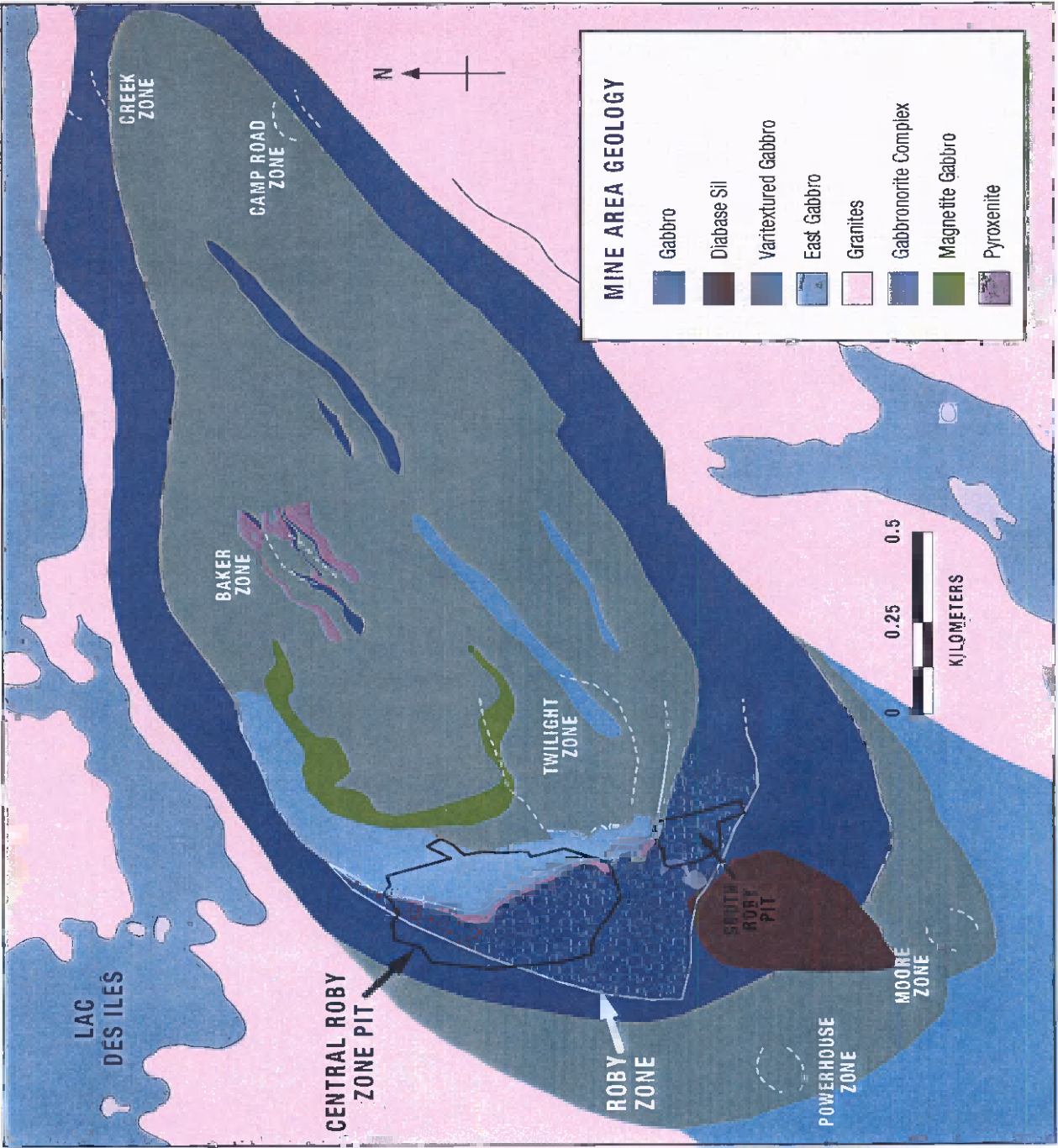
Geology of the Lake Superior Region



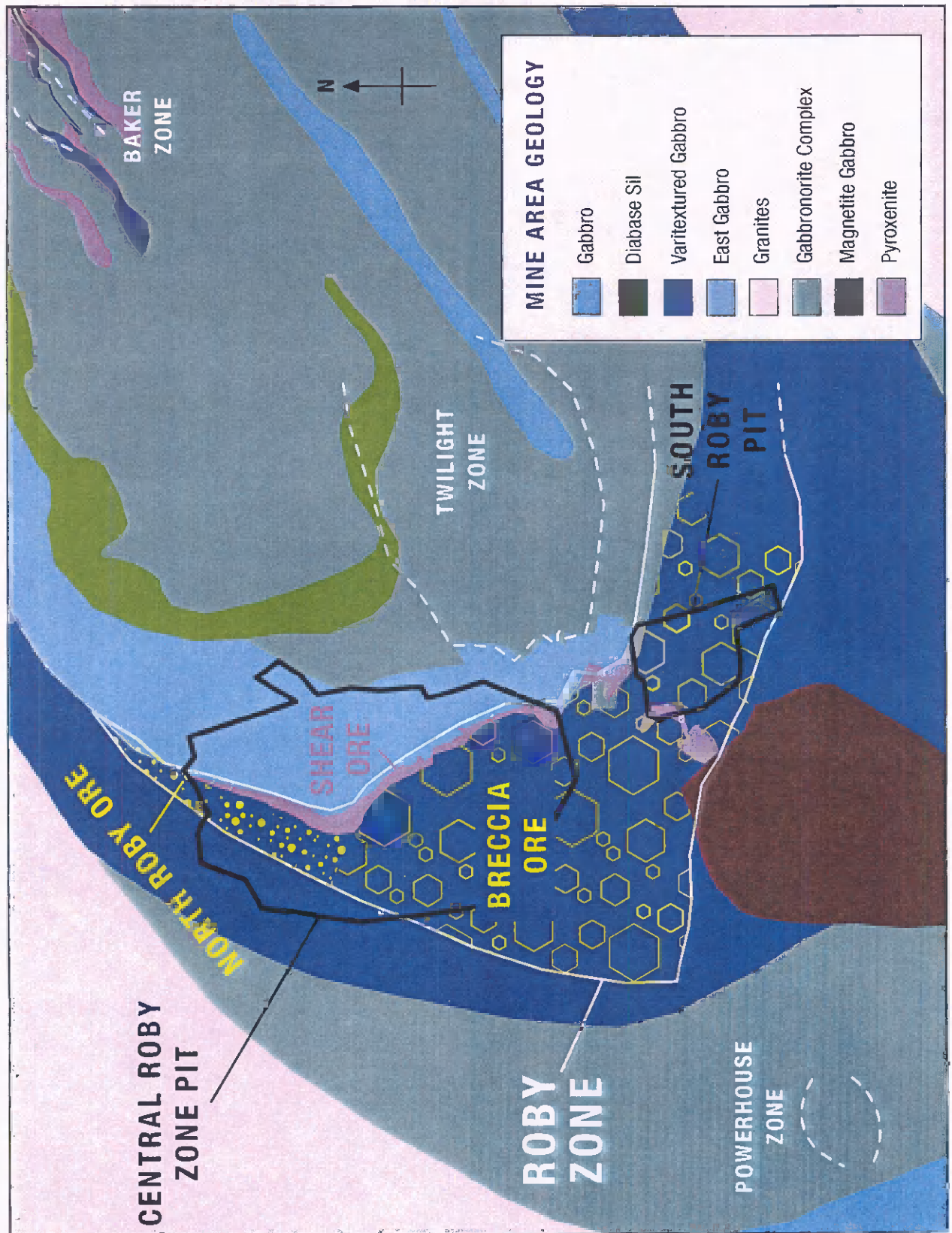
Geology of the Lac des Iles complex



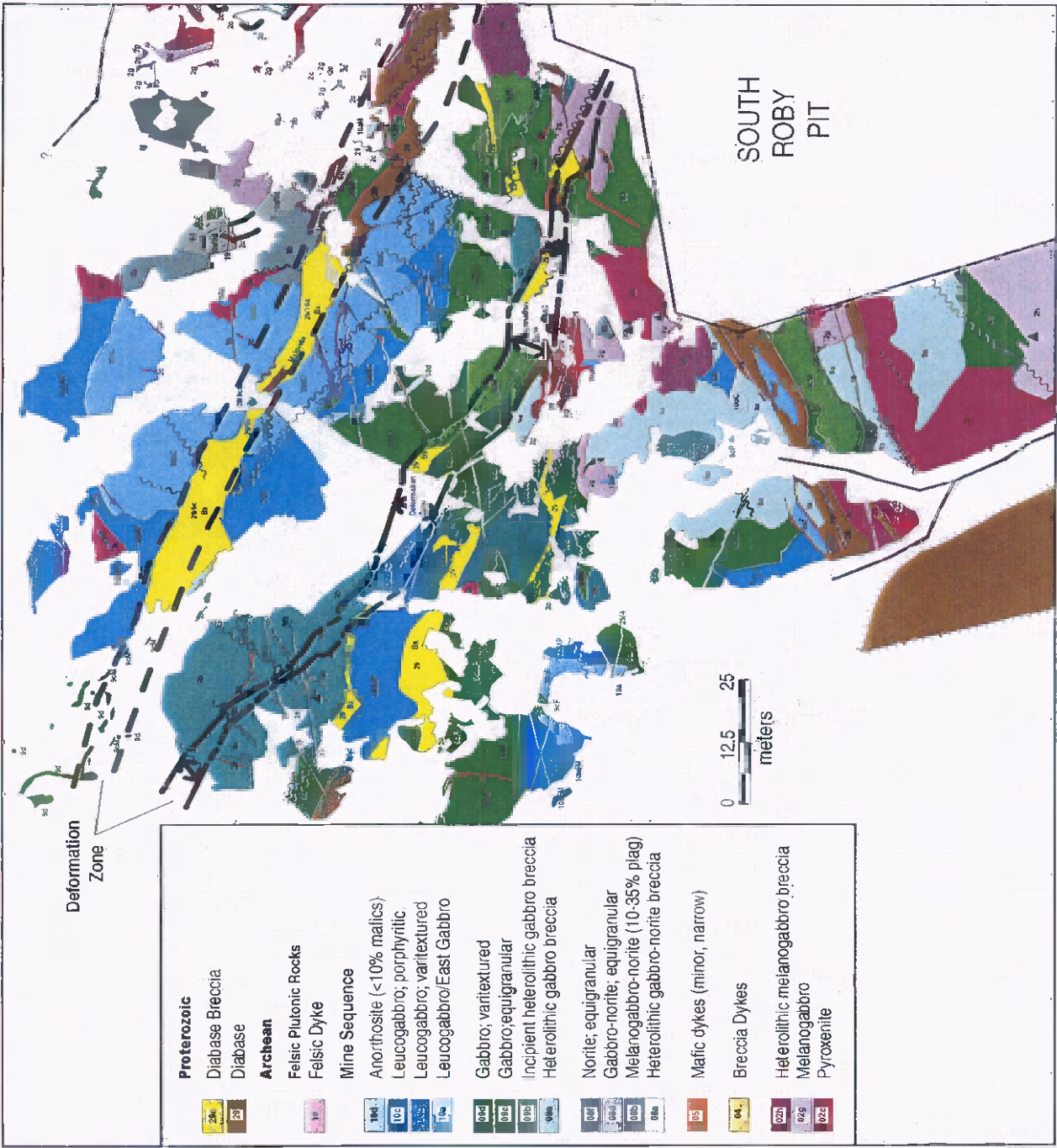
Geology of the mine complex



Geology of the Roby Zone

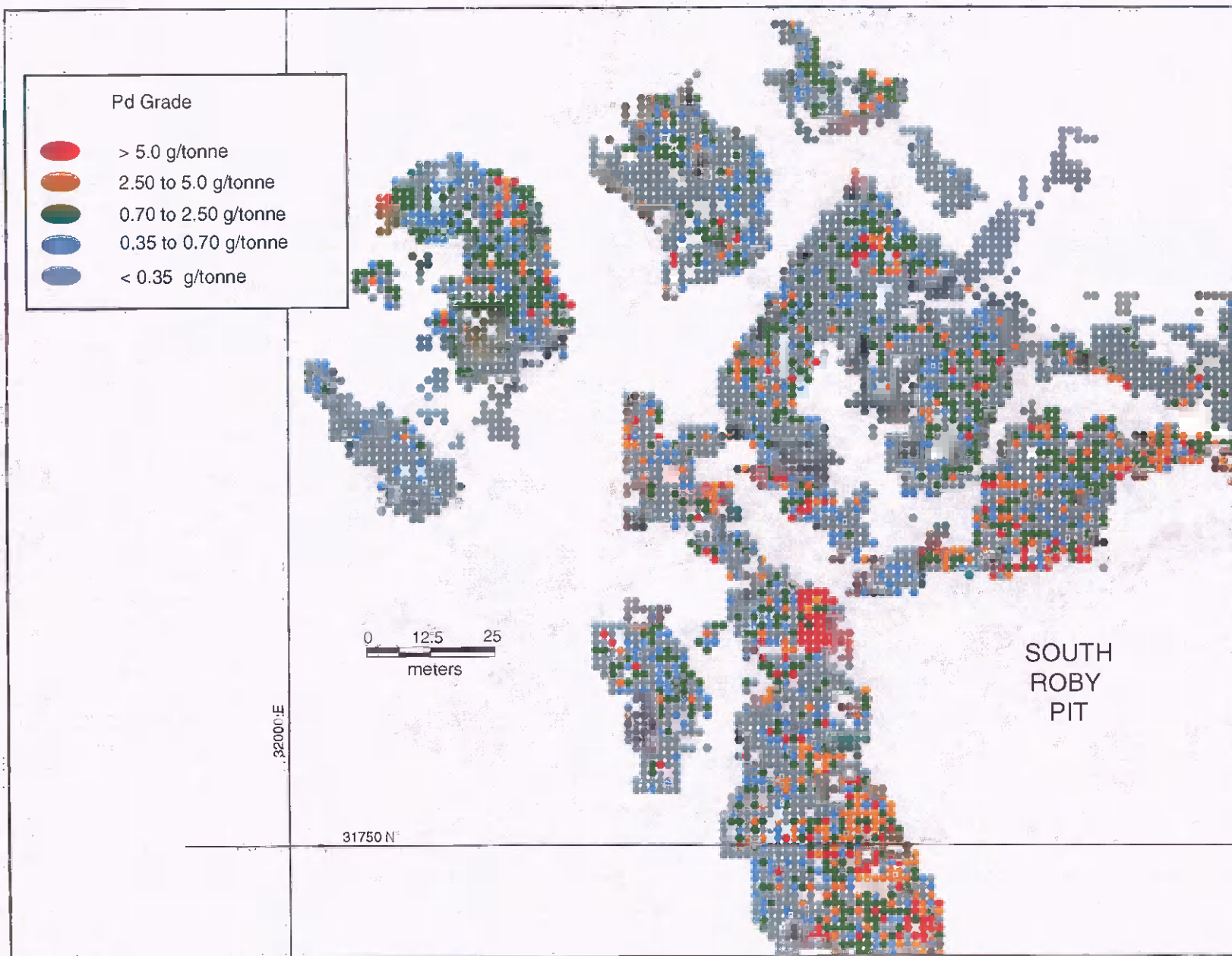


Detailed geology, South Roby Zone



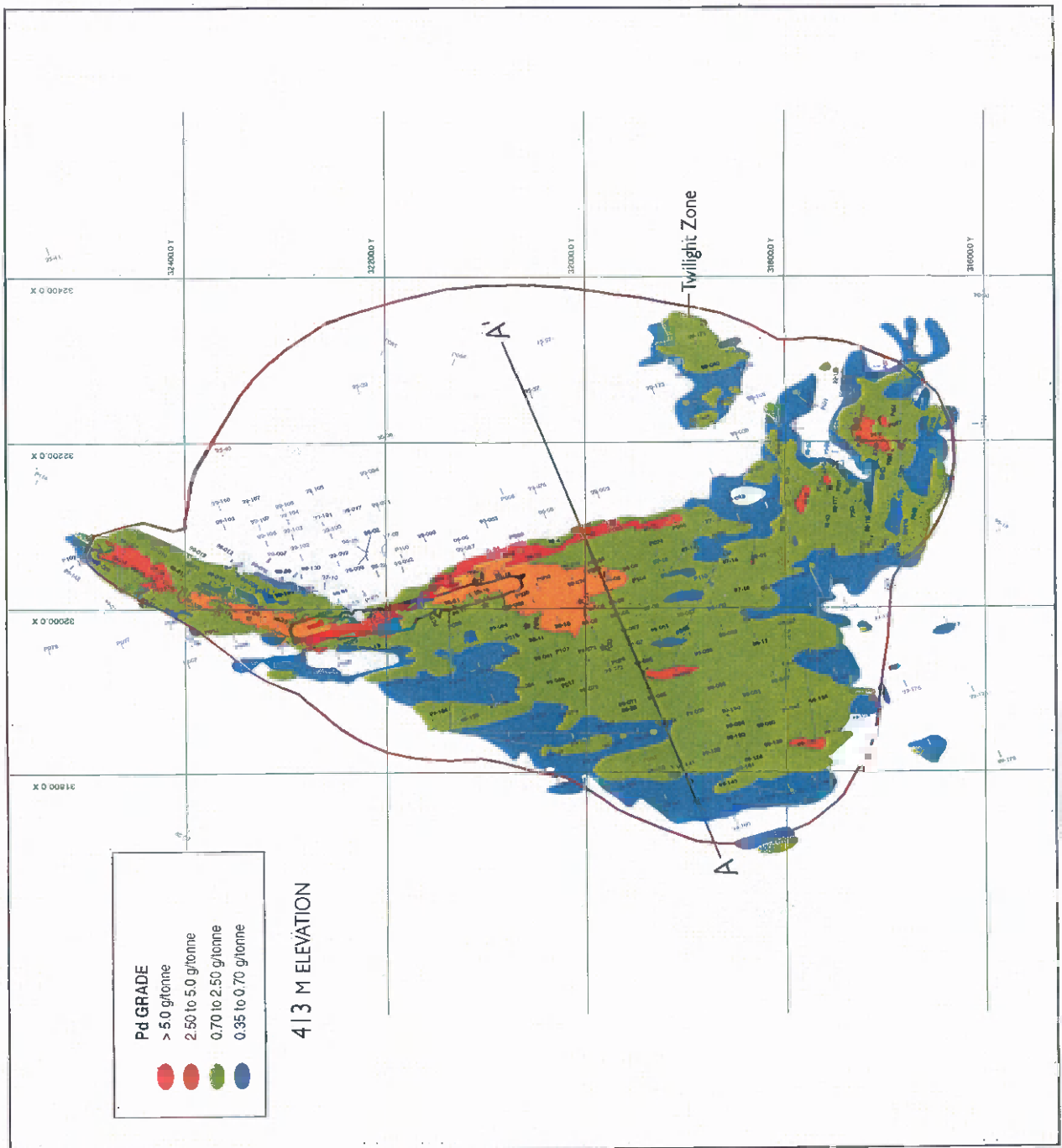
Surface Pd assay, South Roby Zone (overlay on "Detailed geology, South Roby Zone")

PAGE 13

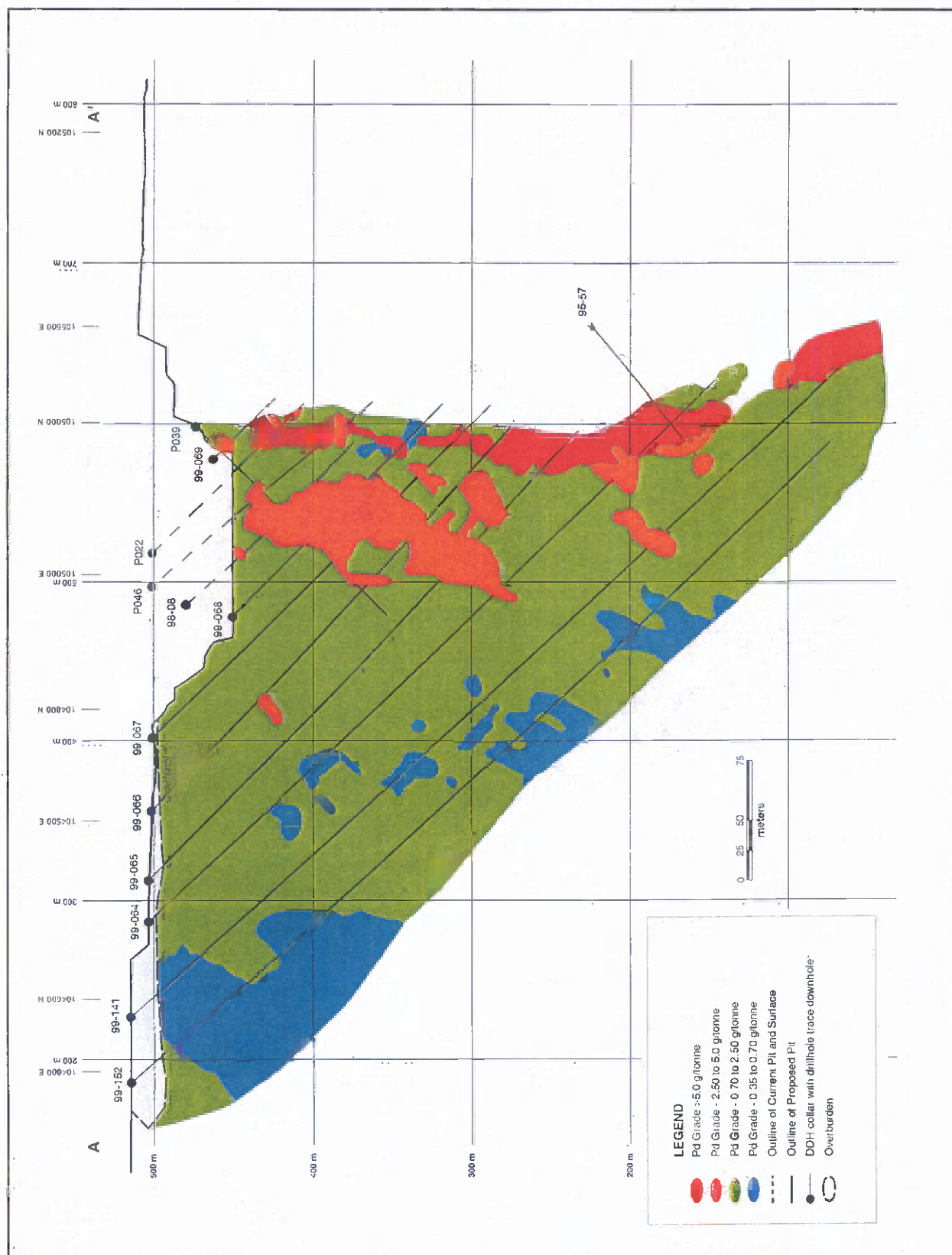


METALS FOR CLEAN AIR

Roby Pit Level Plan Grade Model



Roby Zone Grade Model Cross Section 509 N



GEOARCHAEOLOGY AND DEGLACIATION HISTORY OF THE THUNDER BAY AREA

Brian Phillips, Lakehead University

Scott Hamilton, Lakehead University

Joe Stewart, Lakehead University

Pat Julig, Larentian University

**Bill Ross, Ontario Ministry of Citizenship,
Culture and Recreation**

Soon after deglaciation, native peoples established habitation sites. This trip will examine how the local geomorphology controlled the location of these sites and the nature of human activities.

Forty-Sixth Annual Meeting
Institute on Lake Superior Geology
Thunder Bay, Ontario, May 8-13th, 2000

PRE-MEETING FIELD TRIP 3

Geoarchaeology of the Thunder Bay area - Field Guide

Dr. Brian Phillips, Department of Geography, Lakehead University.
with

Dr. Joe Stewart, Department of Anthropology, Lakehead University.
Dr. Scott Hamilton, Department of Anthropology, Lakehead University

Dr. Pat Julig, Department of Anthropology, Laurentian University
Mr. Bill Ross, Regional Archaeologist, Ministry of Culture and Recreation.

Objectives:

The field trip will focus on the deglaciation and lake level history of Thunder Bay and the area immediately to the west of the city - the Kaministiquia river valley, Kakabeka Falls and the Marks Moraine. In particular, evidence of Paleoindian presence associated with the abandoned shorelines, river mouths and deltas of the Lake Minong stage of Lake Superior (circa 9.5 ka B.P.) will be examined (Figure 1).

The trip will start by a visit to the former mouth of the Current River at the north end of the city, where the Simmonds and McDaid sites, both surface sites, are located. Then the trip will head west, up valley, to the more famous Cummins site, a stratified site, part of which is preserved within a fenced area. The third area to be visited will be the Rosslyn delta, on the Kam river, the western extent of Lake Minong. From there the trip will travel west to Kakabeka, where earlier Lake Beaver Bay features will be examined. Finally, the trip will mount the Marks moraine, providing a view that will place the days observations in context with pre and post Marquette ice marginal events and Paleoindian presence.

Introduction

There is a strong though not exclusive relationship of Paleoindian habitation in the Thunder Bay area with the shoreline features of the post-glacial lakes of the Superior basin, particularly those of Lake Minong, established about 9.5 ka B.P. This is perhaps illusory in that they occupied a number of habitats, some far from a lakeshore, but in geomorphological mapping of these shorelines, it is these sites which have been most commonly found and reported.

Evidence accumulated since early excavations by MacNeish (1952) by both amateur and professional archaeologists (Fox 1975, 1980; Dawson 1983; Julig, 1984) suggest that Plano Indians migrated north eastwards from the area of Dakota and Minnesota into the cul-de-sac formed between Lake Agassiz, which covered most of Manitoba and parts of Northwestern Ontario at its maximum extent, the lakes of the Superior basin and the retreating margin of the Laurentide ice sheet. These people, derived from plains cultures characterised by a fluted point lithic

Figure 1 Route Map



technology, probably followed game into the area as hunters, but in staying, diversified in their economic activities and in their tool-kit design, to form a recognisable "Interlakes Composite" (Ross, 1994) characterised by a parallel flaked point lithic technology, making particular use of the red (jasper) taconite material found in the siliceous Gunflint formation that outcrops in the Thunder Bay area.

Deglaciation History

The broad details of the deglaciation of the Superior basin are shown in Figure 2. As ice of the Rainy River and Superior lobes withdrew from central Minnesota, a series of recessional moraines were left in place. The Vermilion moraine, trending northwest to southeast, across northeastern Minnesota, was followed by the Steep Rock moraine on the Canadian side of the border, and the Brule moraine (Figure 3). Guided by these ice marginal positions, Lake Agassiz found an early eastern outlet through the Arrow/Whitefish lakes corridor (circa 11 ka B.P.) and, shortly after, through the Shebandowan lake corridor, ultimately using the Lake Nipigon spillways around 10.4 ka B.P. (Figure 2a) to enter Early Lake Minong (Teller, 1985) which occupied the Superior basin.

There is no reason to believe that the warming climate did not permit environmental conditions to become suitable for the immigration of animal herds and prehistoric peoples into the Thunder Bay region at this time. Only tenuous evidence of such early occupation has been reported (Phillips, 1993; Ross, 1994), though recent work in the Arrow/Whitefish corridor has confirmed a number of sites that most likely represent an earlier Paleoindian presence. (McLeod and Phillips, pers. communication, 1999).

A major event in the deglaciation history of the Lake Superior basin was the Marquette readvance (9.9 ka B.P.), during which the basin and its margins were briefly reoccupied by ice (Figure 2b). Recessional moraines northeast of the Brule moraine were destroyed and more recent features formed in their place. This event also has considerable archaeological implications, since any evidence of occupation prior to the Marquette readvance is likely to have been buried or destroyed completely by the new ice cover.

Hudson Bay ice pushed southwestward into the Lake Superior basin, to halt on its southern shore (Drexler, Farrand and Hughes, 1983). Only in a small portion of Whitefish Bay near Sault Ste. Marie, in the southeast corner of the basin, did Early Lake Minong remain an open lake (Farrand and Drexler, 1985). The Marquette lobe pushed westward against the steep Minnesota shore and, where less obstructed, flowed northwest up the Kaministiquia valley to the Marks moraine (Figure 3). This prominent ridge, rising to over 470 m (1550') in places, curves from Lappe through Mokomon and around the northwest of Kakabeka towards the Pigeon River (Zoltai, 1963, 1965a, 1965b; Burwasser, 1977, 1980; Burwasser and Ferguson, 1980). Contemporaneously, to the west of the Lake Superior basin, the Patricia ice lobe pushed towards the Rainy River district and halted to deposit the Dog Lake moraine. This runs northwest from Lappe and holds up Hazlewood, One Island, Hawkeye and Dog lakes. East of Lappe, where the two moraines meet, a line of glacial debris known as the Mackenzie interlobate moraine can be traced

Figure 2

The Glacial Period and Early Peoples

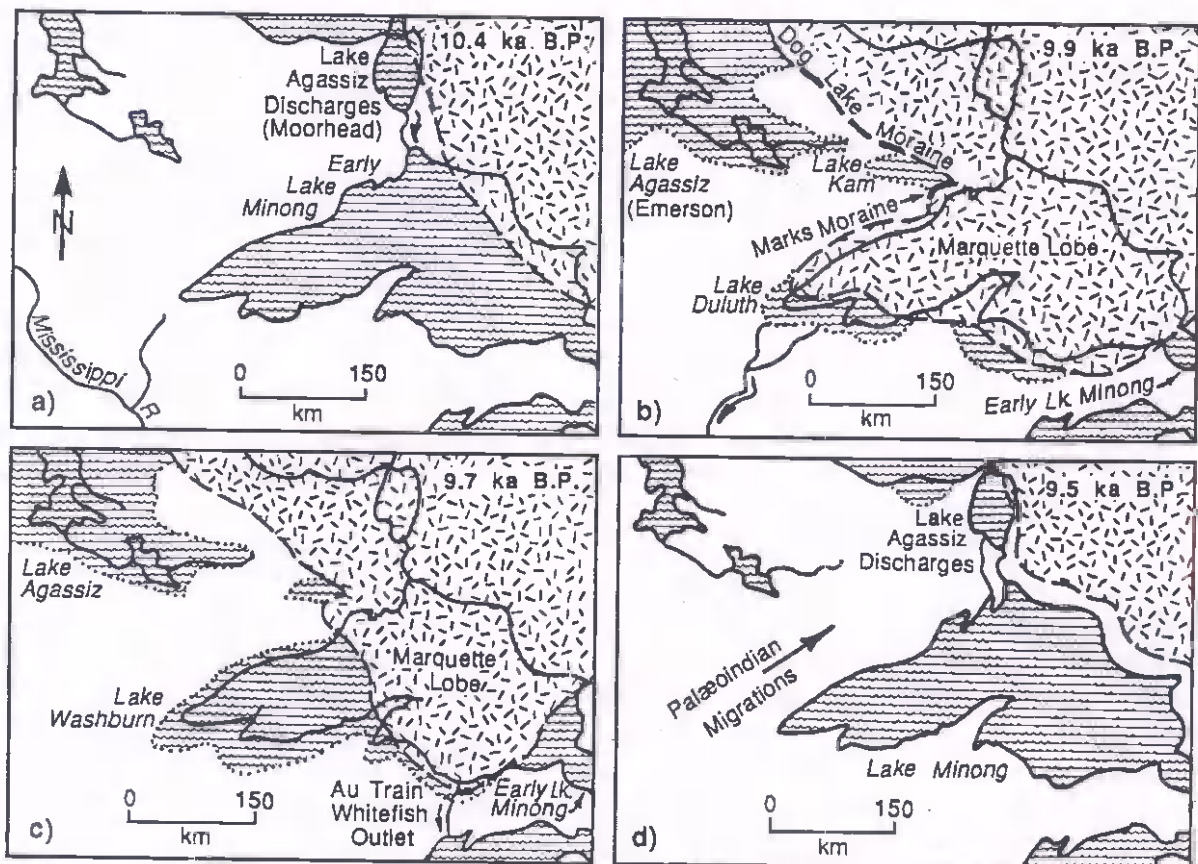


Figure 1: The retreat of the glaciers in the Superior region between 10,400 and 9,500 year ago. (from Phillips, 1993, p. 95)

Figure 3

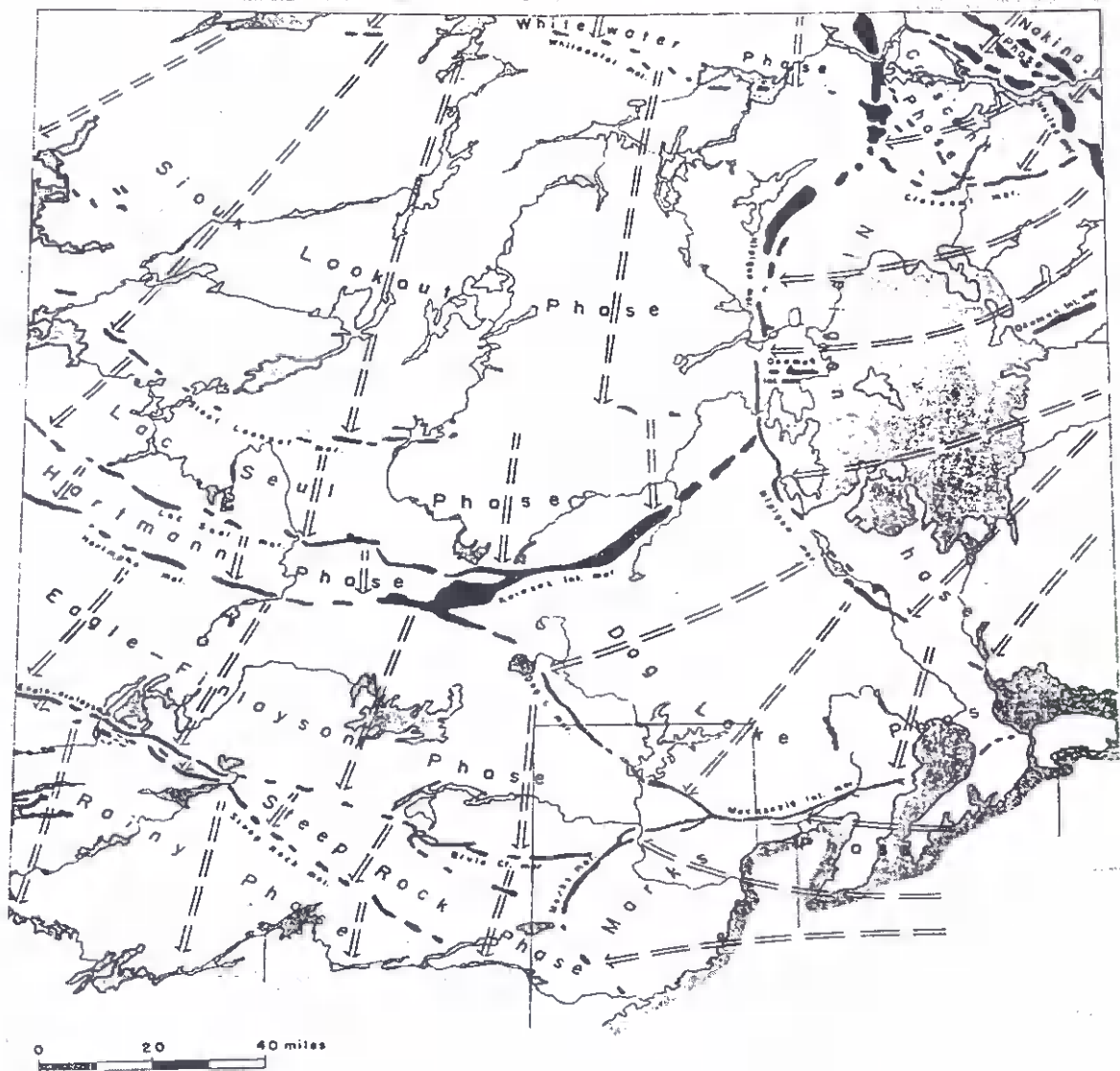


FIG. 5. Map of the Quetico-Nipigon area showing moraines and direction of ice movement during various phases.

from Zoltai, 1965a)

through Pearl and on to the Black Bay peninsula. A glacial lake, Lake Kaministiquia, was formed between the two ice margins (Teller, 1985) and, at Lappe, a huge sand and gravel delta was built by sediment pouring off the interlobate moraine. This scenario is still the subject of debate, however, and there is some field evidence that the Marks and Dog Lake moraines may be older features that were simply reoccupied by ice of the Marquette advance (Julig, McAndrews and Mahaney, 1990; Tickle, 1996; Noble, pers.comm.1999). Lake Agassiz, deprived of its eastern outlets by the advance of the ice, expanded in area and depth (the Emerson phase), and flooded catastrophically through the Clearwater and Athabasca valleys in Saskatchewan and Alberta into the Mackenzie river and the Arctic ocean (Smith and Fisher, 1993).

The Marquette readvance obliterated evidence of the earlier phase of Lake Superior's shoreline history in most of the basin. Shoreline sites that prehistoric people might have occupied in the area of Thunder Bay before 9.9 ka B.P. may lie buried beneath Marquette deposits. In the four hundred years between 9.9 and 9.5 ka B.P., a period about which much more is known, ice withdrew from the Lake Superior basin (Figure 2c/2d). Then, again, Paleoindian people followed game up the Interlakes corridor and into the Thunder Bay region, this time to settle, in part at least, on the shores of Lake Beaver Bay (Stuart, 1993) and Lake Minong (Phillips, 1988).

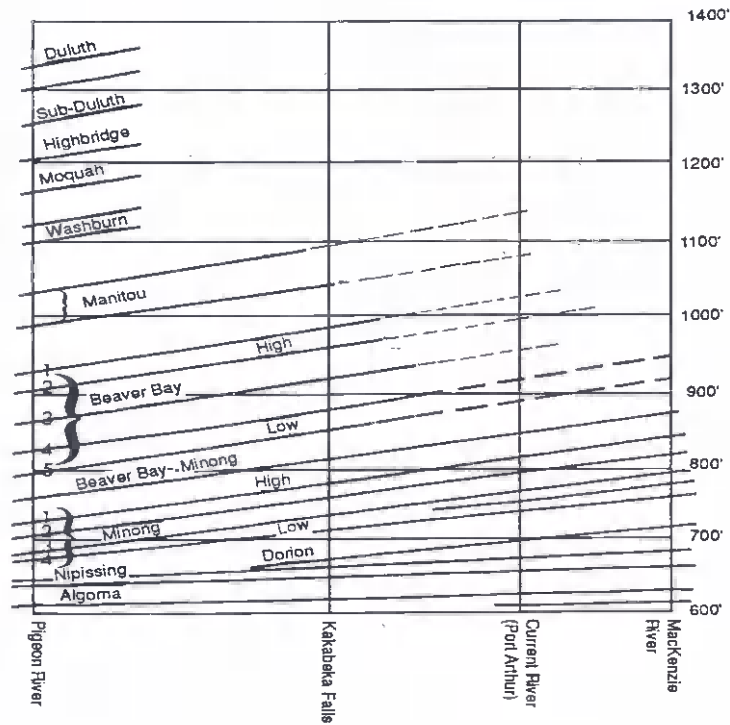
Shoreline History

At the peak of Wisconsinan glaciation, the northeastern margin of the Lake Superior basin was depressed by the weight of ice (isostatic depression) to a greater degree than the less heavily ice-loaded southwestern side. As a result, "rebound" (isostatic recovery) since that time has been greater on the north shore of the Lake Superior basin than on the south shore. Lake shorelines which had been originally horizontal became progressively tilted along an approximate northeast axis, such that a shoreline of the same chronological age increases in altitude from southwest to northeast along the western shore of Lake Superior. A theoretical archaeological site at Grand Marais, Minnesota, found at 184 metres, just above the present storm beach, will be of similar age to another theoretical site in Terrace Bay, Ontario, at the 300 metre contour (117 metres above the present lake), with both sites lying on the same tilted shoreline (Figure 4).

As the Marquette ice lobe wasted back, the eastern and western sides of the basin were exposed as separate entities (Figure 2c). On the eastern side, a fairly stable Lake Minong, controlled by the height of the St. Mary's River outlet at Sault Ste. Marie, extended northwards up the Ontario shore and westwards along the Michigan shore as they were exposed by melting ice (Farrand and Drexler, 1985). In the enclosed western end of the basin, a distinct succession of ever larger but progressively lower level lakes formed (Lakes Duluth, Highbridge, Moquah, Washburn, Manitou, and Beaver Bay), each extending further northeast up the Minnesota - Ontario shore and east along the Wisconsin shore (Farrand and Drexler, 1985). Along the land-based margins of these water bodies were formed various coastal features, such as bluffs and beaches, by which the shorelines can be traced today. As ice vacated the basin, the eastern and western lakes were

Figure 4

Shoreline Diagram for the Pigeon River/Thunder Bay Area



(from Farrand, 1960)

Figure 5

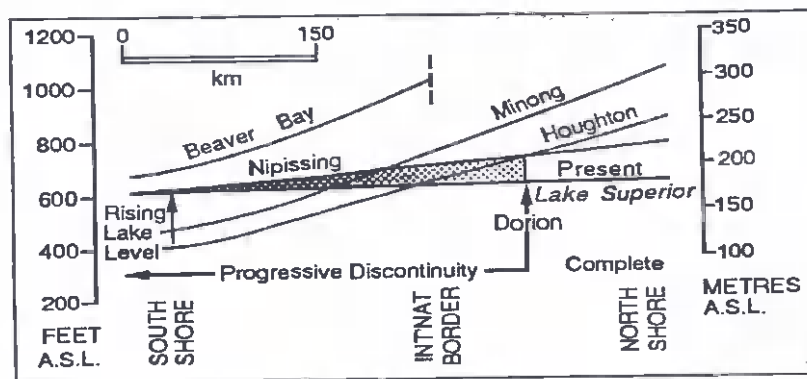


Figure 10: The pre-Nipissing transgression and the resulting loss of potential archaeological sites along the shores of Lake Superior. (from Phillips, 1993, p. 100)

united and the single shoreline of Lake Minong was formed around the basin about 9.5 ka B.P. (Figure 2d). For the next 1,500 years, water-levels in the Lake Superior basin declined as the St. Mary's sill was eroded down to bedrock. A staircase of Post-Minong shorelines were formed, the last and lowest of which was Lake Houghton, about 8.0 ka B.P.

By 8.0 ka B.P., due to isostatic uplift, the rising levels of Lake Huron had flooded into the St. Mary's River and reversed the flow (Larsen, 1987). This back-flooding led to slowly rising water-levels in the Lake Superior basin and culminated in the Nipissing lake stage around 5,000 years B.P.. Known as the pre-Nipissing transgression, this rise in water-level was imposed upon a still-tilting basin. On the north shore, east of Dorion, the rate of isostatic uplift remained more rapid than the rising waters of the pre-Nipissing period, with the result being that the shoreline marking the Nipissing maximum level lies at a lower altitude than all older shorelines, including that of Lake Houghton. On the south shore the pre-Nipissing transgression was more rapid than isostatic recovery, and wave action "inherited" the features of older shorelines, modifying and destroying portions of them, including pre-existing shoreline archaeological sites (Phillips, 1977). This causes a chronological discontinuity which is at its greatest near Duluth and decreases towards Dorion where the Houghton shoreline appears above the present water-level (Figure 5).

In Thunder Bay, evidence of Paleoindian activities would normally have been traced as a continuum from the high Minong shoreline to the Houghton (which lies just below present water-level), but the pre-Nipissing transgression reoccupied the lower and later part of that record and cut a prominent bluff on which the General Hospital, the Court House, St. Joseph's Hospital, and Lakehead University are built. The Nipissing shoreline can be traced up the valley towards Mapleward road and across towards Mount McKay. Thus, the later part of the Paleoindian record, the important transition into the Archaic period, and a good portion of the early Archaic is missing in Thunder Bay.

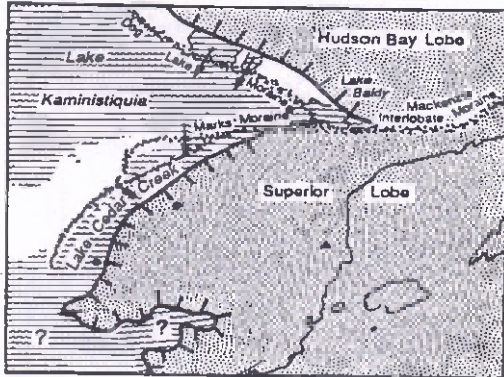
Local History

Figure 6 represents an interpretation of the detailed history of the withdrawal of Superior (Marquette) ice from the Thunder Bay area, based on currently known information, though subject to revision. As ice wasted back from the Marks Moraine, shorelines on the south side of the moraine show that a body of water collected between the moraine and the ice front (Figure 6a). This proglacial lake has been named Lake Cedar Creek (Jahnke, 1993), and there is tenuous evidence that it connected with high level lakes to the south, perhaps of Lake Duluth equivalence. As ice withdrew from the Kaministiquia embayment a small readvance to the Intola Moraine occurred (Figure 6b). The higher levels of lake Beaver Bay have been traced into this moraine (Stuart, 1993), and it is likely that Paleoindian peoples entered the area at about this time, probably using the Marks moraine as a causeway. As ice withdrew further and water level declined to the lower levels of lake Beaver Bay (Figure 6c), the possibility of Paleoindian occupation increases, and by the time Lake Minong was established in the Kaministiquia embayment, there is plenty of evidence to prove their presence

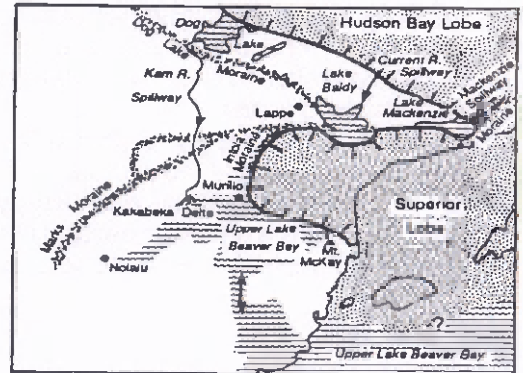
Figure 6

An interpretation of post-Marquette history in the Thunder Bay area.

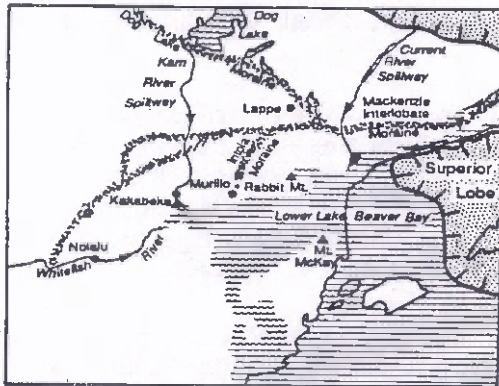
(a) Marquette Ice and Proglacial Lakes area
9,800 years B.P.



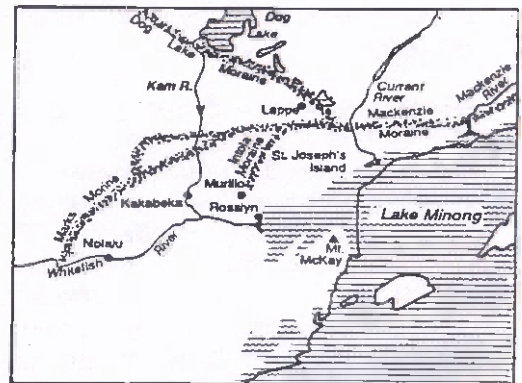
(b) Intola Moraine and Upper Lake Beaver Bay circa
9,700 years B.P.



(c) Lower Lake Beaver Bay circa 9,600 years B.P.



(d) Lake Minong, 9,500 years B.P.



(from Phillips, Hill, Fralick and Ross, 1994)

(Figure 6d). The shoreline diagram (Figure 4) shows that only as ice withdrew from the Thunder Bay region did the sequence of post-glacial lakes extend into the area. Beaver Bay shorelines can be found at Kakabeka, but only the lower levels extend eastwards through the city towards the Mackenzie River.

A few notable sites associated with the Minong shoreline are the Simmonds and McDaid sites at the mouth of the Current River, the Biloski, Catherine and Cummins sites along the north side of the Kam embayment, and the collection of sites on the Rosslyn delta at the mouth of the Kaminitiquia River, though here the association is less simple and deserved more analysis (Hamilton, 1996). Some Paleoindian sites unrelated to shorelines occur at High Falls on the Pigeon river, on Dog Lake (McLeod, 1982), at Harstone Hill, near the junction of the Whitefish and Kaministiquia rivers and near Kakabeka Falls where the Kaministiquia river might have been most logically crossed. The Crane site near Kakabeka, found in a vegetable garden, provides a rich cache of beautifully crafted 10 cm bifaces in a location apparently unrelated to any topographic feature, though on the surface of one of the higher Beaver Bay terraces.

Because of their antiquity and the acidic nature of the Boreal forest soils, only the lithic materials remain to be found at these sites (Figure 7), though it is very likely that many other natural materials would have been used by these peoples.

Field Excursion Stops

Stop 1 - Hillcrest Park Lookout.

The route from the University will be down Oliver Road and, just below the Nipissing bluff, left on High Street and up the hill for a brief visit to Hillcrest Park from where the city can be seen in context with local topography and Lake Superior.

Stop 2 and 3 - Boulevard Park, the Bluffs and Centennial Park Rd.

The route will head northeast on High St., over the St. Joseph/Hillcrest island of Minong times, left on Balsam St, right on Hudson Ave and onto Arundel St. from which a left turn before Black Bay bridge will lead to the Bluffs Scenic Lookout. Walk down to Simmonds site. Then board to cross Black Bay bridge and turn left on Centennial Park Rd to visit the McDaid site, adjacent to the roadway.

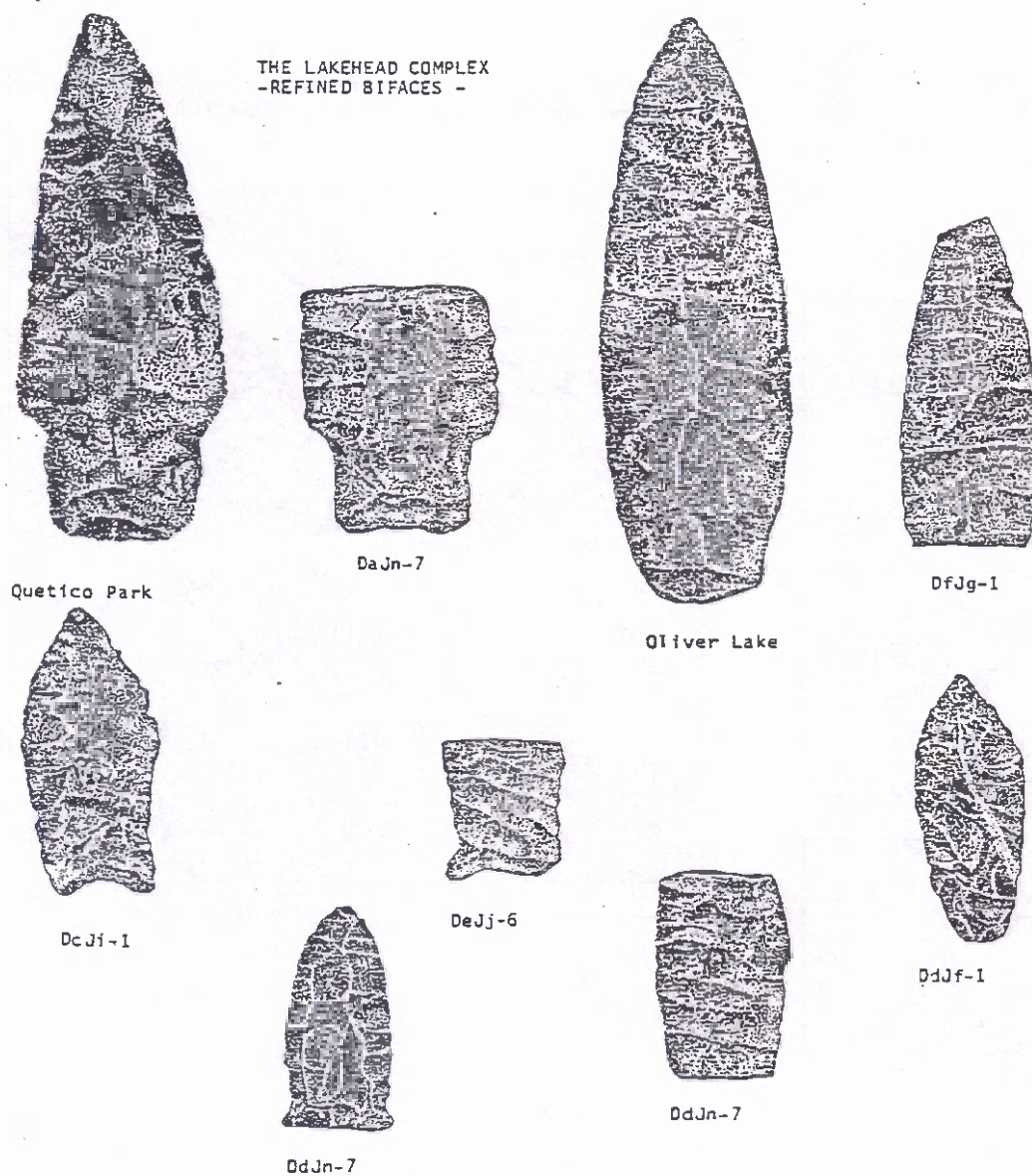
The Simmonds (DcJh-4) and McDaid (DcJh-16) Sites

The present Current river runs in a channel incised into a gently lakeward dipping shelf of Gunflint formation that fronts a bounding rock wall, a structural feature, which now forms the 'bluffs' scenic lookout (Figure 8). In the Minong period, the river carried much water and sediment from inland proglacial lake flows and here evidence supports a major 'bar building' phase of coastal history, a series of river mouth spits or bars being formed on both sides of the river as water level generally declined.

The highest Minong shoreline lay against the bounding rock wall at approximately the 252 m (827') level but between 240 and 236 m (787-774') a series of sand bars formed parallel to the wall on the west bank of the river and a matching series of bars on the east side curve sharply into the then river mouth

Figure 7

Paleoindian bifaces from the Thunder Bay region.



(from Ross, 1995)

from a source on the same rock wall east of the point where it is cut by the river. It appears that the river entered Lake Minong along the rock wall, forming offshore and beach bars which extend south eastwards across the shallow McVickers embayment to the southwest. Simultaneously, longshore transport from the east built bars partially across the river mouth at times, only to be later truncated by fluvial action.

The Simmonds site on the west side, occurring on the parallel bars at about 236 m (774') is matched in elevation and position by the McDaid site on the eastern curving bars (Figure 9). Neither site is a long term habitation site but show evidence of activity typical of a river mouth camping and fishing site. Interestingly, the major bar building episode appears to have been just subsequent to the occupation of these two sites, two large curving bars being formed at 231 and 227 m (758-745') on the east bank, the latter flat topped one largely a subaqueous feature that was probably contemporaneous with the supra-aqueous ridge form of the first. On the west bank a very long bar, now unfortunately truncated at its river mouth end, runs south west, in places broadening to over 100 m (328') in width. In almost text-like manner, the mean grain size and sorting characteristics along its length confirm that it prograded out from the mouth across the McVickers embayment, probably mostly in subaqueous form. Some evidence of Paleoindian activity has been found on the crests of these newer bars but no sites equivalent to those named.

Stop 4 - Mapleward Road, the Cummins Site.

The route will retrace to Balsam and join the Expressway heading southwest. Just beyond the John St. traffic lights the route crosses the McIntyre River. On the north side the low Minong bluff will be seen, with a housing development on top. The Biloski site occupies this surface and a small sand bar at the bluff foot, where a small embayment and secondary river channel once existed.

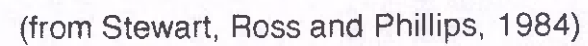
As the Expressway curves to the south, the area on the right holds a lengthy baymouth bar on the end of which lies the Catherine site, unseen from the road. On the summit of Rabbit Mt. (275m, 902'), which rises behind the embayment, lies the Irene site, a lookout site.

The route turns west on Oliver Road and within 2 km rises up the distinct Minong bluff. A left turn on Mapleward Road sees a gentle descent to the Minong shore again, and the Cummins site.

The Cummins Site (DCJi-1)

By far the most significant Paleoindian site is that of the Cummins site (Figure 10). Reported by a local collector in 1962, the site has been excavated and has been the focus of thorough examination. Initially considered a typical surface site, Julig et al. (1986) concluded that it is a rare deeply stratified Paleoindian site, under continued use over a long period of changing environmental conditions. Though the basic tool kit is that of Plains Plano culture (Dawson, 1983; Julig, 1984), the wide range of artefacts imply a diversification into woodworking, fishing and beaver trapping in addition to the regular hunting of caribou and perhaps bison (McAndrews, 1982). The Cummins Site was a major regional tool making

THE M^CDAID SITE — geomorphological basis and
archaeological test and transect locations



centre and the presence of exotic lithic components implies a broad geographical interaction with other groups in the region (Julig, 1984).

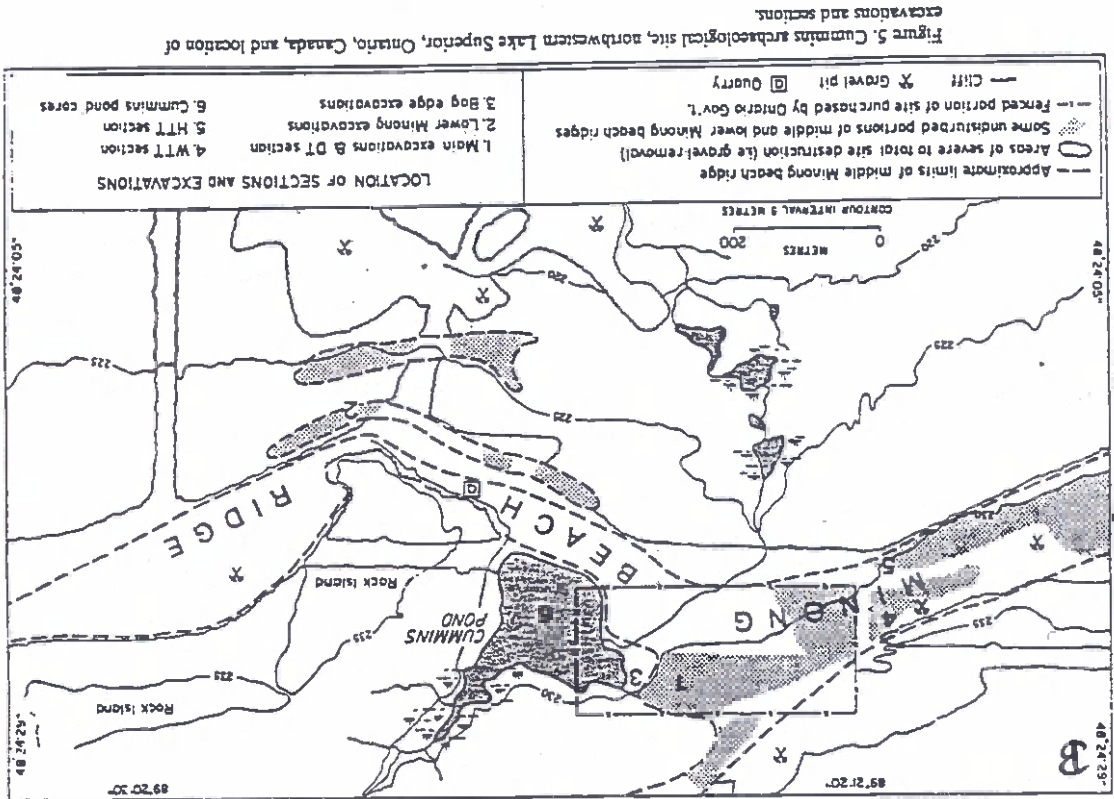
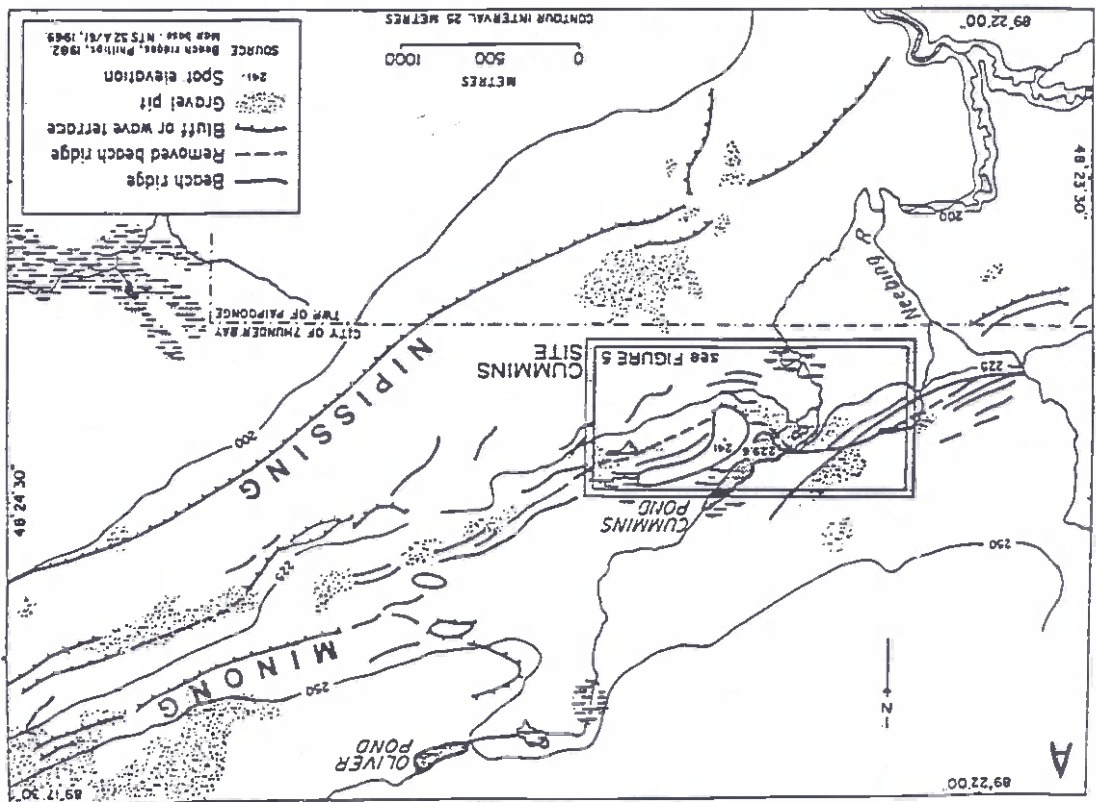
While the archaeological value of the Cummins Site was being revealed, broader questions remained concerning the choice of this and other sites for habitation, especially in relation to lake margin history.

The Cummins Site occurs across the surface of several large sand and gravel bars which trend northeastwards from the then Neebing river mouth across a broad southerly gentle dip slope of local Gunflint shales, overlain by a thin water washed silty till. In these shales occurs jasper taconite, the major source of tool making material. Figure 10 shows the fenced area of the site and the morphological details (Julig et al., 1990). Figure 11 shows the paleogeographic reconstruction of events believed to have formed the area (Phillips, 1982). Longshore transport from the Neebing rivermouth built a series of progressive bars across the shallow water rock shelf, recurving into a minor river valley which formed a sheltered embayment to the west of a rock island. A further bar was built along the front of existing ones eventually crossed the embayment in tombolo-like form to enclose a small lagoon, the Cummins Pond, after which lower and later shoreline features mimicked the established plan shape. A pollen core taken in the pond suggests this enclosure took place before 8.1 ka B.P. (Julig et al., 1986).

The plan shape and structure of the bars is not compatible with a declining water level margin, indeed the accumulation and building up of these large features on a gentle shelf slope is unlikely without transgressive wave action. Even so, such action across a gentle shore slope would not ordinarily build up a large supra-aqueous bar without some initial encouragement to accumulate sediments along a line rather than disperse them in a sheet. The key to the existence of these bars in this location is underlying bedrock control. While Julig (1984) determined through resistivity and ground radar studies that variation in sediment character and bedrock irregularities could be traced, a more simple reconstruction of the bedrock profile is also possible. By surveying down the exposed bedrock dip slope north of the site, the trend (see inset, Figure 11) showed that beneath the site must lie a marked rock step, typical of many that occur in the present topography of these flat-lying shales, and often sharpened by wave action. Accumulation of sediments took place firstly against this rock step and subsequently over the top of it. It is possible that some till remained in the angle of the step. To the west of Mapleward Road an exposure of coarse sand and gravel with angular shale inclusions at the rear of the bar in contact with bedrock suggests overwash of the type that would be expected in this scenario (Figure 12). Once the linear feature was established, longshore sediment supply would extend its length and add to the lakeward face. However, overwash and building in elevation is characteristic of wave action either in an extreme storm event or a transgressive lake margin. Again, a 'bar building' period seems evident at Cummins. Julig (1984) found confirmatory evidence of habitation of these bars during the process of their accumulation, a buried layer of water worn taconite artefacts being associated with a period of overwash and construction.

While occupation of the site when a lakeshore location offered advantages of longshore access, lagoon fishing and perhaps water transport, a cremation

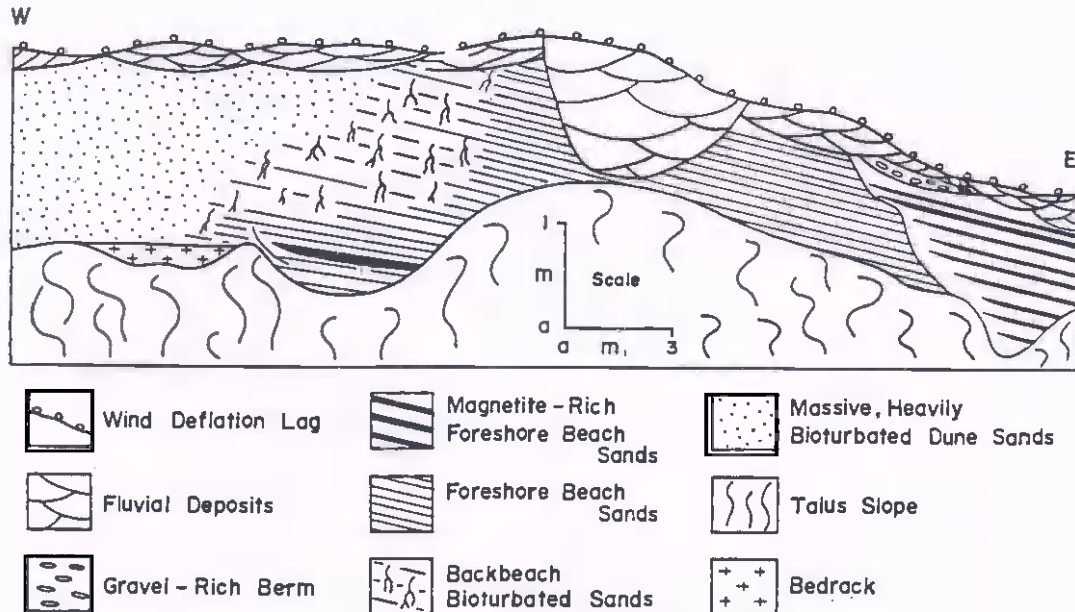
Figure 10



Jullig, McAndrews, and Mahaney (1990), based on morphological mapping by Phillips (1962)

Figure 12

Schematic representation of the cliff face at the Cummins site.



Sedimentology

A well-developed, beach deposit is exposed at the Cummins site. Figure 30 shows a schematic representation of the sandy cliff face where the strandline deposits are visible. Low-angle, planar cross-stratified sands of the foreshore dominate the exposure. Individual laminae dipping 3° to 12° lakeward (original swash-backwash surfaces) are arranged into packets which erosively truncate one another at very low angles. The planar cross-stratified sands are transitional both laterally and vertically into massive sands through a bioturbated zone. the bioturbated area represents a sparsely vegetated backshore environment while the massive sands were deposited as aeolian dunes.

The foreshore sands are erosively truncated in the eastern portion of the cliff by magnetite-rich sands. Internal structures indicate that they are also foreshore deposits. The magnetite-rich foreshore laminae were formed during regression when storm wave activity reworked the lower portion of the beach, erosively truncating the older deposits. During the storm events sand was removed and stored in offshore bars. In intervening periods of fair weather, sand moved from the offshore bars back onto the beach and was winnowing by small waves, producing the magnetite-rich lag deposits filling scour truncations.

The beach assemblage is overlain by erosively based, trough cross stratified sands. These were formed after subaerial exposure of the area. Major rainfall events caused streams to flow off the adjacent rock knoll dissecting and reworking the upper layer of beach deposits.

Phillips, Fralick and Ross, 1987.
From INQUA Field Guide C-12,
Eds. Geddes, Kristjansson and Teller.

burial site on the rock island (Dawson, 1983) was dated to 8,480 ± 390 years BP. Julig (1984) found artefacts both just below the aeolian sands that characteristically modified the topography of the shoreline features once the offshore shelf was exposed by lower lake levels (circa 8.0 ka BP) and in peat that accumulated in Cummins Pond. Indications are that long after the lake ceased to lap the beach face, the site remained used, at least until 7.5 ka BP (Julig et al., 1986).

Stop 5 (several) - The Rosslyn Delta

The route will retrace to Oliver Road and continue west, turning south at the Golf Course towards Twin City Crossroads. Immediately south of the driving range, the road descends a gentle bluff around the 268m (880') contour that is a Beaver Bay shoreline, and then, beyond two turns in the road, the descent down the Minong shoreline arrives on the extensive, flat surface of the offshore shelf and delta of the Kaministiquia River where it entered Lake Minong. This surface, 222m (729') at the crossroad, is composed mainly of sandy silts, with boggy areas underlain by more clayey silts. From the crossroads the route will turn west along Hwy 11-17, where, within 2 kms, is again the Minong bluff, forming an abrupt rise at the Weighscales. Turning left, once on the terrace, the road leads south to several fields in which lithic material has been found. These sites, the Dairy Farm, Breukelman Evergreen and Halow A, B and C, will be viewed (Figure 15) and then the route will follow the north side of the Kaministiquia River towards Stanley. En route the upper and lower Drezecky sites will be seen (Figure 17), as well as the Pawlick site which can be seen on the south side of the river. From Stanley the route will climb the terraces of the Stanley delta and return to Hwy 11-17.

The following extract of text and figures derives from a paper by Scott Hamilton, that has been submitted for publication review to the Canadian Journal of Archaeology. Entitled "Archaeological Predictive Modelling in the Boreal Forest: No Easy Answers", the paper addresses the difficult problem of developing archaeological predictive models to aid in Forest Class environmental assessment. Among the many challenges is the issue of development of temporally sensitive palaeo-environmental reconstructions, and also models of land use that are relevant to the political economy and social organization of the ancient peoples under consideration.

"The apparent correlation between relict shorelines and Plano archaeological sites is well known in the Thunder Bay area where landscape evolution has been the subject of some study (Figure 13). While many Plano shoreline archaeological sites are documented, the best known ones are lithic quarry and workshop sites such as the Cummins, Biloski, Simmonds, and McDaid Sites (Julig 1984, 1994; MacNeish 1952; Hinchelwood 1990; Phillips 1988, 1993). These large sites are well removed from modern shorelines, but are associated with the late Pleistocene shorelines of Glacial Lake Minong and Gunflint Formation bedrock exposures that are suitable for lithic tool production (Figure 13) (see Julig, McAndrews and Mahaney 1990; Phillips 1988, 1993). However, these special-purpose sites do not characterize the full Paleo-Indian settlement pattern. Rather, the dense recoveries

from, and ready visibility of, these sites has resulted in their over-representation in the published archaeological literature. Archaeological reconnaissance upon agricultural fields within the Kaministiquia River delta has revealed a number of Plano or probable Plano sites in a wide range of landscape contexts (Hamilton 1996) (Figure 14). Many of these sites are found at or near Lake Minong shoreline elevations (i.e. 750 feet or 228.6 metres A.S.L.). When placed in a hypothetical Lake Minong environmental context, they are found associated with:

- 1) springs flowing into sheltered coves of Lake Minong (Figure 15: Breukelman Evergreen);
- 2) on points of land that protruded out into Lake Minong to the northeast and southeast of the Breukelman Evergreen site cluster (Figure 15);
- 3) upon low, well-drained sandy knolls surrounded by poorly drained floodplain/deltaic sediments (Figure 15: Halow C);
- 4) along deltaic backwater channels (Figure 16: DbJi-8, DbJi-7, DcJi-28, DcJi-32); or
- 5) on sandy storm beaches developed upon relict deltas (Figure 16: DcJi-30, DcJi-31).

Other sites were found upon:

- 6) raised Pleistocene terraces overlooking the present Kaministiquia River channel (Figure 17: Pawlick);
- 7) on high valley rims that offer panoramic views (Figure 17: Drezecky E);
- 8) along draws leading down to the Lake Minong shores (Figure 16: Halow A and B);
- 9) along upland streams well removed from late Pleistocene shorelines (Figure 17: DcJj-12, DcJj-13); and
- 10) upon well-drained upland knolls on what likely were formerly discontinuous permafrost uplands (Figure 18: Breukelman Field: lithic scatters A to E).

Probable Plano lithic scatters have also been found upon isolated bedrock controlled knolls offering panoramic views of floodplains adjacent to the Lake Minong shoreline (Figure 19: Belluz Farm). This range of landscape characteristics is certainly not exhaustive, but is sufficient to demonstrate the diverse microhabitats frequented by Plano people.

These sites are consistent with other Lakehead Complex sites in terms of a very strong preference for Gunflint Formation lithic materials (Hamilton 1996). However, virtually all of these sites are well removed from bedrock exposures, are

Figure 13

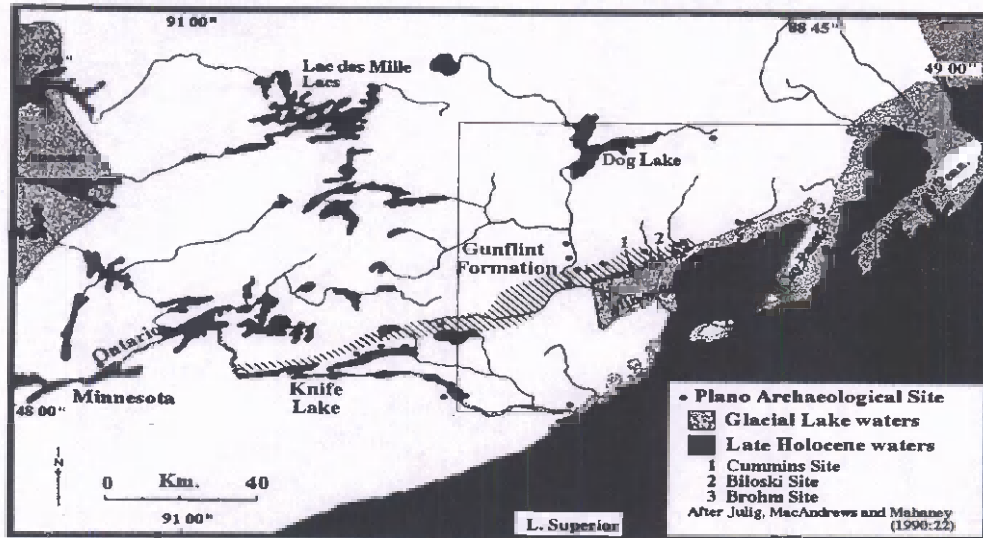


Figure 6 The Correlation between Lakehead Complex sites, Minong beaches and Gunflint Formation outcrops (Hamilton 1996:34)

Figure 14

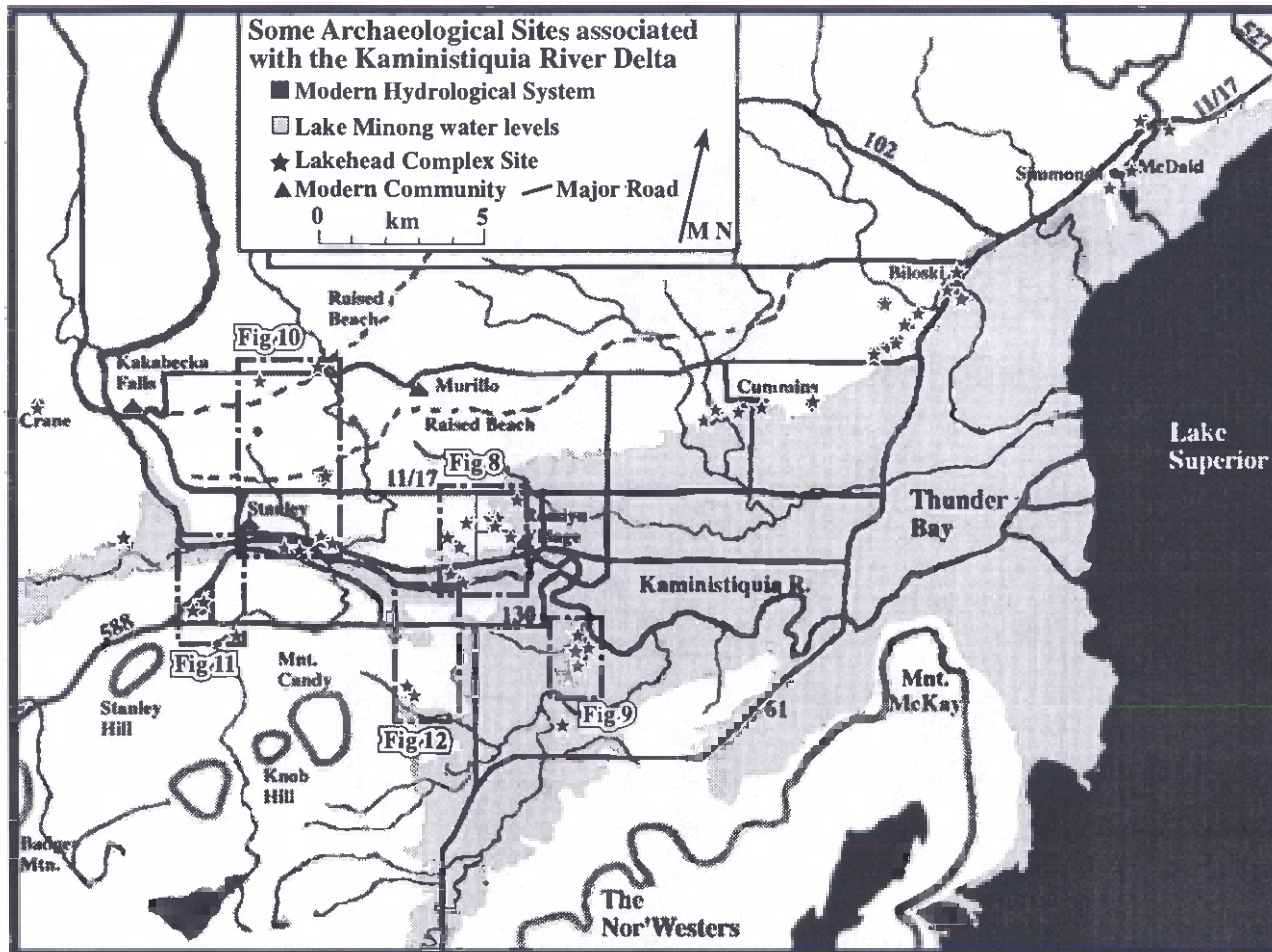


Figure 7 The distribution of confirmed or probable Lakehead Complex sites on the Kaministiquia River delta (Hamilton 1996: 36)

Figure 15

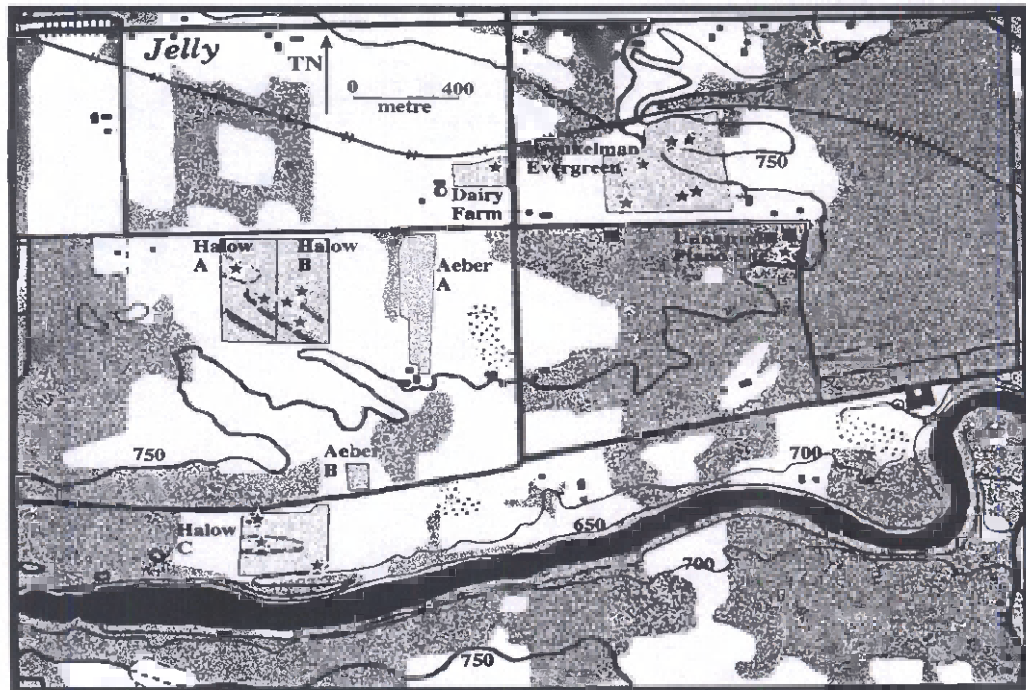


Figure 8 Confirmed or probable Plano sites on well drained sediment near Lake Minong shorelines (750 feet ASL) (Hamilton 1996:59)

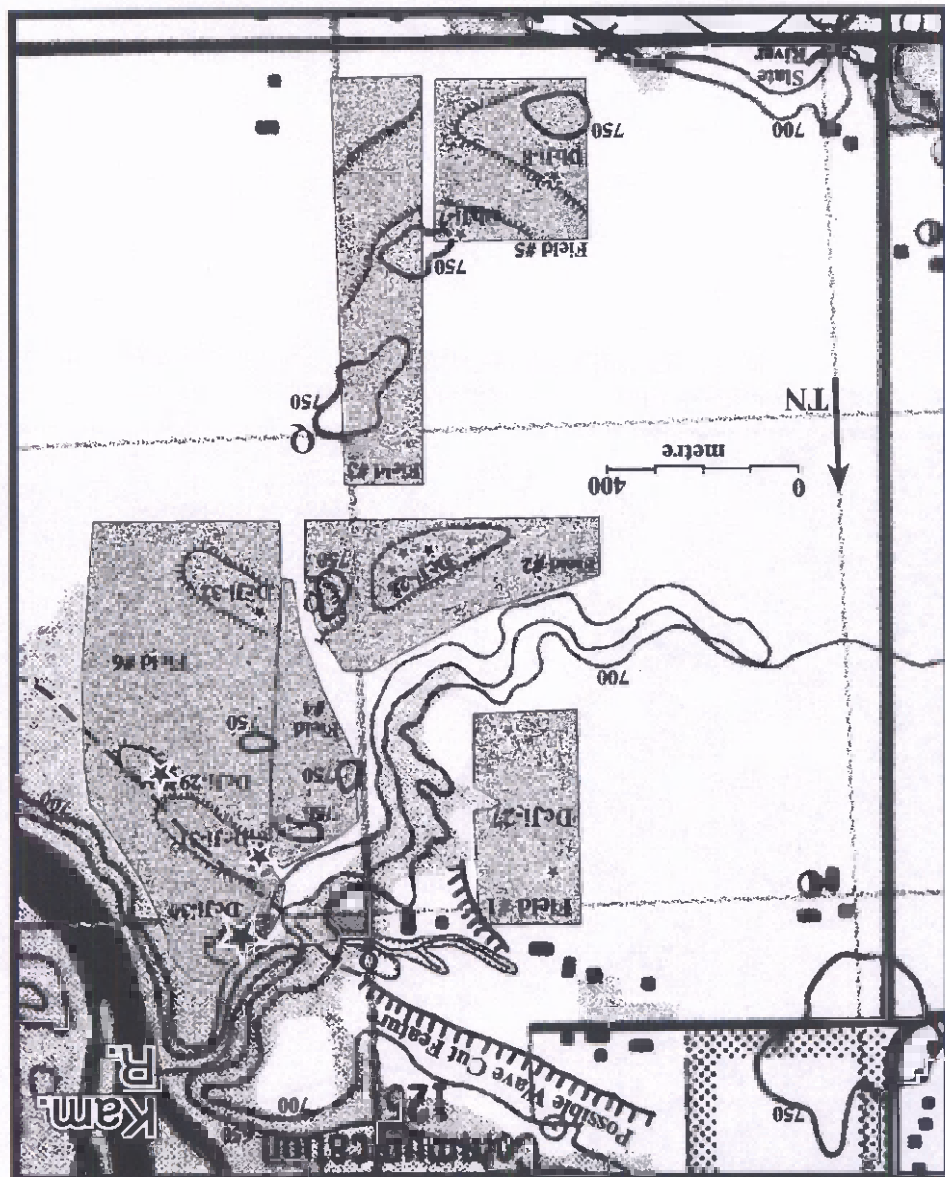


Figure 16

Figure 9 Confirmed and probable Plano Sites on Minong storm beaches, and along deltaic backwater channels (Hamilton 1996:65).

Figure 10 Confirmed and possible Plano sites along late Pleistocene and early Holocene river terraces (Hamilton 1996:51).

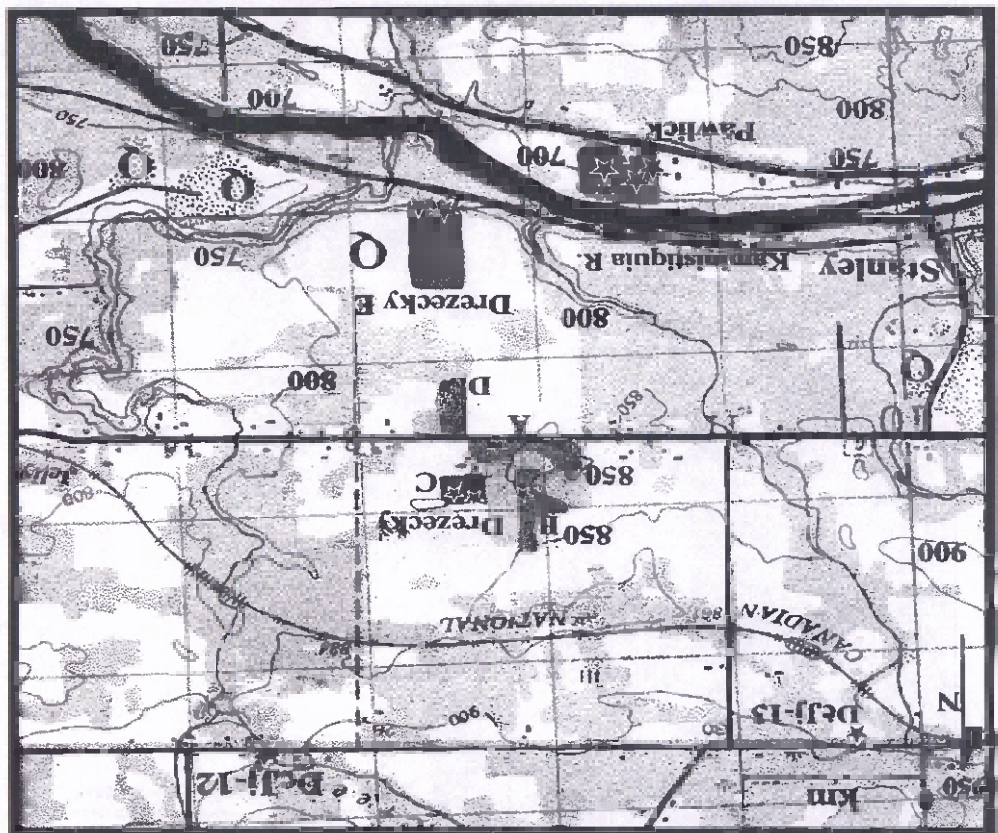


Figure 17

Figure 18

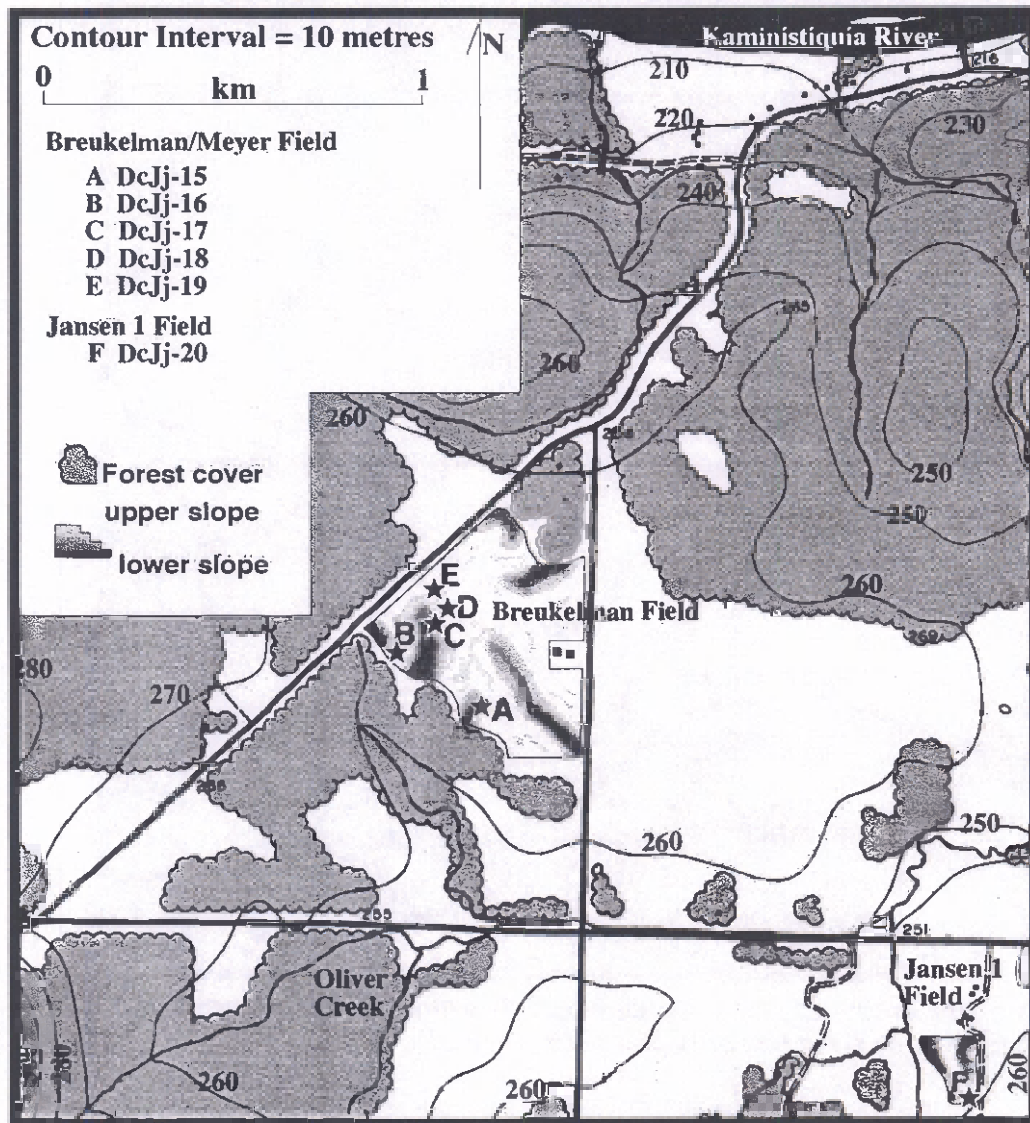


Figure 11 Possible Plano sites located upon well-drained knolls in upland regions well removed from Lake Minong (Hamilton 1996:47).

Figure 19

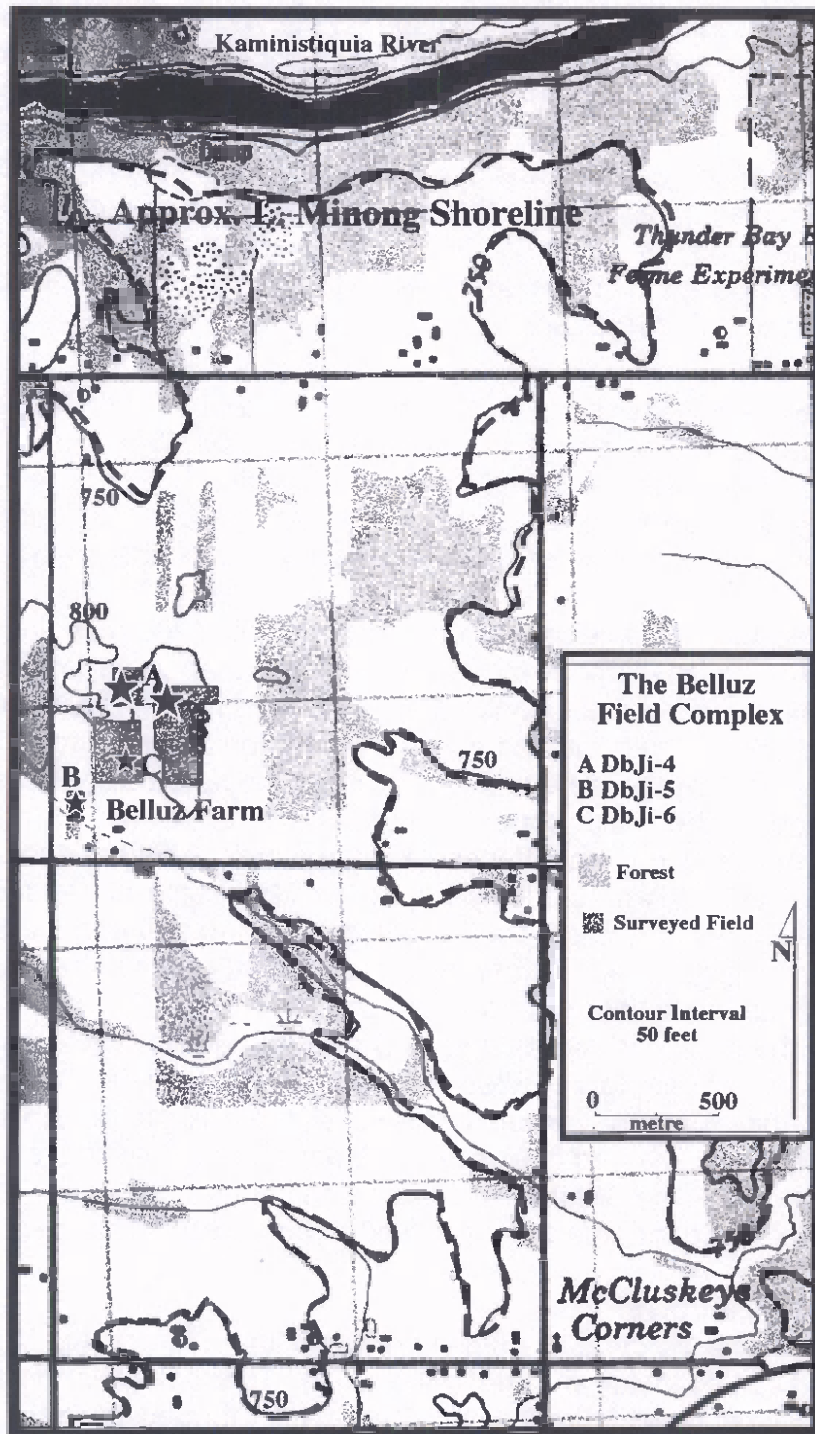


Figure 12 Probable Plano sites on a bedrock controlled knoll overlooking Pleistocene floodplain zone (Hamilton 1996:57).

significantly smaller than the quarry-workshop sites (such as the Cummins Site), and yield many fewer artifacts. When compared to the quarry/workshop sites, these small sites also yield a much higher relative frequency of tools, preforms and utilized flakes compared to debitage (Hamilton 1996). This indicates that the small sites represent encampments, hunting stands and food procurement sites, rather than lithic extraction and reduction stations. Such observations are hardly surprising, but they do serve as cautionary tales regarding the dangers of predicting site distribution on the basis of our current incomplete heritage inventory. These examples are also important in that they demonstrate that modelling ancient human behaviour requires ongoing refinement, an understanding of human forager behaviour, and a well-developed sense of the nature and structure of the ancient landscape and its microhabitats."

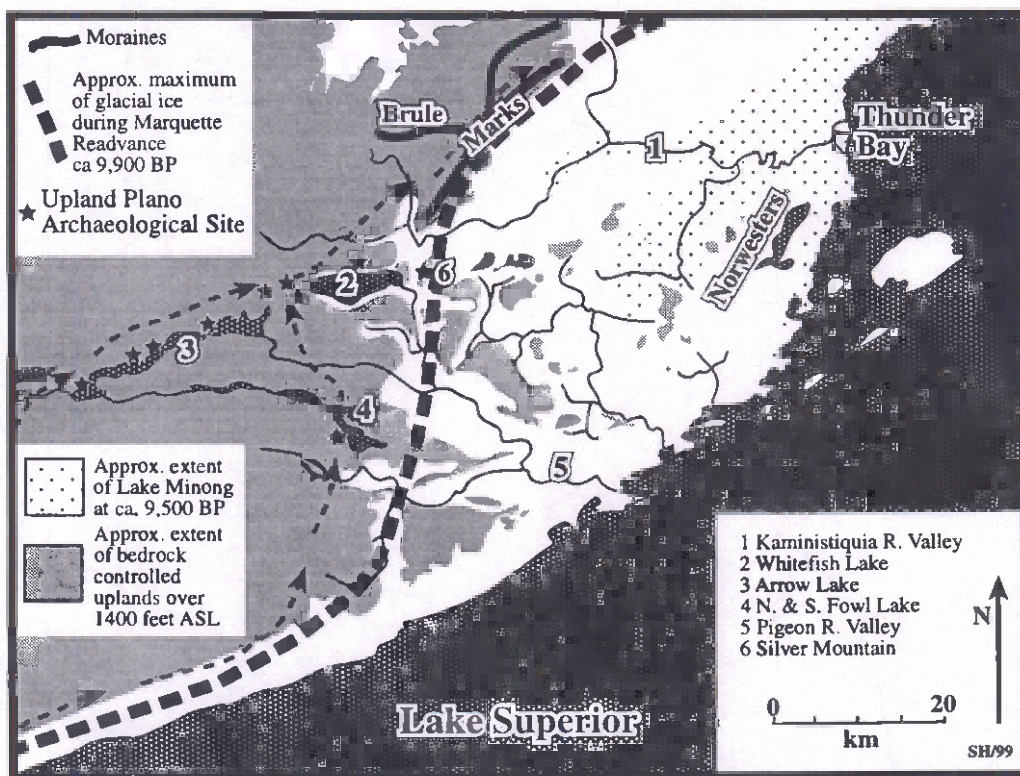
As one leaves the Rosslyn delta (Minong, 9.5 ka B.P.) and moves up the present Kaministiquia valley, one is in effect travelling back in time. The Stanley delta was built into Lake Beaver Bay (9.7 ka B.P.), the first Superior lake to occupy the area recently vacated by Marquette ice retreating from the Marks moraine. Though perhaps less conspicuous, Paleoindian sites of this period are also present.

The larger step back in paleogeography is to consider the area that lies to the west of the maximum position of the Marks moraine, an area which remained unglaciated by the Marquette readvance and which was freed from ice around 11.0 ka B.P. Figure 20 shows the projected maximum position of the Marquette lobe and the large area covered by the waters of a pre-Marquette lake (Early Lake Minong), now approximated by the 1400' contour. On the basis of some long known sites, both in the interior and on the Minnesota northshore, Phillips and Hill (1995) proposed that a Paleoindian routeway along the Minnesota shore turned inland just east of Judge C. Magney State Park, towards North and South Fowl lakes and the Whitefish Lake region. A number of more recently discovered sites in the Whitefish - Arrow Lake corridor supports the contention that Paleoindian peoples were present in the area perhaps before and during the Marquette advance, and that the corridor continued to be used into the post-Marquette periods of Lake Beaver Bay and Lake Minong. Though conjectural, there is the possibility that after the retreat of Marquette ice had begun, the Marks moraine itself provided a high ground routeway from the Whitefish Lake area along the north side of the Kaministiquia valley and into the area to the north and east of Thunder Bay.

Stop 6 - Kakabeka Falls

Just west of the junction of Hwy 11-17 and the Stanley turn off, the huge structure of the Stanley delta is seen (Figure 21). This delta was formed as the Kaministiquia River entered Lake Beaver Bay. A well formed bluff representing a lower Beaver Bay phase of 260m (853'), runs along the north side of the main highway, and the surface below it (250-240m; 820-787') is heavily exploited by the sand and gravel industry. The present Kaministiquia river has incised deeply into the delta, creating a terraced valley side.

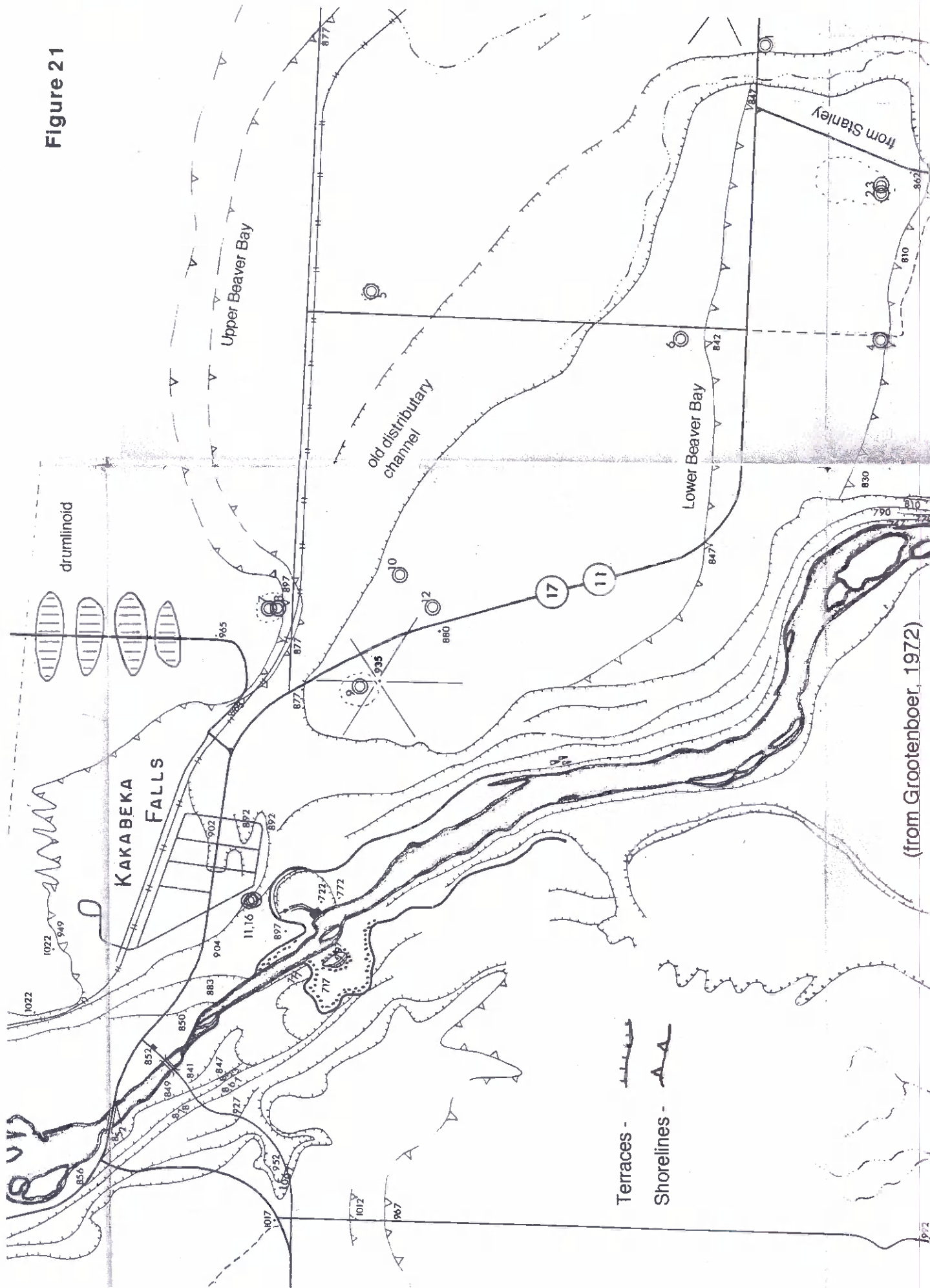
Figure 20



(After Phillips and Hill 1995:22-24, Hamilton 1996)

Figure 13 Some Plano Archaeological Sites on the Uplands west of Thunder Bay

Figure 21



Kakabeka village is built on the floor (277m; 909') of an old distributary of the Kaministiquia river which cut through a higher terrace still (Figure 21). Remnants of this terrace surface at around 300m (984') overlook the village. They represent the highest level of the Stanley delta, as it formed into Upper Lake Beaver Bay, and on the one on the west side of the Kaministiquia River, the Crane site was found.

Kakabeka Falls is a major scenic attraction. It is formed as a result of a very resistant chert bed that caps the underlying, softer shales. It was just this sort of rock material that was desired for tool making. Today, a fairly spectacular gorge lies below the falls (Figure 22). There has been some debate concerning the age of the gorge, since unless inherited and exhumed from a previous stage of river incision prior to the Marquette readvance, there is only post Beaver Bay time for it to develop its current grandeur. An interesting abandoned falls and plunge pool on the west side of the gorge lies a short walk from the Park Information Centre, and represents a wide 'horseshoe' fall that would have been more spectacular than the present falls.

Stop 7 (several) - The Marks Moraine.

The route will turn on Hwy 590 just west of Kakabeka, run across the High Beaver Bay terrace remnant (look for the cemetery on the right) and due west until turning north towards the Marks Moraine. As the dirt road rises up the outer face of the Marks moraine it crosses two distinct benches at 426m (1400') and 442m (1450'). These shorelines of pro-glacial Lake Cedar Creek have been traced all along the outer slope of the Marks moraine (Jahnke, 1993) and represent stillstands in a lake earlier than and isolated from the lakes of the Superior basin.

The Marks Moraine.

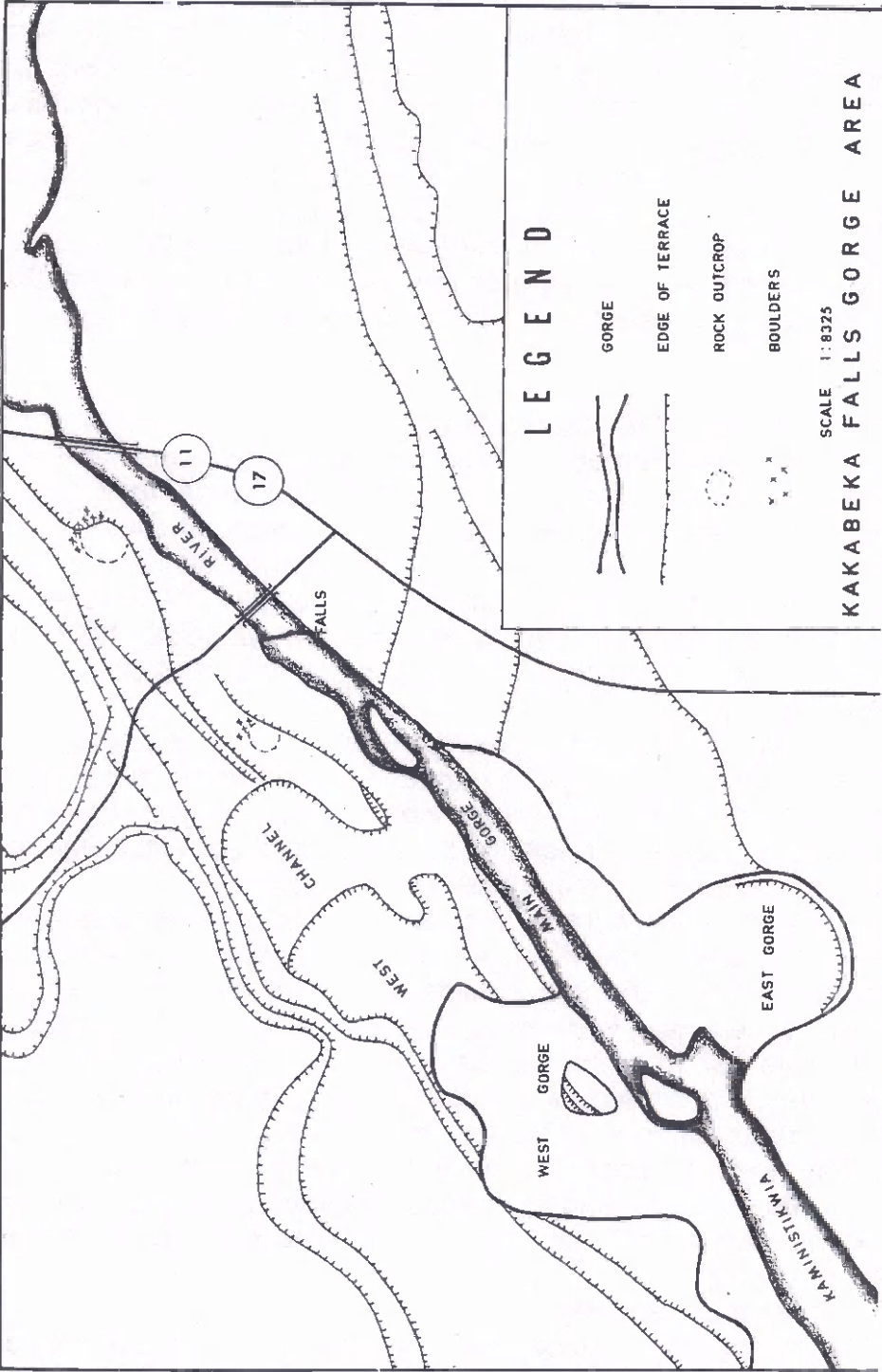
Stopping on the crest around 460m (1510'), the view south over the Whitefish valley and the rugged borderland country of the mesa-like NorWesters is spectacular. It is not hard to envisage ice pushing its way up to the Marks moraine, nor is it hard to imagine pro-glacial Lake Cedar Creek occupying the narrow strip between the moraine and the retreating ice margin.

The moraine is not a simple structure. Although characterised by a till which contains diagnostic pieces of Sibley red sedimentary rocks, along much its length the Marks moraine joins and smothers isolated rock outcrops which probably greatly influenced the extent to which the ice pushed inland. The top of the moraine is remarkable flat in many places and is pockmarked with kettles. Spillway channels cross the feature in places suggesting that at one point Lake Kaministiquia to the north was a few metres higher than Lake Cedar Creek to the south. The moraine appears to have briefly existed as a narrow ridge between these two water bodies, and a possible post-Marquette routeway for animals and Paleo-Indian people, from the unglaciated (Marquette) Arrow/Whitefish region and Interlakes corridor.

The Conmee Pit and spillway.

Continuing north across the moraine and then turning east brings the excursion to a distinct channel cutting from north to south across the moraine surface. Here Conmee Township has several gravel pits. Apart from the huge boulder beds that suggest high discharges down this spillway at times, the pit

Figure 22



(from Grootenboer, 1972)

reveals another key fact in understanding local deglaciation. Figure 23 shows a rare occasion when the western pit face was cleanly exposed. Tickle (1996) interpreted the sequence as consisting of two tills separated by fluvioglacial sediments. However, while one till was of diagnostic Sibley origin, the underlying one was of a northern provenance typical of the Patricia or Rainy River lobes.

Elsewhere, at several places along the Marks moraine a thin Sibley till is plastered over fluvioglacial sediments. The observation suggests that the bulk of the Marks moraine may in fact be pre-Marquette in origin, with thin Marquette ice reoccupying the moraine. This has been the growing opinion of another local field worker (T. Noble, pers.communication, 1999) and looking at Figure 3, the possibility of the Marks moraine being the result of pushing the eastward part of the former Brule moraine is not unreasonable. Several exposures of contorted till structures also support this idea.

Marks Moraine scenic lookout.

Turning left out of the pit and again at the crossroad brings one to a point in the road from which a view east over the Kaministiquia valley, Thunder Bay city and the distant Sleeping Giant is obtained. The stop is useful only to impress viewers with scale of things. It is easy to imagine ice grinding past Mt. McKay and flowing up the valley, just as it is easy to imagine that a slight tilt of the present lake would bring the margin of the lake right up into the Kaministiquia and Whitefish valleys.

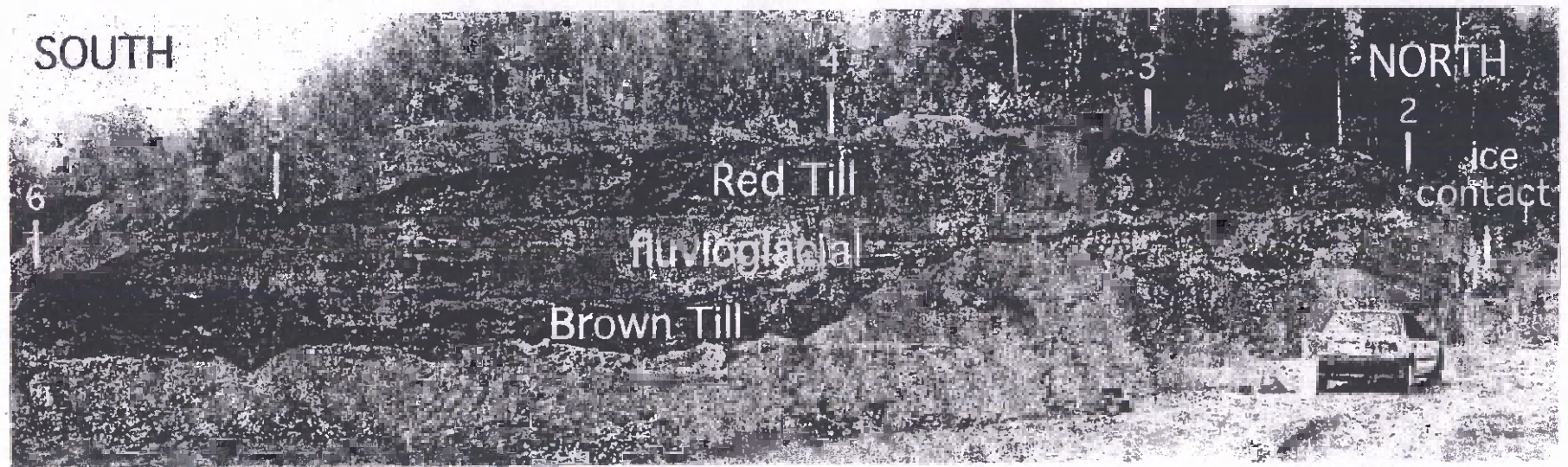
Returning to the crossroad, the route will turn left down the face of the Marks moraine. The road is straight, save for a jog around a kettle hole, and, lower down, rock is exposed in the fields and along the stream beds. Turning right at Hwy 11-17 brings the excursion back to Kakabeka Falls. From here the route will take Oliver Road back to the University. The road climbs out of Kakabeka over the Upper Beaver Bay terrace remnant and then runs due east to Lakehead University.

One landform of note on the way is a crag and tail feature, showing the flow of ice up the Kaministiquia valley. The 'Old Barn' restaurant lies on the tail and on the crag is the farmhouse, open for bed and breakfast (another time, eh!). Just east of Murillo but to the north of Oliver Road is the Murillo drumlin field, a series of long narrow forms parallel to the strike of the Gunflint shales, suggesting a good deal of bedrock control.

The excursion is over. We hope that participants obtained an informed glimpse of the deglaciation history and Paleoindian presence in the area, and enjoyed doing so.

Figure 23

Figure 1.1 The Study Site - Conmee Township gravel pit



Note: Vertical sections taken at numbered locations.

(from Tickle, 1996)

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GEOLOGY OF THE PALEOPROTEROZOIC GUNFLINT FORMATION

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The Gunflint Formation records a chemically dominated shelf sequence in which silica- and iron-rich precipitates predominate. This trip will examine the role that both physical and chemical processes played in various shelf environments.

FIELDTRIP 4

DEPOSITIONAL ENVIRONMENTS OF THE PALEOPROTEROZOIC GUNFLINT FORMATION

Pier Pufahl and Philip Fralick

INTRODUCTION

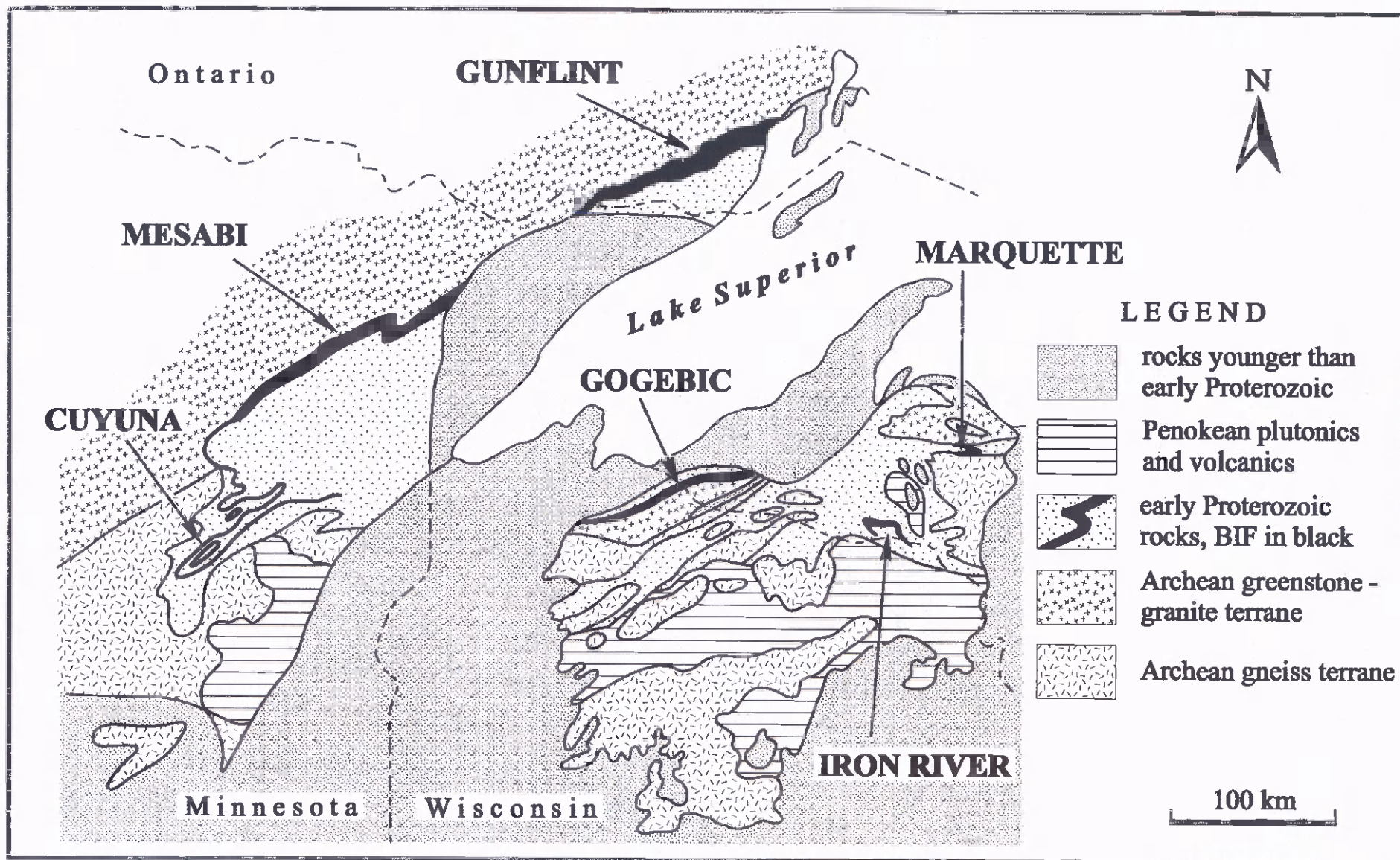
Paleoproterozoic iron formations host most of the world's iron ore and many other important mineral deposits (Simonson, 1985). Consequently they have been the focus of hundreds of mineralogical, geochemical and sedimentological investigations. However, in spite of this immense amount of data no consensus has ever been reached as to their origins. Some of this controversy stems from a poor understanding of the sedimentological environments in which they accumulated (Simonson, 1985). They were deposited during a unique interval of earth history when grossly different atmospheric and hydrologic conditions prevailed (Cloud, 1973), and therefore lack precise modern analogues (Simonson, 1985; Fralick and Barrett, 1995). Modern subaqueous hot-spring activity has provided clues to understanding the precipitation of iron-rich sediments, but the mineralogy and areal extent of modern deposits is not comparable to the large Paleoproterozoic iron formations deposited during the Earth's early history (Fralick and Barrett, 1995). Detailed petrographic investigations of iron formations have contributed to interpreting the paragenesis of syndepositional, diagenetic and metamorphic mineral assemblages, but they have done little to provide detailed, reliable information regarding iron formation depositional processes. Minerals and textures are commonly altered to secondary assemblages, and when primary minerals are present theoretical chemical modelling has been limited in its ability to explain their genesis (Fralick and Barrett, 1995). Paleoenvironmental inferences from grain-size distributions in iron formations are also plagued with uncertainty because they are commonly indurated, diagenetically altered, and frequently fine-grained (Simonson, 1985).

Fortunately, these problems can be overcome by combining a regional basin analysis of clastic and volcanic rocks surrounding iron formations with detailed examination of vertical and lateral facies trends within chemical sedimentary successions (Fralick and Barrett, 1995). Associated sedimentary and volcanic rocks provide data on depositional settings. Sedimentary structures and lithic associations within iron formations give information on physical sedimentary processes, and regional basin analysis provides a framework within which to view iron formation genesis. Combining these techniques makes paleogeographic reconstruction and depositional modelling possible (Fralick and Barrett, 1995). This is the same approach used by sedimentologists and stratigraphers to reconstruct clastic depositional systems.

Location

The main portion of the Gunflint iron range extends 170 km northeast from the United States, at Gunflint Lake, to Loon Lake, Ontario (Figure 1). In Minnesota the Gunflint forms a narrow belt

Figure 1. Distribution of Paleoproterozoic iron formations in the Lake Superior region.



20 km long which is truncated to the southwest by the Duluth Complex. Erosional remnants 98 km northeast of Thunder Bay at Schreiber, Ontario, suggest a pre-erosional continuation of the chemical sedimentary rocks to the east.

Previous Work

The Gunflint iron formation was discovered in 1850. The earliest recorded geological investigation of the Gunflint was conducted by E. D. Ingall in 1887 (Ingall, 1888), who briefly described the iron-bearing strata near Silver Mountain and Whitefish Lake. Other early accounts were made by Smith (1905) and Silver (1906). Van Hise and Leith in 1911 presented a general overview of the iron bearing rocks in the Thunder Bay district. In 1924 J. E. Gill was the first to describe the Gunflint in detail, and in 1926, its stratigraphy northeast of Silver Mountain. In 1927 Gill published again, this time on the origin of the chemical sediments. T. L. Tanton described the iron prospects at Mink Mountain in 1923, and in 1931 gave an overview of the general geology in the vicinity of Thunder Bay.

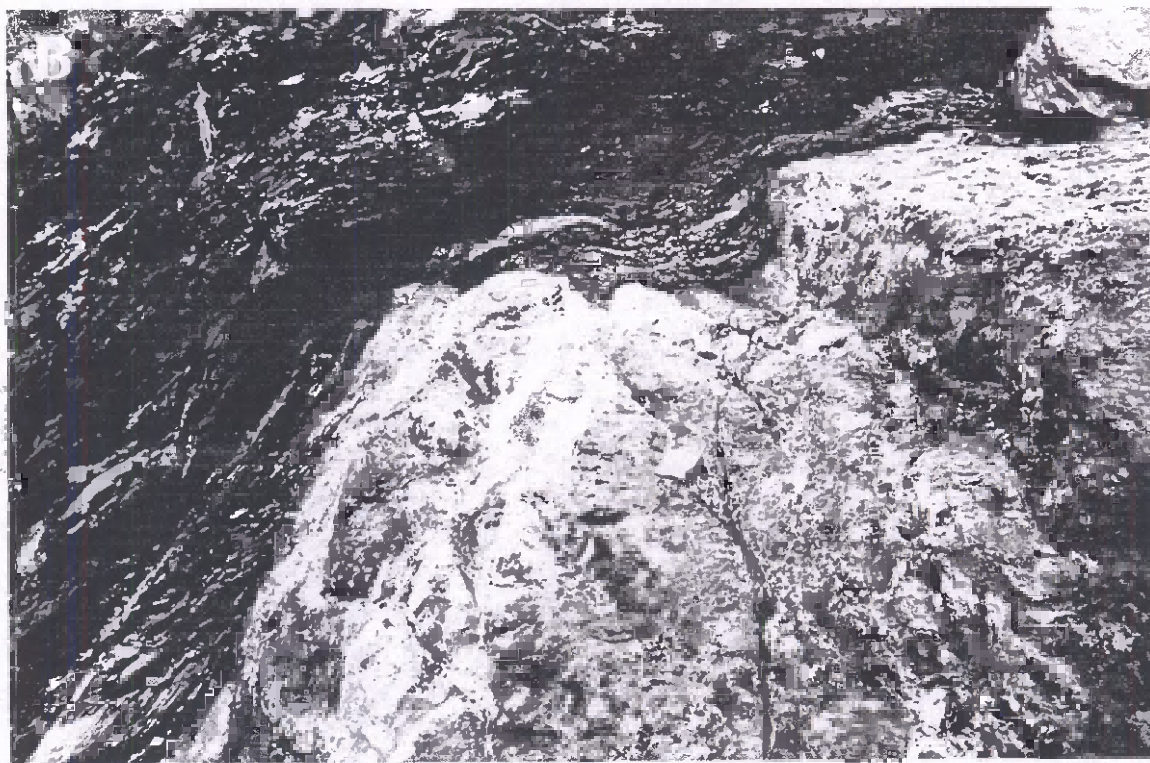
In 1956 the iron formation was divided by Goodwin (1956) into six major sedimentary facies representing two depositional cycles and four members, a 1) Basal Conglomerate Member, 2) Lower Gunflint Member, 3) Upper Gunflint Member and 4) Upper Limestone Member; which are largely correlative with Broderick's (1920) four-fold division of the Gunflint iron range in northeastern Minnesota and Wolff's (1917) subdivision of the Biwabik iron range in east-central Minnesota (Figure 2). He also drew attention to evidence of widespread volcanism associated with deposition of the chemical sediments and suggested a volcanic source for the iron and silica. In 1960 Goodwin and Moorehouse published the first detailed geological maps of the region (scale 1:31 680) and reported on the economic potential, mineralogy and chemistry of the iron bearing strata, and modified Gunflint stratigraphy into a Lower and Upper Gunflint depositional cycle composed of 10 members; a Basal Conglomerate, Lower Algal Chert, Lower Shale, Lower Taconite, Local Lava Flows, Upper Algal Chert, Upper Jasper, Upper Shale, Upper Taconite and an Upper Limestone member. Fralick and Barrett (1995) have recently redefined Gunflint stratigraphy as simply consisting of 2 members; a lower member comprised of basal algal and oncolitic cherts, and cherty grainstones, and an upper member similar in stratigraphy to the lower member. They have interpreted this succession as representing a wave- and tide-dominated inner shelf sequence.

Floran and Papike (1975), Randazzo and Markun (1980), Loughheed (1983) and Simonson (1987) have given detailed petrographic descriptions of mineral assemblages comprising the Gunflint iron formation. The Gunflint volcanics (Goodwin, 1956, 1960) have been petrographically described by Hassler and Simonson (1989). Their REE geochemistry has been determined by Kissin and Fralick (1994). Carbon and sulfur isotopic studies have indicated a hydrothermal exhalative origin as the source for the iron (Cameron, 1983; Carrigan and Cameron, 1991). Other geochemical investigations (Kronberg and Fralick, 1992) have attributed the alteration of the Archean basement to the penetration of Gunflint-derived fluids under diagenetic or low-grade metamorphic conditions. The overlying Rove Formation was described by Morey (1967, 1969). Detailed accounts of microfossils preserved within Gunflint chert are recorded by Tyler and Barghoorn (1954), Barghoorn and Tyler (1965a,b), Edhorn (1973) and Awramik (1976).

Figure 2. Proposed correlations of Animikie Group sediments in the Lake Superior region.

Northwest segment				Southeast segment					
Gunflint range, Minn. - Ont.	Mesabi range, Minnesota	Cuyuna range, Minnesota	Western part Gogebic range, Michigan	Eastern part Gogebic range, Michigan	Western Marquette range, Michigan	Eastern Marquette range, Michigan	Amasa uplift, Michigan	Iron River - Crystal Falls range, Mich.	Menominee range, Wisconsin
Goodwin, 1958	White, 1954	Morey, 1978	Aldrich, 1929	Sims and others, 1984	Cannon, 1986	Gair and Thaden, 1988	Cannon, 1986	Cannon, 1986	Bayley and others, 1966
Animikite Group	Rove Formation	Virginia Formation	Rabbit Lake Formation	Tyler Formation	Michigan Formation	Michigan Formation	Michigan Formation	Dunn Creek Slate	Badwater Greenstone
	Gunflint Iron - Formation	Biwabik Iron - Formation	Trommaid Formation	Ironwood Iron - Formation	Copps Formation	Goodrich Quartzite	Goodrich Quartzite	Michigan Slate	Michigan Slate
	Basal con - glomerate member			Emperor Volcanic Complex Blair Creek Formation		Negaunee Iron - Formation	Negaunee Iron - Formation	Amasa and Fence River Formations	Vulcan Iron - Formation
					Palms Formation	Siama Slate	Siama Slate	Hemlock Formation	Fetch Formation
						Ajibik Quartzite	Ajibik Quartzite	Hemlock Formation	

Figure 3. Photographs of Gunflint sediments outcropping at the spillway section near Kakabeka Falls. A) Trough cross-stratified "lapilli" tuff (reworked volcanoclastic rip-ups). B) Large algal bioherm draped by shales at the base of sequence.



Several isotopic investigations have constrained the depositional age of the Gunflint Formation between 1.6 and 2.0 Ga. Hurley *et al.* (1962) obtained a minimum age of deposition of about 1.6 Ga by both Rb-Sr and K-Ar methods, to which they applied an empirical correction factor, which allowed them to propose an age of approximately 1.9 Ga (Stille and Clauer, 1986). Faure and Kovach (1969) also obtained a Rb-Sr isochron age of approximately 1.6 Ga. More recently, Stille and Clauer (1986) obtained a Sm-Nd isochron age of about 2.08 Ga for volcanic slates interbedded with iron-bearing strata of the Gunflint Formation.

The absolute age of the Gunflint Formation was determined by Fralick *et al.* (1998) by conducting U-Pb, single zircon geochronology on a sample population from reworked tuffs outcropping near Kakabeka Falls (Figure 3), a euhedral zircon gave a concordant age of 1878 ± 2 (Figure 4) which Fralick *et al.* (1998) argue is the depositional age of this unit.

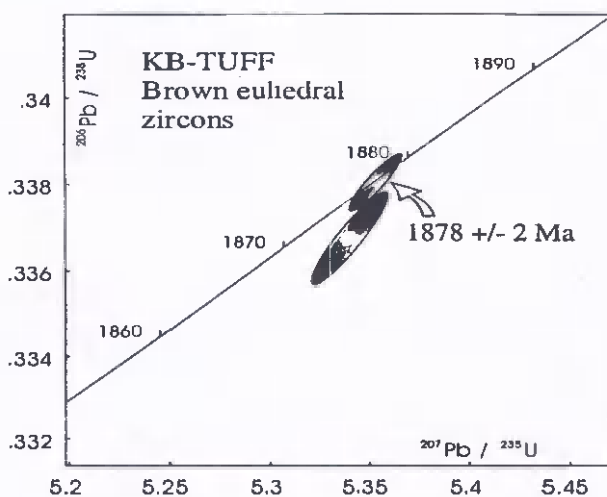


Figure 4. Concordia diagram for zircons from the upper Gunflint Formation (Fralick, *et al.*, 1998)

Animikie Basin

The Paleoproterozoic sedimentary rocks deposited in the Animikie Basin form a southward-thickening wedge covering the southern margin of the Superior province, which is truncated in east-central Minnesota and northern Wisconsin by the "Penokean" magmatic terranes". Sedimentation began approximately 2.1 Ga ago and ceased roughly 1.85 Ga ago (Morey, 1983), prior to or during the Penokean orogeny of Goldich *et al.* (1961). The nature of the sediment varies considerably, ranging from volcanic and clastic to the chemical precipitates which form the thick successions of iron formation. The termination of the Penokean orogeny marked the onset of an intrusive igneous phase which emplaced subduction related tonalitic and granitic plutons into the Animikie sediments and the arc related volcanics of the Wisconsin magmatic terranes. The present form of the basin was

achieved around 1.1 Ga ago (Silver and Green, 1972; Hanson, 1975; Wanless and Lovebridge, 1978) when a north-northwest trending branch of the Midcontinental Rift System (King and Zietz, 1971) separated the Animikie sediments into a northwestern and southeastern segment (Morey, 1983).

The northwestern segment of the Animikie Group unconformably overlays the Superior Province and consists of a basal sandstone-siltstone (Pokegama Quartzite, Mahnomen Formation), iron formation (Gunflint, Biwabik, Trommald iron formations), and a thick, upper, shale-siltstone sequence (Rove, Virginia and Rabbit Lake Formations). In east-central Minnesota the Thomson Formation which has similar lithology to the Rove, Virginia and Rabbit Lake Formations, has also been considered correlative to the Animikie Group (Morey and Ojakangas, 1970). However, recent structural interpretations have complicated the correlation of these units (Ojakangas, 1995); northward thrusting during the Penokean Orogeny has juxtaposed units of differing Early Proterozoic ages in east-central Minnesota (Southwick *et al.*, 1988; Southwick and Morey, 1991; Hemming, 1994). Similar structural complexities also hinder correlation within the southeastern segment in northern Wisconsin and the Upper Peninsula of Michigan (Klasner *et al.*, 1991; Gregg 1993).

The sedimentary rocks of the Animikie Basin were deformed during the Penokean orogeny, forming a zone of deformed and metamorphosed Paleoproterozoic and Archean rocks along the southern margin of the Superior Province (Sims, 1991). It is thought to have occurred between 1.9 and 1.76 Ga (Van Schmus, 1976; Van Schmus and Bickford, 1985). In the Great Lakes region it forms the transition between the Late Archean crust of the Superior Province and the Paleoproterozoic continental crust of the Mazatzal and Yavapai provinces of the southwestern United States (Sims and Peterman, 1986; Hoffman, 1989; Barovich *et al.*, 1989). Deformation along this tectonic front involved thrusting and development of basement gneiss domes (Southwick *et al.*, 1988). Rocks of the orogeny are exposed in east-central Minnesota, northern Wisconsin and the Upper Peninsula of Michigan.

The orogeny consists of a northern zone of deformed Animikie strata and a southern assemblage of arc-related volcanics and synorogenic, calc-alkaline plutonic rocks of the Wisconsin magmatic terranes, separated by the Niagara Fault Zone in northern Wisconsin and the Upper Peninsula of Michigan (Sims *et al.*, 1989) and by the Malmo Discontinuity in east-central Minnesota (Southwick and Morey, 1991).

The earliest sedimentological models for the Animikie Basin relied on the work of James (1954) who concluded that iron formation successions evolved from a "miogeosynclinal" sequence to a "eugeosynclinal" assemblage during Paleoproterozoic time. James' original ideas were then discarded in the mid 1970's in favour of deposition of Animikie sediments in a rift-like basin (Cambray, 1978). In this model rifting was followed by the ultimate closure of the basin with the Penokean orogeny. In the early 1980's Young (1983), by analogy with other Phanerozoic and Proterozoic depositional basins, interpreted the tectonic and sedimentary history of Proterozoic rocks in the northern Great Lakes region in terms of an aulacogen model. Unfortunately, there was no general consensus on the driving force which caused the basin to form and the characteristic sediments analogous to rift/aulacogen depositional systems are absent. In the mid 1980's continued research lead to significant advancements in understanding the tectonics associated with basin development. These strides lead Hoffman (1987) to suggest deposition of chemical-sediments

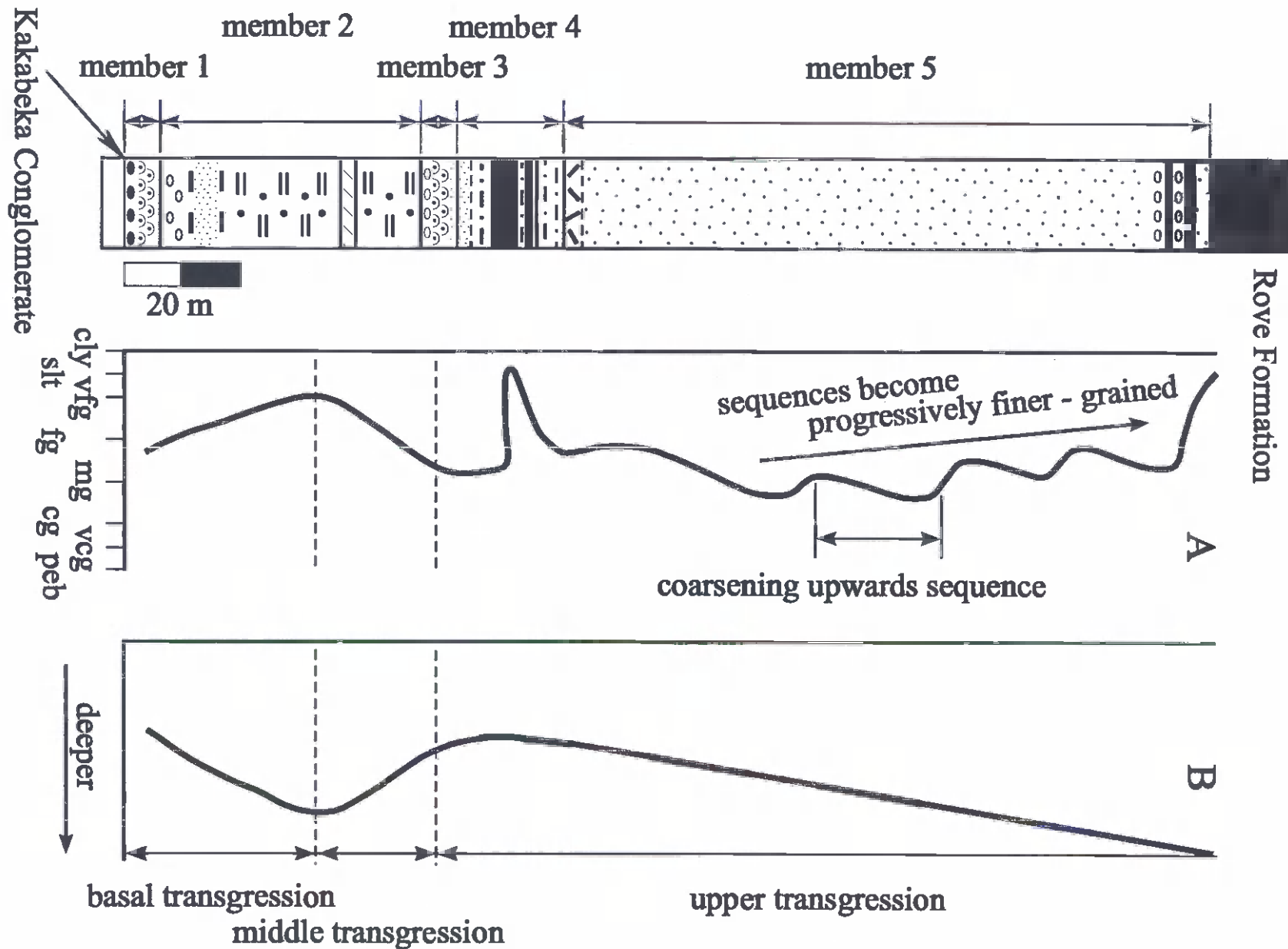


Figure 5. Stratigraphic section through the Gunflint Formation based on drill core from holes a few kilometres north of the US border. A detailed discussion of the lithofacies is presented in the text.

A) Up-section grain size trends.

B) Inferred changes in water depth.

occurred in a migrating peripheral foreland basin which formed in response to crustal loading during the Penokean orogeny. Southwick *et al.* (1988) and Southwick and Morey (1991) then reinterpreted the segment of the Penokean orogeny (Goldich, *et al.*, 1960) in Minnesota. Like Hoffman, (1987) they also concluded that the Animikie Group accumulated in a peripheral foreland basin and that iron formation deposition occurred in three distinct tectonic settings as the orogeny evolved. Recently, Pufahl (1995, 1996) and Hemming (1994, 1995) have demonstrated that this tectonic model is no longer valid. Hemming (1994, 1995) contends that neodimium and lead isotopic evidence for the provenance of the Animikie Group is consistent with the tectonic evolution of the Animikie Basin beginning as a passive margin within a back arc basin, and ending as a telescoped back arc basin that closed as a result of a change in relative plate convergence direction. Nevertheless, in recent years the foreland model has reached widespread acceptance based on high resolution aeromagnetic data, exploratory drilling through structurally complex regions, and sedimentological investigations of clastic units above and below iron-bearing strata (Hoffman, 1987; Southwick *et al.*, 1988; Southwick and Morey, 1991; Ojakangas, 1995).

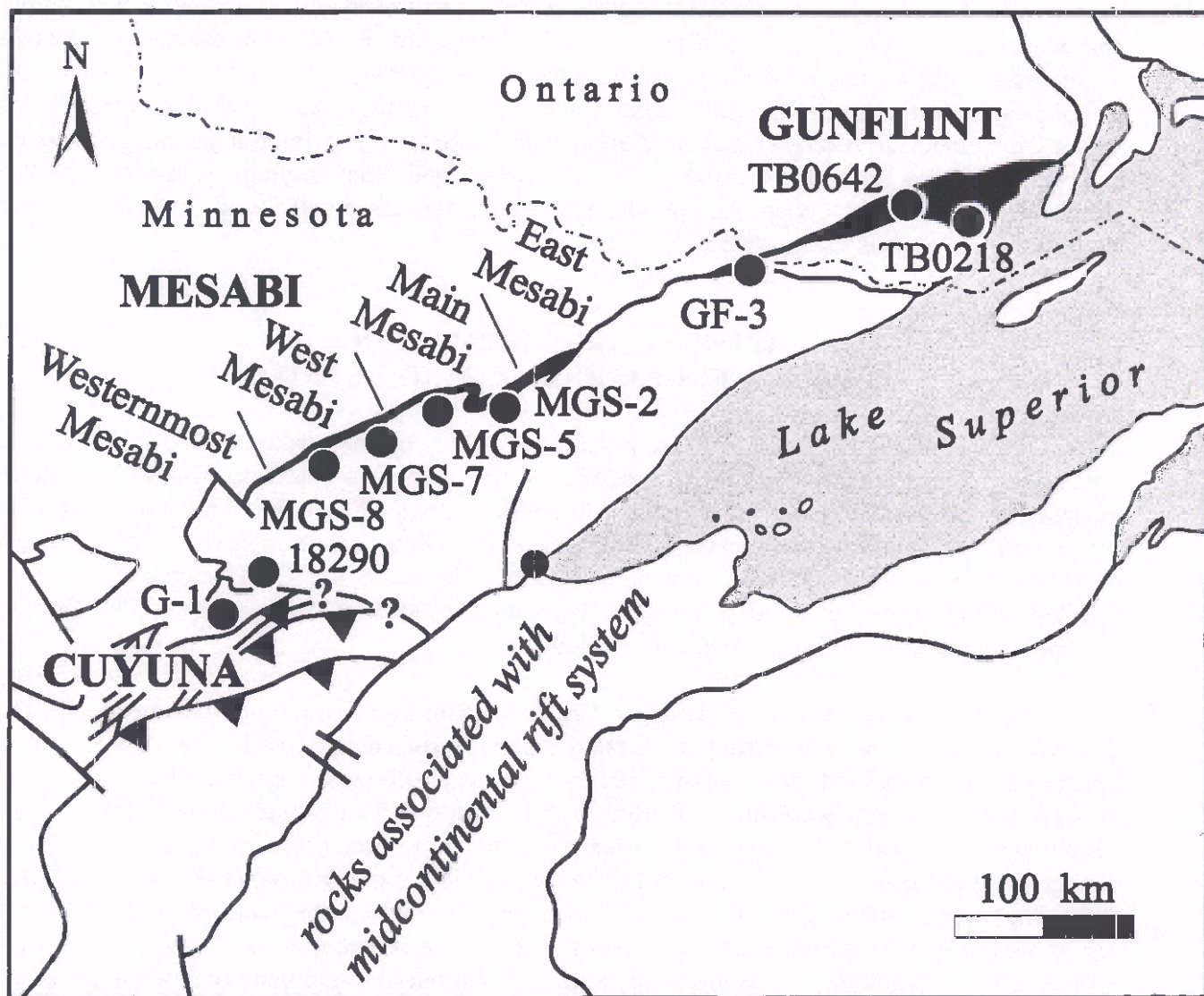
GUNFLINT IRON FORMATION PALEOGEOGRAPHIC RECONSTRUCTION

Interpretations presented here are based on the study of core drilled through the Gunflint Formation. The Formation has been divided into five members based on distinct lithofacies associations and controlled by transgressive-regressive cycles (Figure 5). The strand proximal succession present in Northwestern Ontario offlaps to the south as fine-grained, chemical sediments of the slope are approached in the Cuyuna Iron Range (Figures 6 and 7). This chapter will give detailed interpretations of the depositional environment of each of these members; but first an overview of nearshore sedimentation is presented.

Lateral and vertical facies changes within the Gunflint Formation indicate deposition occurred on a shallow, non-barred, microtidal, storm-enhanced shelf with little clastic influx. Bedforms and internal sedimentary structures composing Gunflint strata are directly comparable to those found in modern nearshore-to-shallow marine clastic and carbonate depositional systems. These facies trends record energy transformations from deep water, offshore waves to nearshore, shoaling waves under fairweather and storm conditions. Minor mafic flows (Goodwin, 1960) and extensive tuffaceous horizons (Hassler and Simonson, 1989) within the Gunflint indicate that the region was volcanically active at the time of basin formation and subsidence (Fralick and Barrett, 1995). The main morphological elements, bedforms and internal sedimentary structures along a typical beach-shoreface-offshore profile are shown in Figure 8. The following summary of these features is taken from Walker and Plint (1992), Einsele (1992) and Clifton *et al.* (1971).

The *foreshore*, also synonymous with beach, consists of the portion above the low tide line and is dominated by the swash and backwash of breaking waves. In the foreshore zone the most characteristic feature is parallel to low-angle cross-bedding dipping seaward. On non-barred high energy shorelines, two planar laminated zones occur, an inner (upper foreshore), planar zone and an outer (lower foreshore) planar zone within the zone of breaking waves. In between these two zones the bedding surface becomes rough due to dotted and irregular erosional and depositional features.

Figure 6. Location of drillholes depicted on Figure 7.



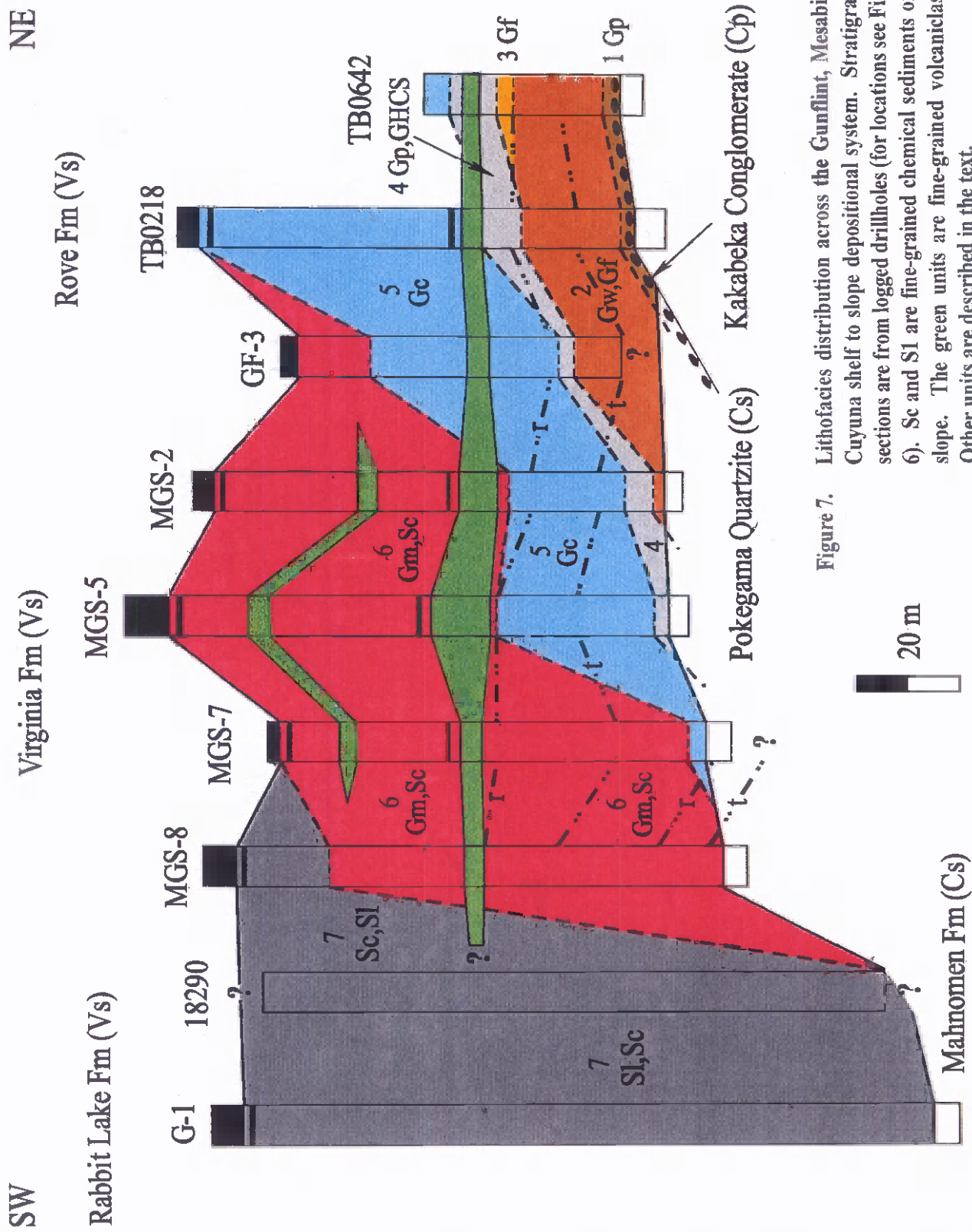
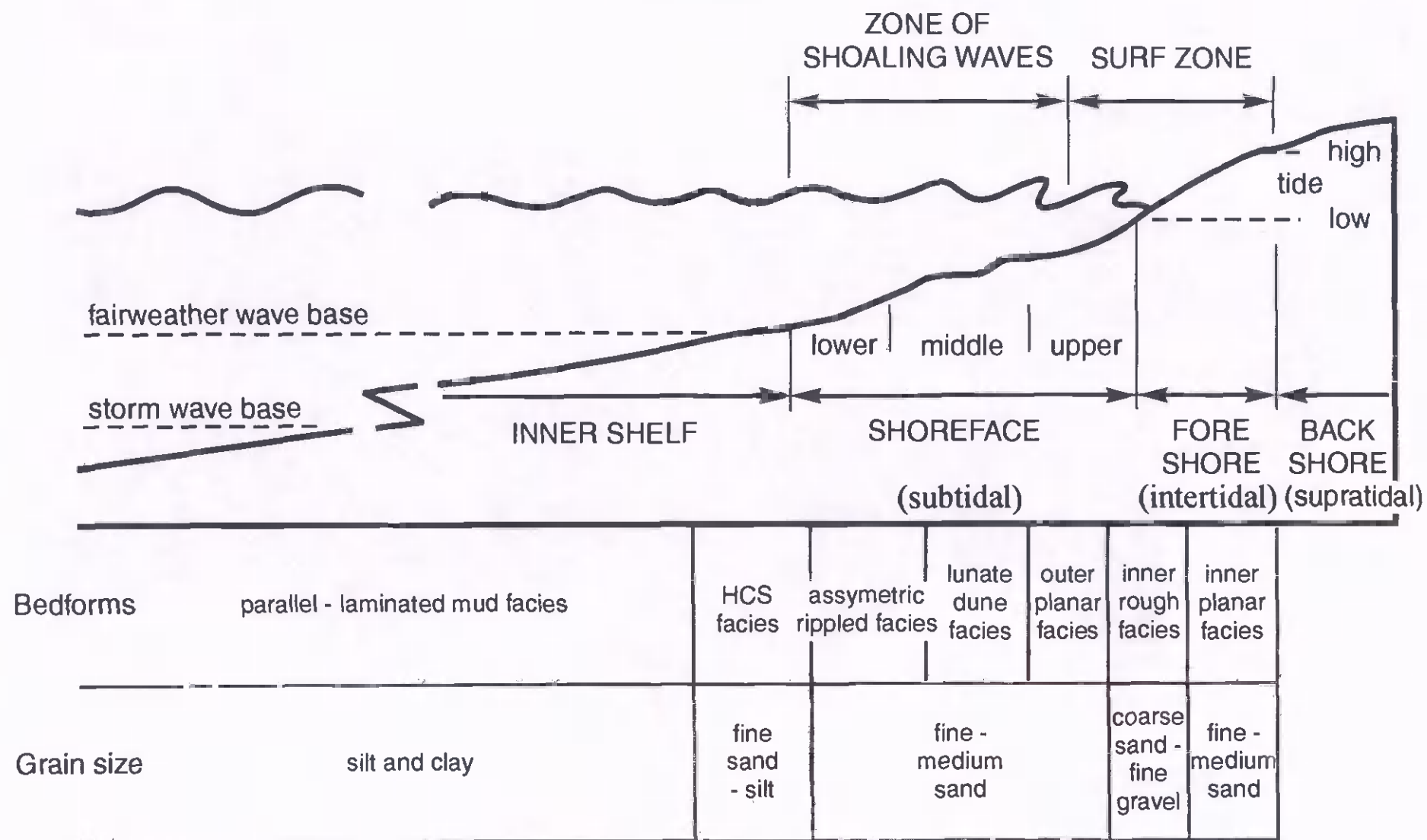


Figure 7. Lithofacies distribution across the Gunflint, Mesabi and Cuyuna shales to slope depositional system. Stratigraphic sections are from logged drillholes (for locations see Figure 6). Sc and S1 are fine-grained chemical sediments of the slope. The green units are fine-grained volcaniclastics. Other units are described in the text.

Figure 8. *Morphological elements, bedforms and internal sedimentary structures along a typical beach-shoreface-offshore profile.*



In tidally dominated systems the foreshore and upper shoreface is comprised of flaser, wavy, and lenticular bedded sands and muds typical of sedimentation on inter- and subtidal, mixed sand-mud flats. The *shoreface* lies below low tide level and is characterized by day-to-day sand transport above fairweather wave base. The shoreface normally has a concave-upward profile, which is in equilibrium with the waves that shape it. On the lower shoreface, at depths near or below fairweather wave base (10 to 20 m), are long crested symmetrical wave ripples, produced by storm waves. Landward, on the middle shoreface, these inactive ripples pass into active ripples which become increasingly asymmetric, irregular, and short-crested. Larger, trough cross stratified, lunate dunes are frequently observed on the upper shoreface in the breaker zone. All these bedforms are associated with small-scale or larger-scale cross bedding which is predominantly oriented toward the land. Because the sediments of this zone are under continuous agitation, the deposition of mud is prevented. The mud-dominated *shelf* is typically below fairweather wave base and has a lower depositional slope than the foreshore and shoreface. Mud is typically parallel laminated and highly bioturbated in modern shallow-marine depositional systems.

In storm enhanced shallow marine non-barred, environments the shoreface processes described thus far, as well as processes in deeper regions of the inner and outer shelves are strongly affected and modified by storm events (Davis, 1992; Einsele, 1992). Storms can deposit sandy material in the supra tidal zone, as well as generate special non-channelized flow conditions in deeper water. These high energy events frequently produce characteristic sheet-like sand and mud beds, called *tempestites*, of considerable lateral extent (Johnson and Baldwin, 1986; Morton, 1988; Nummedal, 1991; Einsele, 1992). Storm-generated bed forms show a distinct trend from the beach into deeper water. The upper shoreface (surf zone) displays dunes oriented parallel to the shoreline. When flow becomes channelized longshore and rip currents erosively scour fairweather sediments, depositing medium- and coarse-grained, trough and planar, cross stratified sands. The middle shoreface is dominated by both flat and swaley bedforms with nearly horizontal or low-angle swaley cross stratification. Because of the continual agitation of sediments through wave action on the upper and middle shoreface the record of heavy storms is usually destroyed. On the lower shoreface (5 to 20 m water depth), the stress of storm surge ebb currents reaches about the same magnitude as the oppositely directed storm-wave induced oscillatory flow (Einsele, 1992). Thus, for a short time, both currents may stir up sand and mud, which are redeposited as parallel and hummocky cross-stratified fine-grained, graded beds at the same location or nearby (Einsele, 1992; Dott and Bourgeois, 1982). On the inner shelf and parts of the outer shelf, particularly in high-energy environments, storm-induced combined flow may still generate HCS or thin, graded, sandy beds resulting from laterally flowing suspension currents. If the suspension originates from storms and the sandy layers alternate with shelf muds, the storm induced beds represent distal tempestites (Aigner, 1985; Einsele, 1992). At greater water depths on the shelf, the current component of the combined storm flow becomes dominant, leading to current rippled, offshore, fine sand and silt beds and graded rhythmites (Aigner, 1985; Einsele, 1992). Since only the largest storms can affect the sea floor in this zone, storm events are followed by long periods of quiescence and mud deposition.

Paleogeographic interpretation

The following discussion has been divided into three sections. Each deals with interpreting the vertical and lateral facies trends comprising the basal transgressive, middle regressive and upper transgressive depositional cycles of the Gunflint Formation in detail.

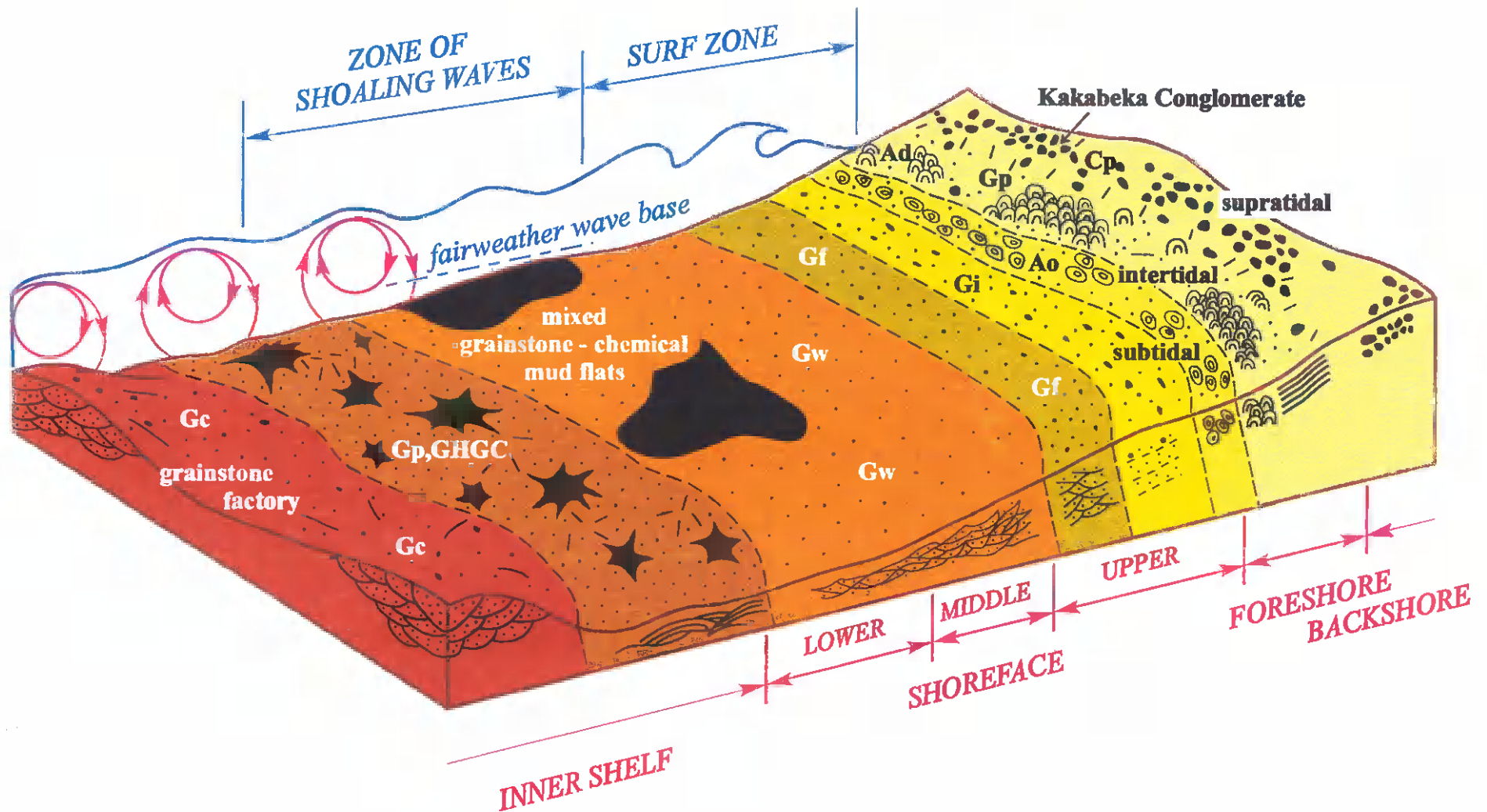


Figure 9. Paleogeographic reconstruction of the depositional system which produced the Gunflint Formation. A detailed discussion of the attributes of this system is presented in the text.

Basal transgression

This 20 m thick, upward fining cycle extends from the base of the Animikie Group to the middle of member 2. The base of the cycle is characterized by the Kakabeka Conglomerate, a backshore, pebbly beach deposit preserved as a transgressive pebble lag (Figure 9). This intraformational conglomerate was formed by the localized weathering of the peneplain. Then, through subsequent storm-generated wave action (swash/backwash) these intraclasts were reworked into crudely stratified, matrix supported, pebbly, supra tidal, beach deposits. Ojakangas (1983) has made a similar interpretation of the conglomeratic unit forming the base of the Pokegama quartzite in the Mesabi iron range in east-central Minnesota.

As the basal transgression progressed these coarse beachface sediments were on-lapped by the fine- and medium-grained, parallel laminated grainstones (**GP** facies) of member 1. The hematite-enriched transition from member 1 may indicate a change in the redox state of the depositional system accompanying a rise in sea level or, more simply, it may represent oxidation of iron-bearing sediments by late phase oxygen-rich meteoric waters percolating along this lithologic boundary. Deposition of this thin, parallel bedded grainstone unit marks the onset of intertidal sedimentation on the foreshore (Figure 9). Member 1's relative thinness (5 m) and apparent lack of desiccation features suggests that the intertidal zone was not well developed on the Animikie shoreline, a situation typical of modern, microtidal environments (tidal range <2 m) (Reineck and Singh, 1975). LLH stromatolitic bioherms (**Ad** facies) at the base of member 1 may have developed in much the same way modern, columnar stromatolites and crypt algal knolls form in the low energy intertidal flats of Shark Bay, Australia (Hoffman, 1976; Schreiber, 1986). Here columnar stromatolites develop in the lower intertidal zone in response to strong wave action. An oncolitic horizon (**Ao** facies) above these intertidal, algal mats marks the transition into the subtidal (Figure 9) (Logan *et al.*, 1964; Ginsburg, 1960, Sellwood, 1986). Ginsburg (1960) compared modern oncolites from the Florida and Bahama Banks to Devonian, Permian and Precambrian examples and found they formed 1 to 8 feet below mean water level at or near the subtidal transition. Oncolites from the Paris Basin, a non-barred, shallow epeiric sea covering much of northwestern Europe during the Middle Jurassic, were also determined to characterize this transition (Sellwood, 1986).

The overlying flaser and wavy bedded chert-carbonate grainstones (**Gf**, **Gw** facies) of member 2 represents subtidal deposition on a shallow, storm enhanced, non-barred shoreline (Figure 9). Similar Middle and Upper Cambrian, cyclic, peritidal, carbonate, tempestites compose the Elbrook and Conococheague Formations in the Virginia Appalachians (Demicco, 1985, Koerschner and Read, 1989). This 1.6 km thick carbonate sequence formed on the Cambro-Ordovician continental shelf; an aggraded, rimmed, mature, passive continental margin (Koerschner and Read, 1989). Cycles typically range between 1 and 7 m in thickness, possess a sharp basal transgressive surface, are overlain by a subtidal unit passing into a tidal flat cap, and record regional changes in sea level over very short intervals of time (Koerschner and Read, 1989). Three types of cycles are recognized, thin-based cycles, thrombolitic-based cycles, and thick based cycles. The latter has been divided into three subcycles, those which are digitate-algal-bioherm-based, grainstone-based and ribbon-carbonate-based. Ribbon-carbonate-based cycles most closely resemble member 3 and therefore will be the only cycle fully reviewed in the following discussion. For a full treatment of these cyclic, peritidal carbonates the reader is referred to Koerschner and Read (1989).

Ribbon-carbonate-based cycles are composed of 1 to 5 m thick fining and coarsening upwards sequences of flaser bedded coarse-grained carbonate grainstones/packstones which are comprised of wavy and lenticular bedded finer-grained pelletal carbonates and rare lime conglomerates intimately interbedded with parallel laminated dolomitic mudstones. The term comes from the field appearance of the alternating thin-beds (ribbons) of differentially weathered limestone and dolostone (Demicco, 1983). The coarse-grained ribbon carbonates and pellet packstones are most similar to the flaser bedded, chert-carbonate grainstones (Gf facies) of member 3 in the Gunflint chemical sedimentary succession. Coarse-grained ribbon carbonates are unlayered or have low-angle cross bedding with weakly graded foreset laminae. Pellet packstones have horizontal laminations, climbing wave-ripple lamination and wave ripple cross-lamination. Mud-poor pellet-rich ribbon carbonates and mud-and-pellet rich ribbon sequences most closely resemble wavy bedded, chert-carbonate grainstones (Gw facies). Mud-poor, pellet-rich ribbon carbonates are planar laminated; mud-and-pellet-rich ribbon sequences are comprised of wavy and lenticular bedded pellet layers interbedded with form-discordant wave ripples, starved ripples and parallel laminated dolomite mudstones (Koerschner and Read, 1989).

Demicco (1983) and Koerschner and Read (1989) interpret these ribbon sequences as recording shallowing by aggradation from storm-influenced subtidal settings floored by interlayered lime muds and sands, to subtidal sand flats on a mature, rimmed shelf. Coarsening-upward sequences record low energy, subtidal conditions, followed by upward-shallowing into high energy shallow water settings (ie. surf zone and breaker zone). Koerschner and Read (1989) contend these coarse-grained ribbon carbonates formed on sand flats, where coarse layers were reworked during storms and less common fine layers settled out as storm/tidal energy waned. The rare lime conglomerate interbeds were deposited during storms which swept semi-lithified, peloidal sands from shallow subtidal and intertidal settings. Fining upwards ribbon sequences record transgressive high-energy conditions that gave way to low energy conditions as the platform shallowed. Fine-grained pelletal, ribbon carbonates formed on lower energy, subtidal, mixed sand flats in response to wave action and grade landward into mixed sand-mud flats (Koerschner and Read, 1989). Fairweather reworking of wave rippled beds alternated with deposition of thick storm layers of pelletal sediment with amalgamated hummocky cross stratification, and wave and current ripples (Koerschner and Read, 1989). Similar clastic, shallow, tidal marine sequences occur in the modern North Sea (Demicco, 1983).

Unlike the Cambrian ribbon carbonates of the Virginia Appalachians these flaser and wavy bedded chemical sediments did not form by aggradation of upward shoaling grainstone sequences. Instead, vertical changes in iron formation facies reflect a rate of relative sea level rise which consistently outpaced that of sedimentation.

Middle regression

This minor, 20 m thick, regressive phase begins in the middle of member 2 and extends to the top of member 3. It records the gradual onset of shallowing from the middle shoreface to the lower foreshore. The subtidal mixed grainstone-chemical mud flats of member 2 coarsen upwards to 3 m thick packages) of flaser bedded chert-carbonate grainstones. These coarse packages formed during regressive offlap of lower-middle shoreface facies and, like the flaser bedded package at the base of this member, represent dune development in the surf zone of the upper shoreface.

As sea level continued to fall these flaser and wavy bedded chert-carbonate grainstones were transgressed by the coarser, higher energy, stromatolitic and oncolitic, chert-carbonate, grainstones of member 3. These storm enhanced, crypt algal, lower foreshore deposits mark the regressive peak of this middle cycle. Like the **Gf** facies chert-carbonate grainstones of member 2, those comprising member 3 also record lunate dune development on a subtidal/lower intertidal chert-carbonate grainstone flat.

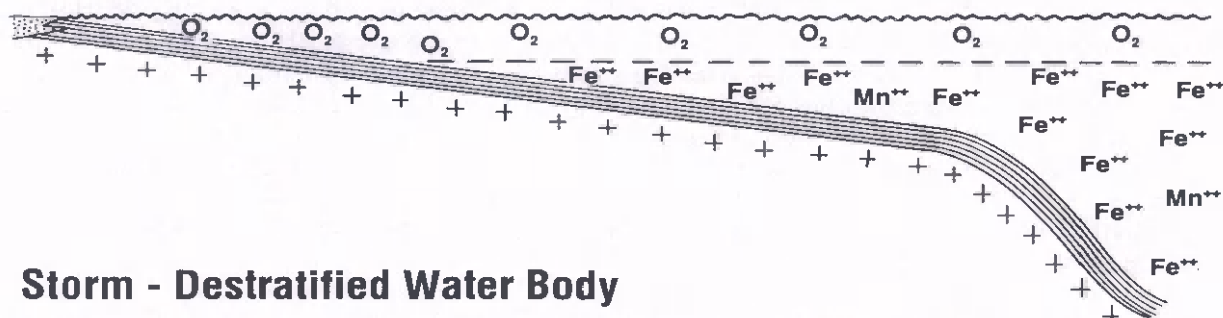
The 6 cm thick, **Gi** facies grainstone bed at the base of member 3 in O'Connor township is interpreted as being a proximal tempestite deposited in this shallow, high energy environment. This coarse, storm generated lag is thin, possesses a base which appears to erosively scour the flaser bedded chert-carbonate grainstones of member 2, contains abundant magnetite- and hematite-rich rip ups and pinches laterally towards Blake township; all are characteristics of nearshore sheet storm deposits (Einsele, 1992). With a return to fairweather conditions this storm bed provided an ideal substrate for stromatolites to colonize. It was firm and not easily reworked by normal shelf currents (Fralick, 1988). Development of crypt algal knolls on this intraformational lag was however not widespread as they colonized only the most stable regions sheltered from fairweather reworking. The thin, discontinuous, stromatolitic and oncolitic horizons which developed above this basal, crypt algal layer suggests subtidal, stromatolitic growth occurred on a substrate that underwent periodic, fairweather reworking. The development of these poorly formed, discontinuous LLH and SS stromatolitic bioherms is typical of crypt algal growth on shallow, high energy, subtidal sand sheets of the upper shoreface and lower foreshore (Pratt *et al.*, 1992).

Upper transgression

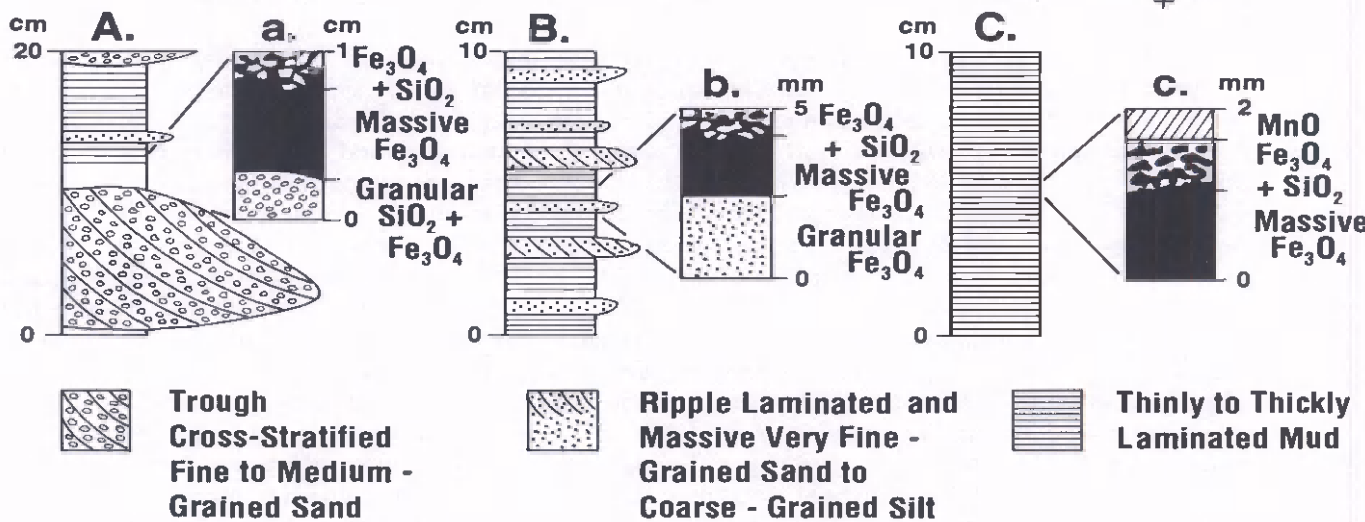
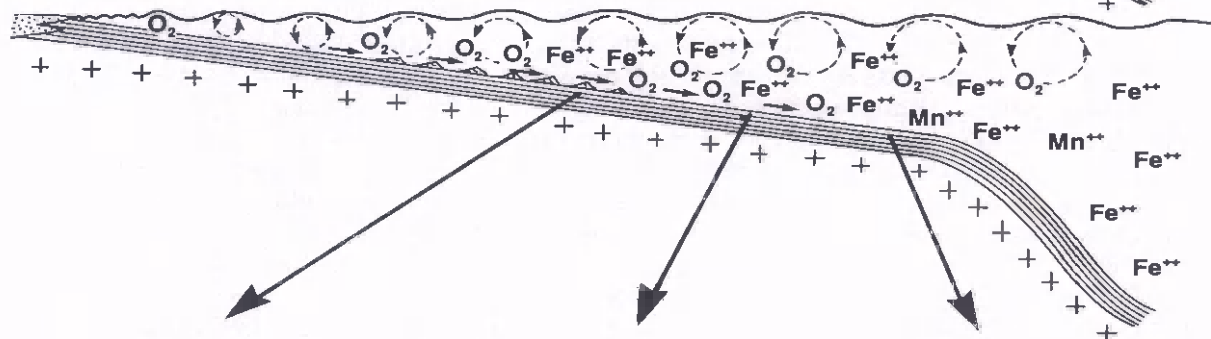
The upper transgressive cycle extends from the top of member 3, through member 5, possibly to the top of the Rove Formation; a vertical distance of almost 700 m. Its onset is rapid and records the progressive onlap of offshore facies over those of the lower shoreface. The thin package of hummocky cross stratified chert- and hematite-rich grainstones (**GHCS** facies) at the base of member 4 in O'Connor township marks this transition. Modern hummocky-cross-stratified sands occur below fairweather wave base on the lower shoreface of storm dominated, shallow marine depositional systems at or near the transition to the inner shelf (Hamblin and Walker, 1979; Dott and Bourgeois, 1982; Aigner, 1985; Walker and Plint, 1992; Einsele, 1992; Davis, 1992). The parallel bedded, hematite-rich, slaty and cherty iron formation beds (**Gp** facies grainstone) to which this basal hummocky-cross-stratified unit grades are also characteristic of storm deposition near this change (Figure 9). They are identical to modern, proximal offshore, storm-sand layers (proximal tempestites) and their more distal equivalents, graded rhythmites (distal tempestites) (Reinick and Singh, 1972; Aigner, 1985). Like the parallel bedded chemical sediments these storm beds also possess a sharp base, and in some instances may grade upward into plane laminated mud layers. In modern tempestites the sand layers are either evenly laminated or are laminated rhythmites, in which the lower laminae are thicker and coarser-grained and grade upwards into thinner and finer-grained laminae (Reinick and Singh, 1972). They are the result of deposition from shallow marine, storm generated turbidity currents (Reinick and Singh, 1972; Walker, 1984; Einsele, 1992; Davis, 1992). The fine- and medium-grained chert- and hematite-rich grainstone beds comprising this member represent episodic development of storm-generated, chemical sedimentary beds. Thicker, graded, coarser beds developed during the most severe storms. The hematite-rich slaty iron formation packages between storm-grainstones represents the return to fairweather sedimentation on the lower

Figure 10. *Depositional model for the precipitation of iron-rich sediments on the Animikie shelf. The model is based on data collected from the Gunflint and Gogebic iron ranges. Samples from these areas were examined in thin section under a petrographic microscope and using a scanning electron microscope with an energy dispersive system. Whole rock analysis was also performed using an inductively coupled plasma. During the Paleoproterozoic the world ocean contained much less oxygen than at present (Cloud, 1973). This enabled Fe^{++} and Mn^{++} to stay in solution. Oxygen would only be present in appreciable quantities in the shallow, near shore environment where reasonable levels of light penetration to the bottom would allow the cyanobacteria colonies in this area to thrive, and thus release oxygen (1). During storm events the combined effects of an extreme low pressure system over the area and wave forcing caused the buildup of water in a strand-proximal position. Upon waning of the storm event this water would be returned to the mid and outer shelf via a storm surge ebb flow. The deliverance of more highly oxygenated waters to the deeper shelf would cause the rapid precipitation of iron oxide (probably as iron hydroxyoxide) (2). At high Fe/Mg ratios it is likely that precipitation of iron compounds would dominate during this rapid precipitation phase. As oxygen and/or iron is/are depleted precipitation rate slows and silica gel, which is slowly but constantly precipitating from seawater, increases in importance as a constituent of the layer. Finally, with depletion of iron in the aqueous phase due to precipitation, and a resulting decrease in the Fe/Mg ratio, manganese begins to effectively scavenge oxygen forming MnO-rich tops to some layers. If another storm event occurs before the Fe/Mg ratio of seawater has decreased to the point manganese precipitation becomes significant the upper portion of the layer will not form. Smaller storms are able to cause the above processes to operate in the mid-shelf area, whereas only large storms effect the outer shelf. Thus, manganese concentrations are much higher in rocks formed in the outer shelf-slope environment. A) Typical sequence found in the mid-shelf where thick (1 cm) chemical sediment laminae directly overlay storm generated granule lenses. B) Typical laminae from the outer mid-shelf. C) Thin, complete laminae deposited in the outer helf.*

1. Fairweather Period - Stratified Water Body



2. Storm - Destratified Water Body



shoreface. In Blake township these chemical sedimentary storm layers lay directly on the grainstones of member 2.

The pulse of volcanism preserved in the middle of member 4 as a 6m thick volcanoclastic horizon (**Vs** facies) occurred swiftly, outpacing iron formation deposition. Its sharp bottom contact indicates rapid influx of volcanoclastic material into the depositional system. Its gradational upper contact suggests delivery of volcanoclastics to the basin waned gradually over time. The lapilli tuff (**Vt** facies) which characterizes the base of member 5 is interpreted as being reworked and deposited by storms from more proximal, nearshore settings. The massive, medium- to coarse-grained chert-carbonate grainstone bed (**Gi** facies) characterizing its base is similar to the stromatolitic, storm generated lag in member 3; its lower bounding surface is erosive into the top of member 4 and it also pinches laterally towards Blake Township. Similar volcanoclastic beds within the Gunflint Formation have also been interpreted by Hassler and Simonson (1989) as being reworked and redeposited by turbidity currents.

The coarsening upwards, massive and cross-stratified, chert-carbonate, grainstone successions (**Gc** facies) of member 5 mark the progressive onlap of inner shelf facies over those of the lower shoreface parallel laminated and hummocky cross-stratified grainstones (**Gp** and **GHCS** facies) (Figure 9). Similar successions deposited in coastal and open marine environments comprise the Upper Cretaceous, Sussex and Shannon sandstones of the Powder River Basin in northern Wyoming. In Wyoming cycles coarsen from silty offshore mudstones, through thin bedded, storm- and tide-generated sandstones to cross stratified sandstones with abundant clay rip-ups. The depositional model proposed for the Shannon and Sussex sandstones consists of migrating, offshore bar complexes that have their surfaces covered by sand waves (Berg, 1975; Brenner, 1978; Hobson *et al.*, 1982; Davis, 1992). With some modification this model may also be applied to the coarser-grained, shoaling upwards sequences comprising member 5 of the Gunflint Formation.

Like bar development in clastic systems genesis of chemical sedimentary, offshore bar complexes depends on, 1) the availability of coarse sediment to offshore environments and 2) the presence of currents capable of shaping this sand into well defined, migrating, sand bodies. In clastic, shallow marine depositional systems coarse detritus is supplied to offshore areas by storm (Berg, 1975; Walker, 1984; Tillman and Martinson, 1984) and tide (Spearing, 1976; Brenner, 1978; Seeling, 1978) generated currents. However, unlike their clastic counterparts, bar genesis in chemical sedimentary systems relies on offshore precipitation of iron oxides and silica gel from upwelled enriched bottom waters (Borchert, 1950; Drever, 1974; Morris, 1993), and subsequent current reworking of the substrate into rip-up grains (Simonson and Goode, 1989) (Figure 10). Consequently this process does not depend on systems tracts capable of delivering coarse detritus to the shelf. Bar development in a chemical sedimentary, shallow-marine depositional system is therefore governed by the position of areas on the shelf with high, primary chert/carbonate grain production and the presence of currents able to rework these chemical precipitates into bar complexes. The relative stratigraphic position of member 5, over facies characteristic of the lower shoreface (**Gp** and **GHCS** facies), suggests that regions of primary grain genesis were below fairweather wave base and that storm-generated, oscillatory flow was responsible for reworking the substrate into rip-up grains.

Finer-grained, shoaling upwards, chemical sedimentary successions from the Gogebic Iron Range in Wisconsin (Figure 11) (Pufahl, 1994) are identical to the coarsening upwards successions

comprising the Sussex and Shannon sandstones in Wyoming. They are regarded as being the distal equivalents to the coarse offshore bars of the Gunflint (Pufahl, 1994). Cycles in Wisconsin coarsen from parallel laminated, magnetite-rich slaty iron formation in much the same way the Cretaceous upwards shoaling cycles coarsen from shelf mudstones.

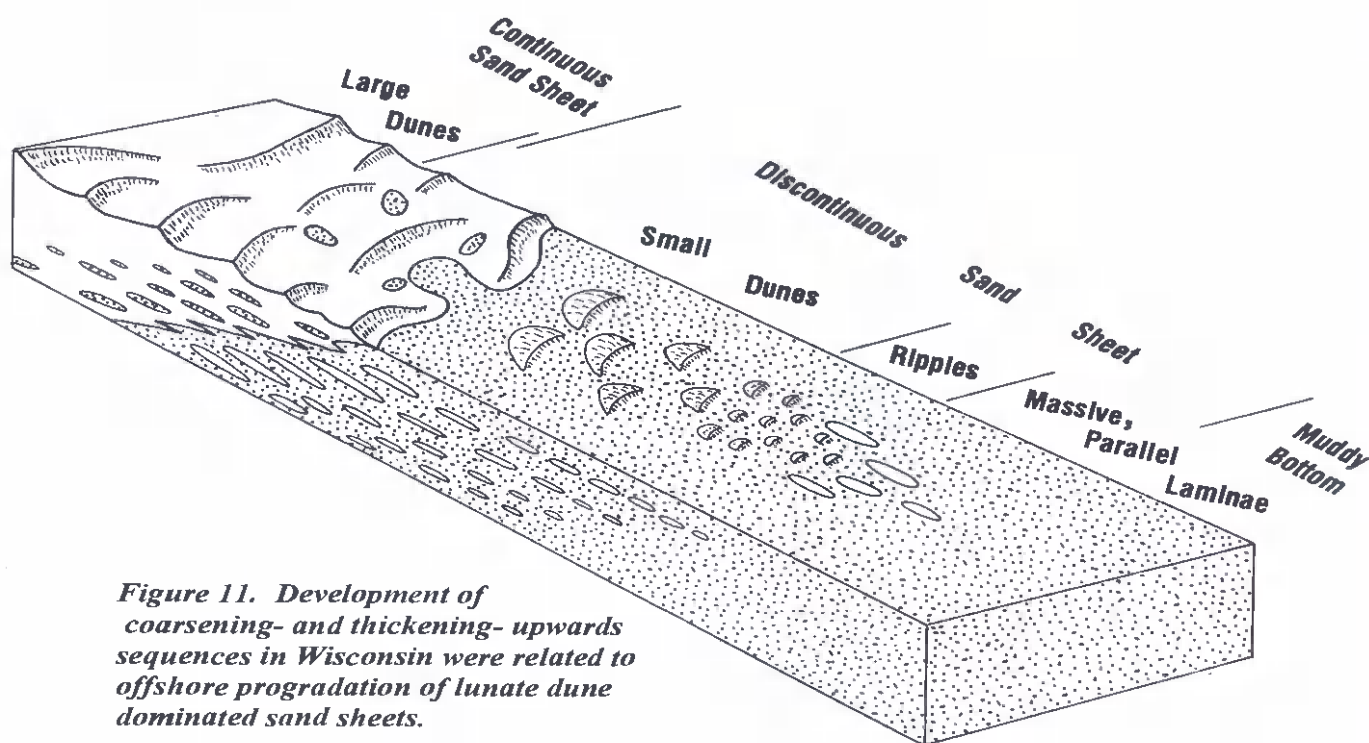


Figure 11. Development of coarsening- and thickening-upwards sequences in Wisconsin were related to offshore progradation of lunate dune dominated sand sheets.

In a broad sense these offshore regions of primary grain genesis are similar to the "subtidal carbonate factories" responsible for grain production on modern and some ancient carbonate platforms (James and Kendall, 1992). The coarse nature of the Gunflint shoaling upwards cycles in Blake and O'Connor townships suggests bar development occurred at or near a "grainstone factory". As the transgression progressed these factories migrated in step with shoreline retreat resulting in the deposition of slightly finer-grained bar complexes further from regions of primary grain production. The existence of these offshore factories would account for the absence of a mud dominated offshore, which is so characteristic of clastic shallow-marine depositional systems (Walker and Flint, 1992).

Alternatively, these coarsening upwards, Gc facies grainstone successions may represent paracycle development (Vail *et al.*, 1977) in regions of high primary grain genesis during relative sea level rise. Paracycles form in response to a punctuated rise in overall sea level. They record a period of rapid, relative sea level rise and still stand, followed by another relative rise with no

significant fall intervening (Vail *et al.*, 1977). The sequence that accumulates is coarsening upwards, suggestive of continued shoaling as sediments accumulate to sea level. In this scenario the medium-grained, chert-carbonate grainstones at the base of coarsening upward cycles would represent grainstone deposition during rapid sea level rise as regions of primary grain genesis migrated shoreward. The coarse-grained tops of cycles would represent accumulation of grainstones near to base level where breaking waves were able to winnow the substrate during an interval of relative calm. At the present time data is insufficient to choose whether an autocyclic (depositional) or allocyclic (sea level) mechanism governed cycle genesis.

The Gi facies chert-carbonate grainstone beds which characterize transitional facies into the overlying Rove in Blake township are interpreted as representing deposition from suspension clouds generated by either storms or tectonism during basin subsidence.

OVERVIEW OF THE ANIMIKIE BASIN

The iron-bearing rocks in Ontario and Minnesota form a chemical sedimentary wedge which fines and thickens from coarse, wave reworked, nearshore deposits of the Gunflint Formation to offshore, parallel-laminated, chemical mudstones of the Trommald iron formation in east-central Minnesota. Lateral and vertical facies transitions are directly comparable to those observed in modern shallow marine clastic and carbonate depositional systems. They record energy transformations from deep water, offshore waves to nearshore, shoaling waves under fairweather and storm conditions. Paleocurrent data suggests deposition of chemical sedimentary rocks occurred on a southward facing paleoslope.

Nearshore chemical sedimentary rocks consist of successions of strand-proximal stromatolites, flaser and wavy bedded chert-carbonate grainstones, and parallel bedded and hummocky cross stratified hematite-rich cherty grainstones deposited on a non-barred, microtidal, storm-enhanced shoreline with little clastic influx. Deposition on the inner shelf was dominated by accumulation of coarse-grained, shoaling upwards, chert-carbonate grainstone successions, possibly associated with offshore bar development. Their genesis is believed to have been the result of a unique interplay between the chemical and physiological conditions which prevailed in the Animikie Basin during iron formation deposition. They formed in proximal offshore areas where iron oxide and silica gel were actively precipitating, and where currents could rework chemical precipitates into rip-up grains and redistribute them. The shoaling upwards successions replace a mud dominated offshore, which is so characteristic of modern shallow-marine depositional systems.

Distally, coarsening upwards sequences become finer-grained and less well developed. They coarsen from parallel laminated, magnetite-rich chemical slates to medium-grained, massive and cross stratified, chert-carbonate grainstones, and are interpreted as recording progradation of offshore bar complexes on the middle shelf below fairweather wave base, away from areas of high primary grain genesis. These successions may also represent paracycle development during an overall rise in relative sea level.

Monotonously interbedded parallel and wavy bedded chemical slates characterize deposition of iron formation on the outer shelf and slope-break. The presence of rare, massive and cross

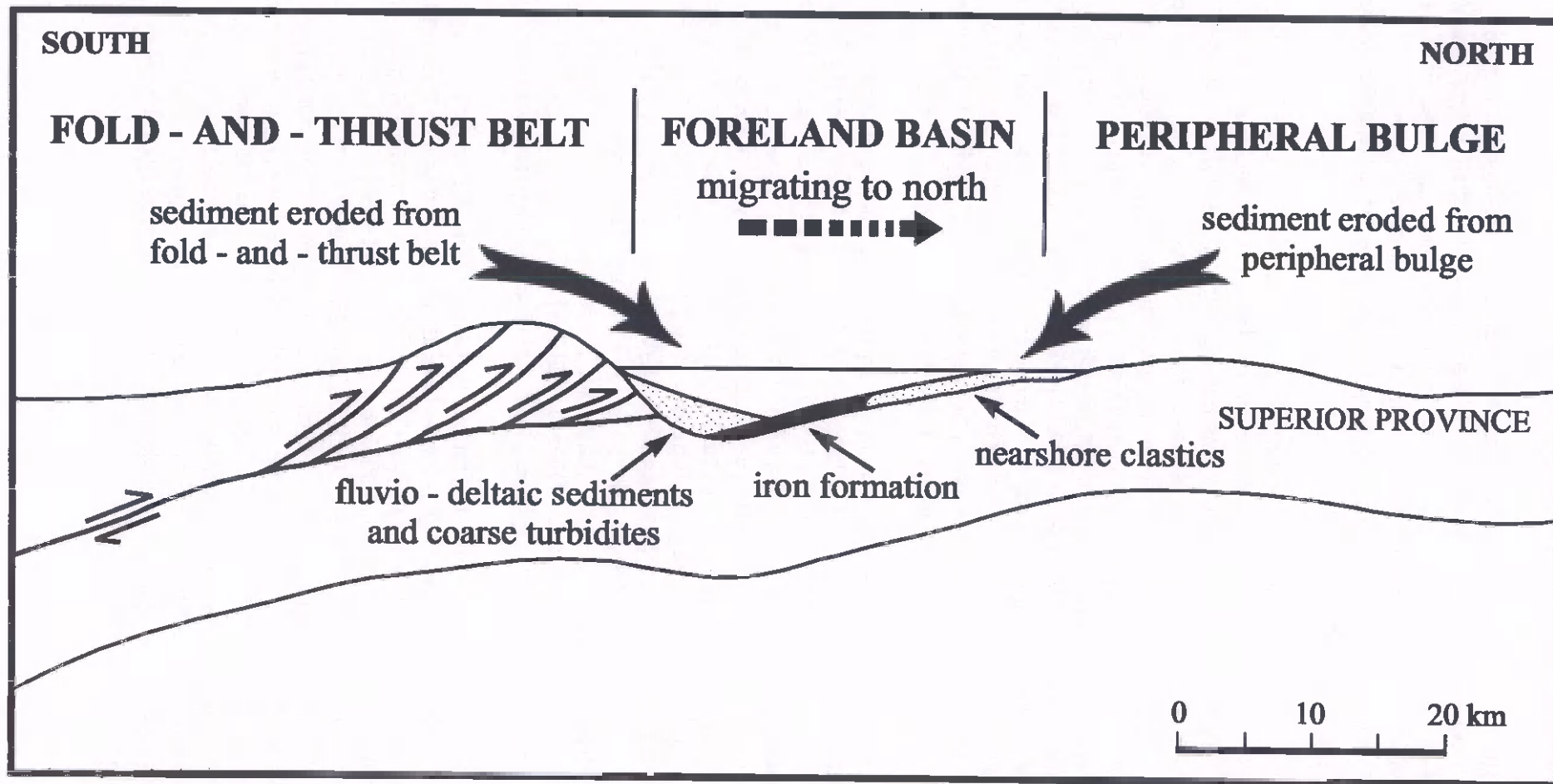
stratified, chert-carbonate grainstones and graded, magnetite-rich slaty laminations indicates that storm-generated turbidity currents originating on the inner shelf, where grain production took place, were an important mechanism for moving sediment onto the distal shelf. The existence of iron formation breccia (Schmidt, 1963) in the Cuyuna north range suggests that mass flow processes generated by large-scale rotational slumping and sliding on the distal shelf were important in moving sediment from the slope to the rise.

Sedimentological and stratigraphic data do not provide definitive answers to the question of plate tectonic setting of the Animikie Basin. However, paleogeographic reconstructions based on vertical and lateral facies transitions present within the Gunflint, Biwabik, and Trommald iron formations, and clastic units above and below iron-bearing strata indicate that it is doubtful deposition of the Animikie Group occurred within a migrating peripheral foreland basin (Figure 12). There is an apparent lack of coarse turbiditic, deltaic, and fluvial sediments, adjacent to the fold thrust belt, and paleocurrent data from clastic and chemical sedimentary rocks are opposite to what should be observed if deposition took place in a foreland basin. Also, if iron formation was to have accumulated in a foreland setting it should be restricted to the distal shelf as weathering of the peripheral bulge would shed clastics into nearshore regions effectively preventing deposition of iron formation. Facies transitions within iron-bearing members indicate the contrary, a complete shelf sequence, not just the distal portion is present.

Although data is insufficient to reconstruct the tectonic evolution of the Animikie Basin in detail, some general inferences regarding the depositional history of the Paleoproterozoic strata in Ontario and Minnesota can be made. Sedimentation within the basin began with an overall marine transgression onto the southern margin of the peneplained Superior province and deposition of the Pokegama Quartzite and Mahanomen Formation. The Pokegama Quartzite accumulated in nearshore, tidally influenced settings while the Mahanomen Formation was deposited simultaneously in more distal shelf regions by pelagic settling and waning turbidity currents. This phase of clastic deposition was probably associated with the early history of basin development. Through time, the continuing erosion of the source terrain resulted in a reduction of clastic supply to the basin, to a point where iron formation began to accumulate on the shelf, away from the land-derived clastics still accumulating in the nearshore. Once the highland was reduced to near base level iron formation was then able to accumulate in the nearshore, and directly onlap the Superior province.

Deposition of iron formation occurred during an overall transgressive marine event punctuated by two minor regressions. Lateral facies transitions within iron-bearing units record a complete shelf sequence which fines and thickens from coarse, wave reworked nearshore deposits to parallel-laminated chemical mudstones of the distal shelf. The existence of minor mafic flows and tuffaceous horizons in the Gunflint, and the presence of three volcanoclastic horizons in the Gunflint and Mesabi iron ranges indicate that the region was volcanically active during subsidence in this phase of basin development. The thickest volcanoclastic horizon forms the base of the Rove, Virginia, and Rabbit Lake Formations, and marks the transition from chemical to clastic sedimentation. This volcanic episode stopped iron formation genesis by choking the depositional system with volcanoclastics. As relative sea level continued to rise deposition within the Animikie Basin was dominated by accumulation of clastics derived from the north by southward flowing turbidity currents.

Figure 12. The Animikie Basin modelled as a migrating peripheral foreland basin. Arguments against this tectonic interpretation are presented in the text.



FIELD TRIP STOPS

ITINERARY

- | | |
|----------------|--|
| Stop 1 | Basal unconformity, stromatolites; Nolalu turnoff |
| Stop 2 | Basal unconformity, stromatolites; Kaministiquia River |
| Stop 3 | Tuffs, volcaniclastic sediments; Kakabeka Falls |
| Stop 4 | Ribbon chert-carbonate; Kakabeka Falls |
| Stop 5 | Basal pebbly sandstone; Oliver Road |
| Stop 6 | Stromatolites, beachrock; Rupert Street |
| Stop 7 | Collapse breccia; base of High Street Park |
| Stop 8 | Grainstones, stromatolites; Current River |
| Stop 9 | Slumped ribbon chert-carbonate; road to Pass Lake |
| Stop 10 | Rippled siltstones; Transcanada Highway |
| Stop 11 | Grainstones with volcanic sherds; Terry Fox Lookout |
| Stop 12 | Grainstones; Transcanada Highway near Current River |

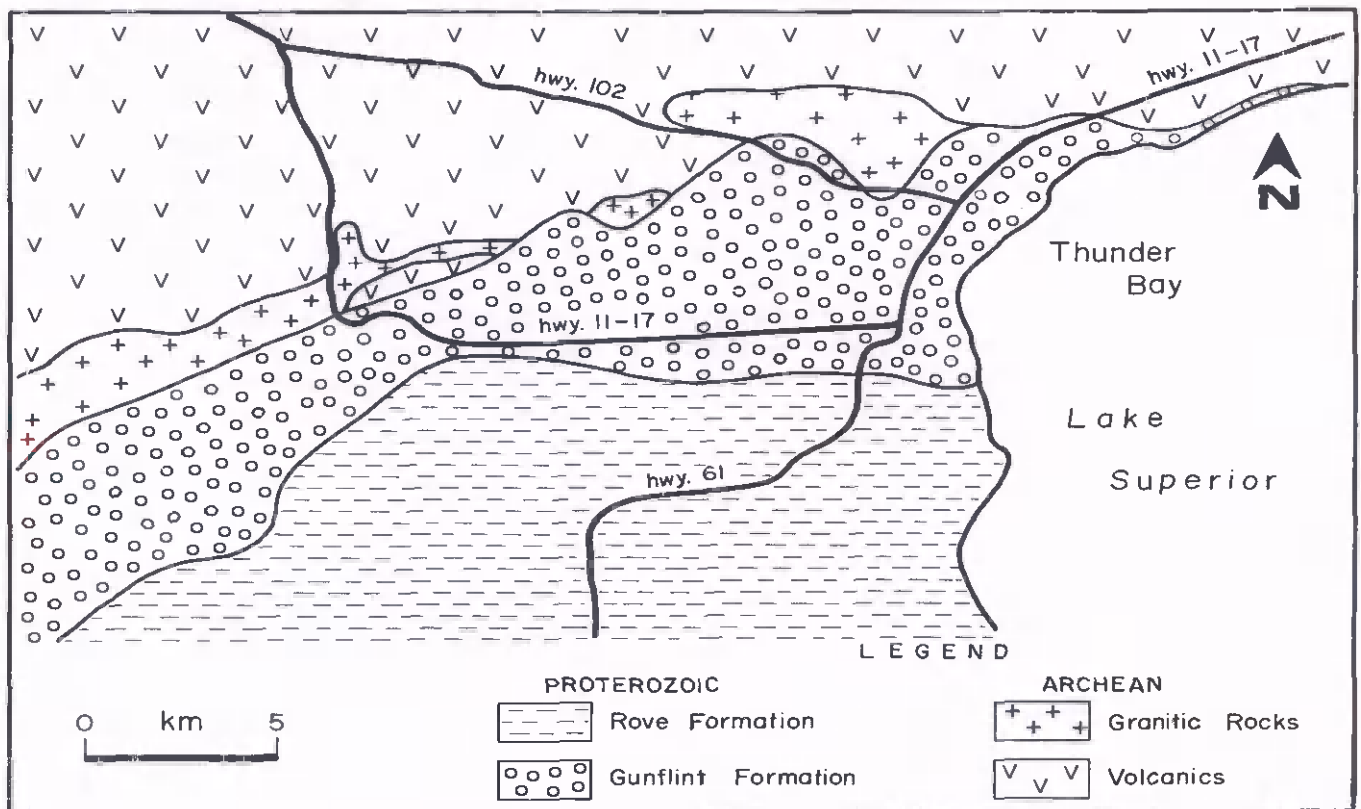


Figure 13. Geology of the Thunder Bay area.

STOP 1 Proceed west on Highway 11/17 (the Transcanada) (Figure 13) past Kakabeka Falls and over the Kaministiquia River. A few hundred metres past this bridge there is a turn off to Nolalu. Take it and proceed approximately 100 metres, the main outcrop is on the right.

Chemical sediments of the Gunflint Formation rest directly on Archean granitoids at this location. This highlights the sporadic development of the thin conglomeratic veneer comprising the Kakabeka Conglomerate in this area. Stromatolites, including some large forms, are developed in these basal cherty carbonates. The stromatolitic ribbon carbonates are abruptly overlain by a grainstone succession.

STOP 2 Proceed back to the Transcanada Highway turn right and cross the Kaministiquia Bridge. Turn left on road immediately after bridge and drive through park to end of road. A path leads to the first set of rapids above the bridge.

The rapids are formed by Archean granitoids. The slow-water area to the south is underlain by the Gunflint Formation. River level will govern what can be observed at this stop. The Kakabeka Conglomerate is patchily developed overlying the Archean at this stop. Silicified stromatolites are developed on the Conglomerate or directly on the granitic basement. This is the location from which samples collected in the 1950's yielded the first Gunflint cyanobacteria described in the literature. The samples were from the silicified stromatolites and were described by Tyler and Barghoorn (1954).

STOP 3 Proceed back to the Transcanada Highway and turn left. Turn right into the main park entrance and park in the parking lot. Walk to the observation platform.

The river gorge is composed of a sequence of volcanoclastic shales (less resistant, darker units) and tuffs (more resistant, lighter coloured units). This sequence represents the major volcanoclastic horizon present in the upper Gunflint Formation and is traceable to the south through the Mesabi Range. Basalts outcropping approximately 30 kilometres to the southwest are probably correlatable with this unit. Unpublished geochemistry (major, trace, REE analysis) performed on samples of the basalt indicate a tholeiitic composition with a deep mantle source (plume).

STOP 4 Walk back to the southeast side of the parking lot.

An outcrop of ribbon chert-carbonate (ankerite) forms this cut. Folds and faults are present and probably represent soft sediment deformation. Commonly, past studies of the Gunflint Formation have described this type of sediment as forming in a deep, quiet water environment and this may be correct. However, similar carbonate sequences are often attributed to deposition in intertidal or lagoonal subtidal environments. The evidence supporting a deep or a shallow setting will be discussed at this stop.

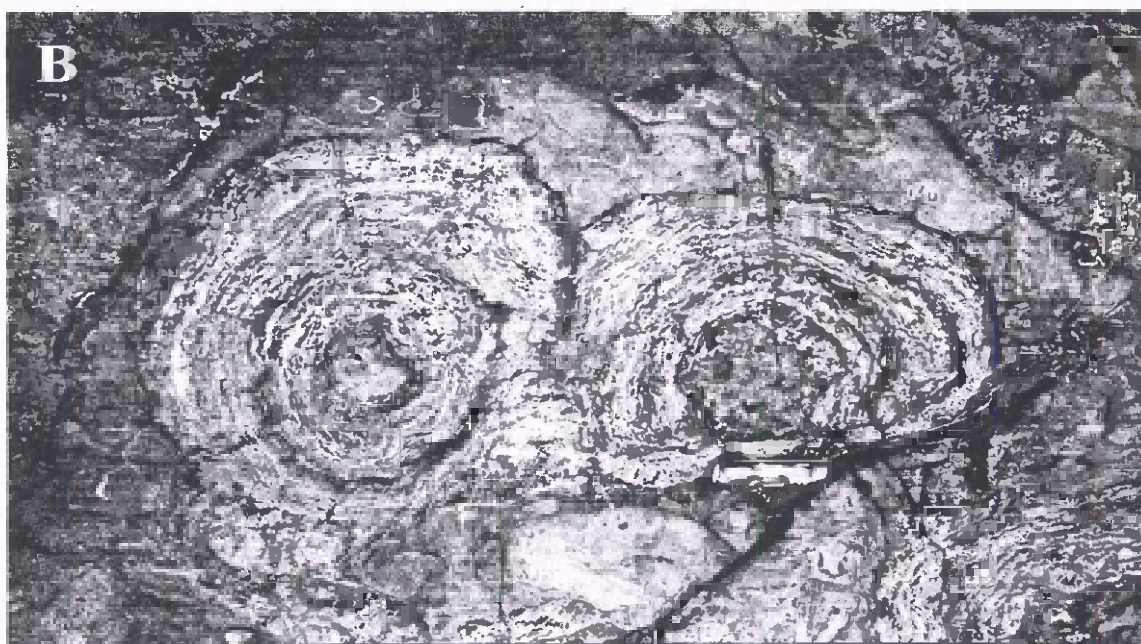


Figure 14. *Stop 6. A) Intraformational beachrock conglomerate formed by storm wave rip-up of partially cemented surficial layers of medium-grained sand in the upper beach environment. B) Plan-view of cabbage-sized stromatolites with medium-grained and granule sand layers banked up against them.*

STOP 5 Turn right onto the Transcanada and turn left onto Oliver Road. Turn left on Mapleward Road and right on Dawson Road. Drive 7.4 km to a small dirt pull off on the left and park. Walk down the path for a few minutes until you see a small, rounded outcrop.

This is an outcrop of pebbly sandstone belonging to the Kakabeka Conglomerate. It overlies Archean granitic rocks and represents gris (angular, granitic debris), probably weathered from surrounding basement.

STOP 6 Turn left on Dawson Road and proceed to Thunder Bay. Turn right on the Transcanada and left on Oliver Road. Continue down Oliver Road Turn right at Hill Street and stop at the corner with Queen Street. This is private property so stay on the outcrop.

This is a no hammer stop. The middle stromatolitic unit forming the base of the upper member of the Gunflint Formation outcrops at this location. The carbonate stromatolites are excellently preserved and are visible in both cross-section and plan view (Figure 14). Carbonate sands are banked up against the cabbage-sized forms and eventually overwhelm them. The stromatolites established themselves on a silicified horizon which at other locations shows evidence of subaerial exposure. This unit provided the necessary hard substrate to allow the stromatolites to develop. The stromatolitic horizon is overlain by beachrock which has been disrupted by a major storm event. Pieces of the underlying stromatolites have been ripped up along with the overlying partially cemented carbonate sands. The beachrock formed in the upper beach environment due to salt spray from breaking waves depositing meniscus cements at grain contacts of the sand. This partially cemented sand layer on the upper beach was disrupted into slabs when a large storm caused waves to break in this environment.

STOP 7 Proceed back to Oliver Road. Turn right. Turn left on Crown Street (at the junction with John Street) and drive to near where it bends right and turns into Dixon Street. Drive one block northwest on a small lane to the base of the High Street Park hill.

The outcrop forming the base of the hill is composed of carbonate breccia with some chert clasts. Wavy laminated dripstone forms discontinuous layers blanketing some of the breccia piles. This is a collapse breccia formed by failure of the roof of a cavern in the ankerite. A number of failures occurred as denoted by the interlaminated dripstones, which were deposited on debris piles on the cavern floor by roof leakage. This unit forms the upper layer of the Gunflint Formation's lower member and is often overlain by stromatolites correlatable with STOP 6. Maximum regression of the Animikie sea occurs at this interval leading to subaerial exposure of Gunflint sediments in the Thunder Bay region. Evidence of subaerial exposure has not been observed in drill core examined from south of the U.S. border.

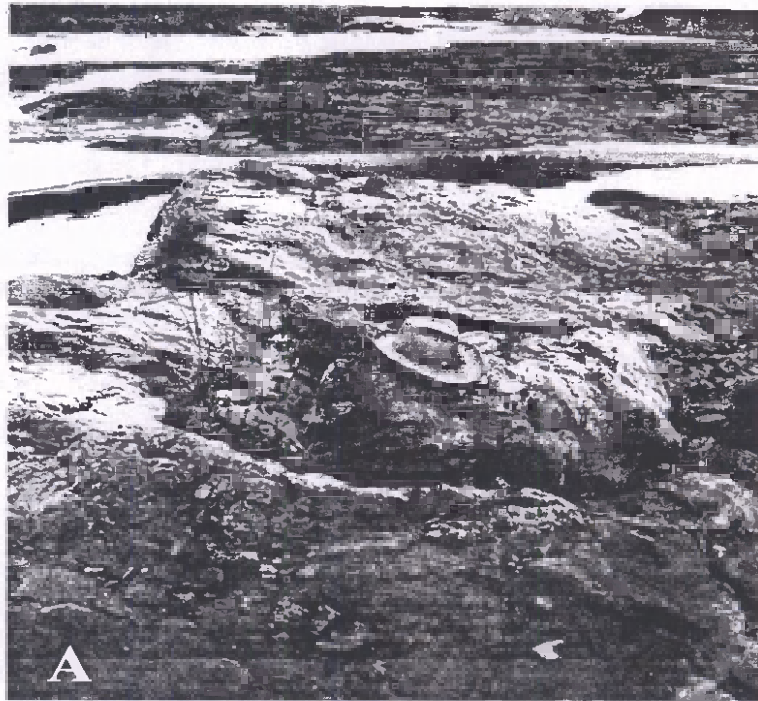


Figure 15. Stop 8. A) Silicified stromatolitic reef overlain by black volcaniclastic shales (background). B) Hummocky bedding plane of ankeritic grainstones.



STOP 8 Drive to Oliver Road and turn left. Proceed to its end and turn left onto Cumberland. Continue driving on Cumberland to Current River. Turn left onto Arundel Street and proceed to the cement bridge across the Current River. Immediately before crossing the bridge turn left onto Lyon Blvd. Pull off to the right into one of the parking areas. Walk to the river then down river to the end of the outcrop.

You are standing on the Paleoproterozoic sea floor. The uneven topography of the ankerite grainstones reflects the presence of dunes and hummocks produced by the storm events which moved the carbonate sands over the bottom (Figure 15). Silicified colonies of stromatolites dot this irregular seascape. This forms the lower unit of a tripartite assemblage. Large silicified stromatolites occur on the other side of the bridge and directly overlying these is a volcanoclastic silty shale. A description of this sequence and its inferred depositional environment follows (taken from Fralick, 1989).

Major Subdivisions

Coalesced domal stromatolites form the microbial bioherms present in the outcrops on Current River. They represent a portion of the upper stromatolitic zone and are underlain by parallel laminated and hummocky cross-stratified carbonate grainstone which has undergone a complicated history of silicification, neomorphic spar growth and dolomitization-ankeritization. The bioherms are draped by graded, centimetre-scale beds of dark grey siltstone and shale (Figure 16).

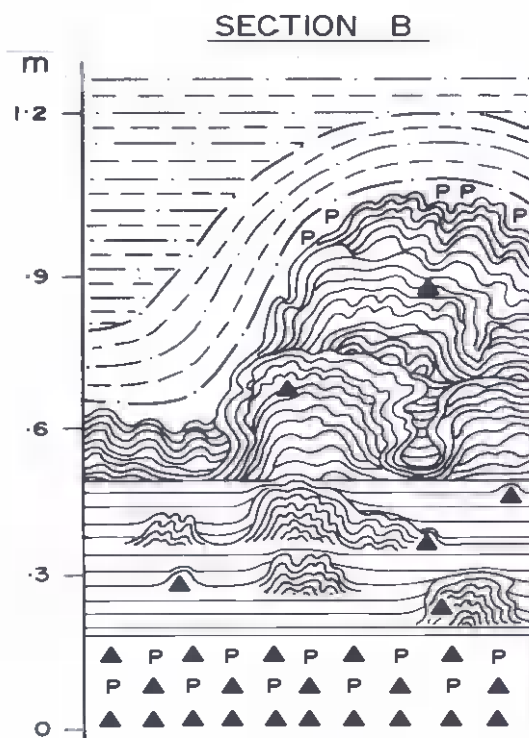
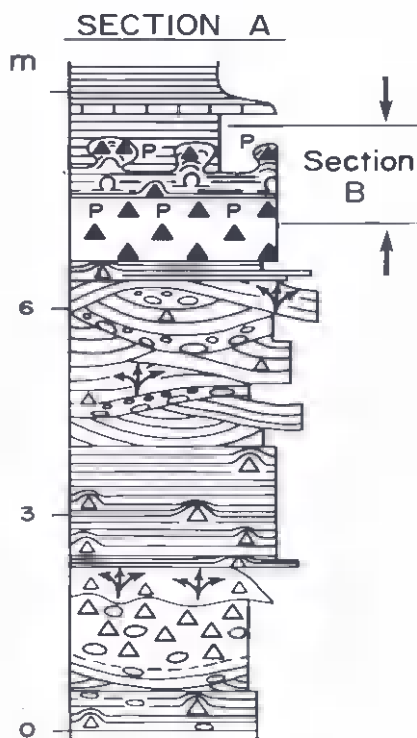
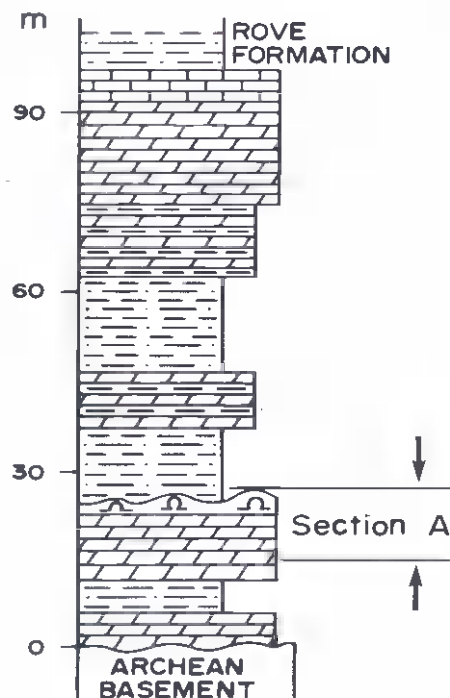
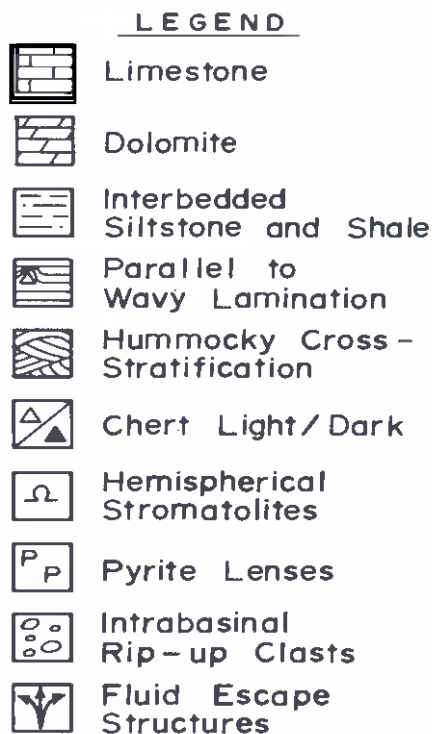
Carbonate Grainstone

The grainstone is divisible into three types: 1) hummocky cross-stratified; 2) parallel laminated; and 3) massive, highly silicified. Hummocky cross-stratification generally has wavelengths, between 0.5 and 1 m. It commonly contains oval and rectangular rip-up clasts which, when abundant, fine upwards in individual beds. Often silicified rip-ups or mixtures of silicified and non-silicified rip-ups are present in non-silicified beds.

Parallel laminated grainstone also contains rip-up clasts, but they are less abundant. The appearance of these units is complicated by the presence of silicified nodules. The grainstone develops a wavy lamination where the parallel laminae bend around silicified areas. The compaction deformed laminae are similar in appearance to hummocky cross-silicification, but can be distinguished by the presence of low-angle erosive truncations in the latter.

Silicification of layers and patches within layers is very common. One bed near the base of the measured section exhibits a marked upward increase in silicification, culminating, at its top in a layer of massive chert. Another highly silicified zone directly underlies the stromatolite bearing horizon. Silicified laminae present in other beds have sharp, flat top surfaces and sharp irregular bottom surfaces. Water escape structures, common throughout the hummocky cross-stratified units, sometimes brittlely deform these silicified layers, forming cross-cutting carbonate dikes which rotate plug-like, angular chert blocks to one side.

Figure 16. Stratigraphic sections at varying scales of the stromatolitic sequence on Current River.



Microbial Bioherms

The base of the bioherm bearing layers is composed of parallel laminated carbonate grainstone, wavy silicified in patches, containing isolated centimetre to decimetre-scale (maximum 20 cm high x 50 cm wide) domal stromatolites. They represent the SH-C/LLH-C type of Logan *et al.* (1964). Grainstone laminae are banked up against the sides of the microbial mounds. Some mounds are buried by the grainstone while others, after being partially covered, re-established themselves. The amount of grainstone decreases upward and the cabbage sized stromatolites begin laterally merging. This trend resulted in the development of up to 1 m high, sometimes overlapping, microbial mounds scattered over the bottom. In areas where early stromatolites flourished raised platforms were built (up to 0.5 m above the surrounding bottom) on which large, coalescing, domal mounds grew, giving a total relief above the adjacent area of well over 1 m.

All stromatolites have been extensively silicified and internal structures are poorly preserved. Lamination in the stromatolites, when visible, is caused by the alternation of light and dark chert bands.

Pyrite layers and irregular blobs are present in the mounds, in places forming up to 20% of the rock. A stratiform, discontinuous pyrite layer also drapes some mounds; the larger ones. Lenses in this unit attain thicknesses up to 10 cm.

Siliciclastics

Siliciclastic, fine grained rocks drape the microbial bioherms. Centimetre-scale graded beds comprising this unit are internally composed of alternating matrix-rich and matrix-poor, millimetre-scale laminae. The matrix-rich layers contain fine grained, sand-sized clasts of very angular quartz and orthoclase floating in a matrix composed of very fine grained dark material containing disseminated pyrite and 1 to 2% non-carbonate carbon. Matrix-poor layers are composed of very angular quartz, orthoclase, biotite and muscovite in clast support (maximum size 0.1 mm). The thinly bedded rocks are overlain by an indistinctly laminated unit which fills in hollows between the stromatolite mounds. It is a poorly sorted, matrix supported silty shale with bedding parallel muscovite flakes up to 5 mm in length. Shallow, saucer shaped scours occur in its upper portion. The scours are overlain by carbonate grainstone laminae up to 1 cm thick. The grainstone unit is 10 cm in thickness and is interbedded with and overlain by well laminated silty shale similar to the lower unit.

DEPOSITIONAL ENVIRONMENT AND HISTORY

Stage A - Storm Dominated Shelf

The presence of hummocky cross-stratified carbonate grainstone with the absence of wave-ripple reworked tops indicates that the sand blankets were deposited between fair weather wave-base and storm wave-base in a relatively siliciclastic-free environment (Figure 17). The carbonate, or its precursor was likely precipitated, eroded and abraded in shallow water areas and transported into deeper water during storm events. Silicified rip-up clasts and crusts denote that silicification was syndepositional (as proposed by Simonson, 1986); the degree of silicification increasing towards the top of the bed indicates that seawater (*sensu lato*) was probably the silicifying agent.

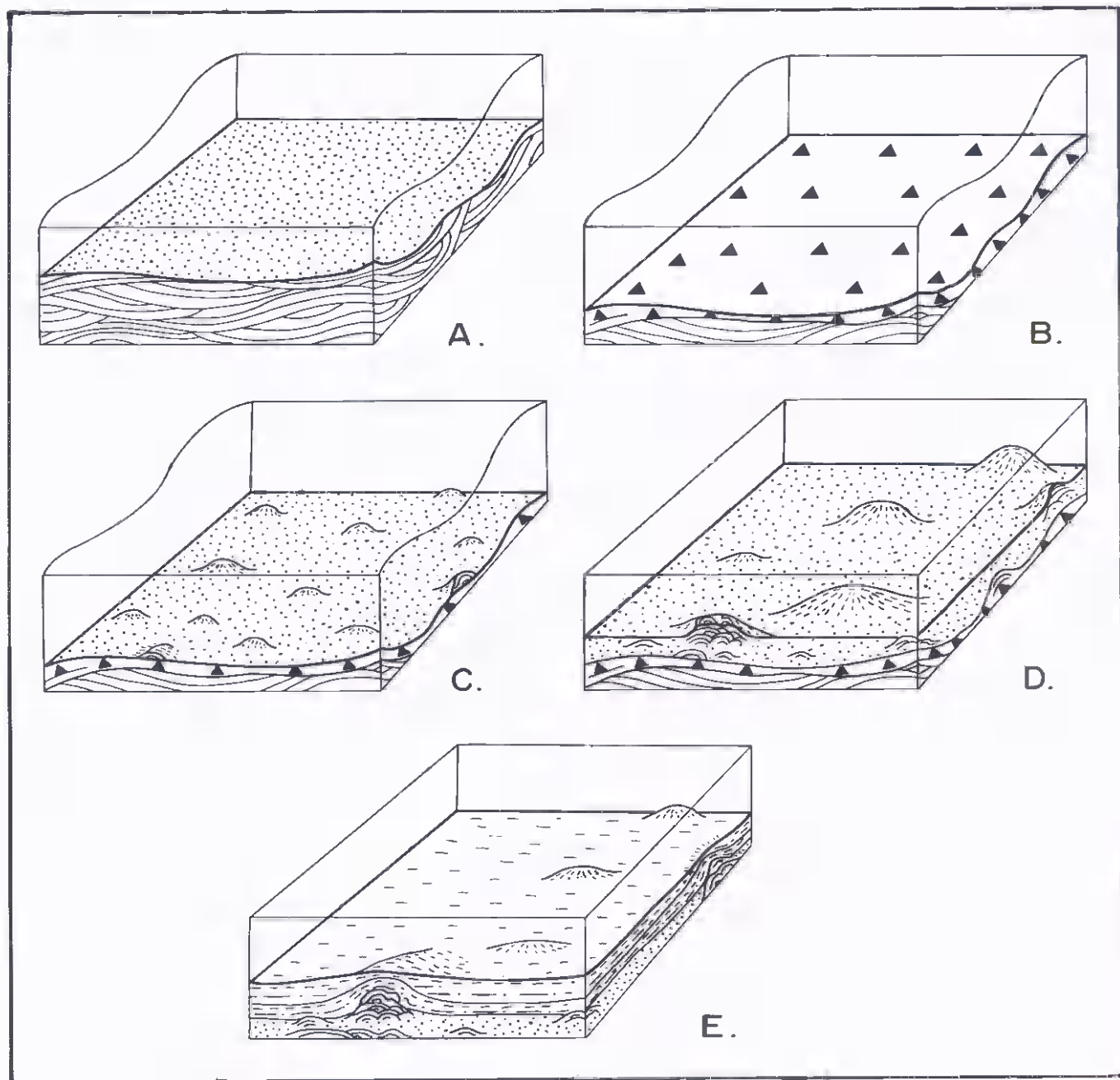


Figure 17. *Developmental phases of the environment represented by sediments present at Stop 8. Each stage is discussed in the text.*

Stage B - Stabilization and Silicification of the Substrate

If seawater was the agent causing alteration of the carbonates then silicification should increase with length of time the unit was positioned close to the water-sediment interface. This scenario implies that the massive chert underlying the stromatolites formed by alteration of carbonates during an interval of lowered deposition rate. This is best explained by increasing the water depth so that the bottom sinks below storm wave-base and slower deposition rates ensue.

Stage C - Stromatolite Development

Stromatolites grew on the silicified, stabilized substrate, forming inverted-bowl shaped mounds on the bottom. The lack of wave agitation and firm substrate proved ideal. Some stromatolites were occasionally buried by grainstone laminae brought in by seaward flowing bottom currents generated during storm events.

Stage D - Large Microbial Bioherm Growth

Individual stromatolite colonies coalesced upwards, forming large mound shaped structures and creating an irregular bottom topography. Little sediment accumulated between the stromatolite mounds as storm events became less capable of supplying detritus to the area.

State E - Volcaniclastic Burial

Growth of the bioherms was abruptly terminated by the deposition of a series of volcaniclastic sediment blankets. The draping of these blankets over the mounds rather than their ponding in intermound areas indicates rain out from above rather than sedimentation from bottom currents.

The composition (fresh detrital biotite), angularity (subhedral feldspars) and sorting (large well formed crystals floating in a dark fine grained groundmass) all point to little or no reworking of pyroclastic material. The layers may have been deposited directly as rain out from ash landing on the water's surface or as storm deposited layers. The latter is preferred, as the microlaminations can then be explained by fluctuations in the energy level. The presence of significant quantities of non-carbonate carbon in the groundmass of these rapidly deposited units may reflect the settling of the volcaniclastics through a sea with a composition similar to microbial vegetable soup. The occasional large storm event eroded saucer shaped scours into the fine grained material and transported minor amounts of carbonate in from the shore-proximal zone.

Lateral Implications

As water depth increased, the bottom changed from a wave dominated condition to a zone of slow sedimentation and thus pervasive silicification. This led to stabilization of the sediment and stromatolites began to develop. With time (increasing water depth) large domal structures grew. This succession was brought to an end in the upper stromatolitic zone of the Gunflint Formation by pyroclastic eruptions and volcaniclastic deposition drastically altering bottom conditions.

STOP 9 Turn left on Lyon Blvd. Then right on Arundel. Turn left on Hodder Ave and drive to the Transcanada. Turn right on the Transcanada and drive to the Pass Lake turnoff. Turn right here and proceed till you encounter a large outcrop of contorted ribbon chert-carbonate on your left.

Figure 18. A) Stop 9. Disrupted ribbon chert-carbonate layers. B) Stop 11. Chert-carbonate silt to pebbly grainstone with rip-ups of volcanoclastic sediment and fingernail-sized shards.

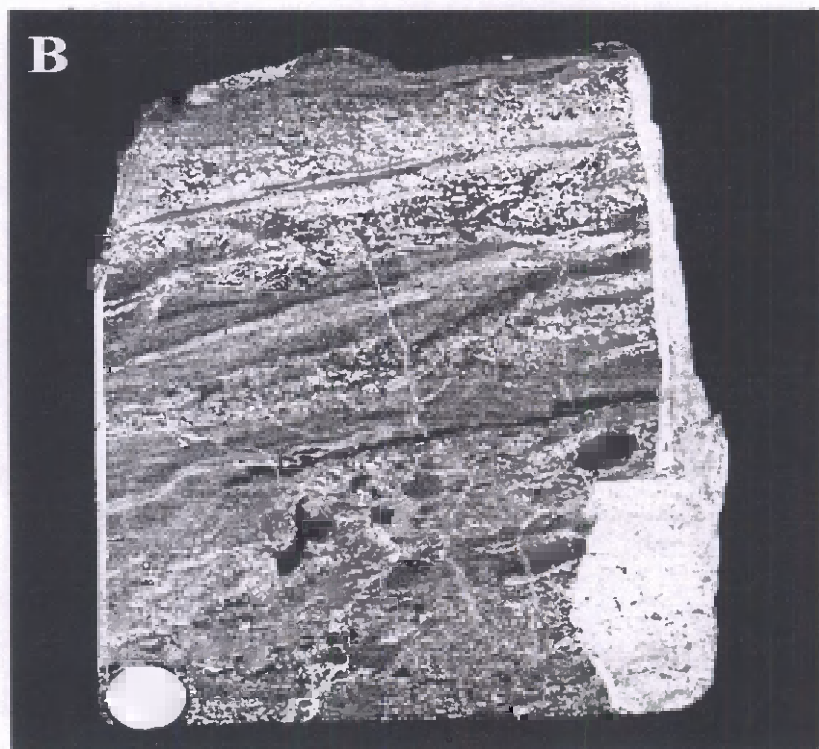
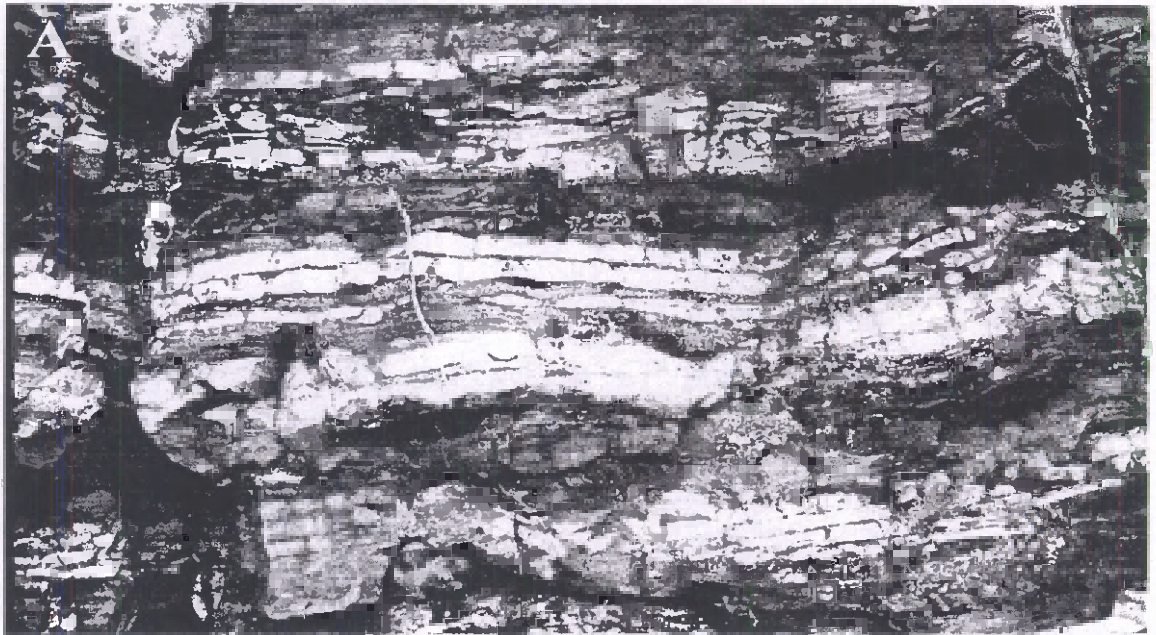




Figure 19. *A) Close-up of internal laminations in a medium-grained sandstone storm layer. The bottom one third of the layer contains upper flow regime parallel laminations. The upper two thirds is hummocky cross-stratified with low angle truncation surfaces clearly visible. B) Outcrops at Stop 12.*

Laterally continuous interlaminated chert and ankerite layers occur in this outcrop (Figure 18A). These layers are intensely folded and disrupted in places. The outcrop provides an excellent opportunity to discuss if the deformation is synsedimentary, why the ribbon chert-carbonates preferentially exhibit this type of deformation and what type of depositional environment is represented by these sediments.

STOP 10

Proceed back to the Transcanada and turn left, towards Thunder Bay. Drive approximately 1.8km until you reach the small side road on the right. Turn right onto the dirt road and park.

Volcaniclastic, very fine-grained sandstones, siltstones and mudstones occur at this location. These are probably correlatable with the interval examined in the gorge at Kakabeka Falls. The very fine-grained sandstones and coarse-grained siltstones are ripple laminated with well preserved internal structure (see polished hand samples). Again the possible depositional processes which may have operated to form this unit will be discussed in the field.

STOP 11

Turn right onto the Transcanada and proceed towards Thunder Bay. Turn right at the access road to the Terry Fox lookout. Park in the parking lot and walk to the large outcrop on the highway; watch out for cars.

A large diabase sill overlies the Gunflint Formation at this location. The Gunflint grainstones present here contain large rip-up fragments and both small and large sherds of volcanogenic origin (Figure 18B). Volcaniclastic shales are present near the top of the sedimentary sequence and this has been described by some workers as the Gunflint-Rove contact. However, interbedding of volcaniclastic shales and chemical sediment is common in the upper portions of the Gunflint Formation, and if more chemical sediment occurs at a higher level this cannot be defined as the contact.

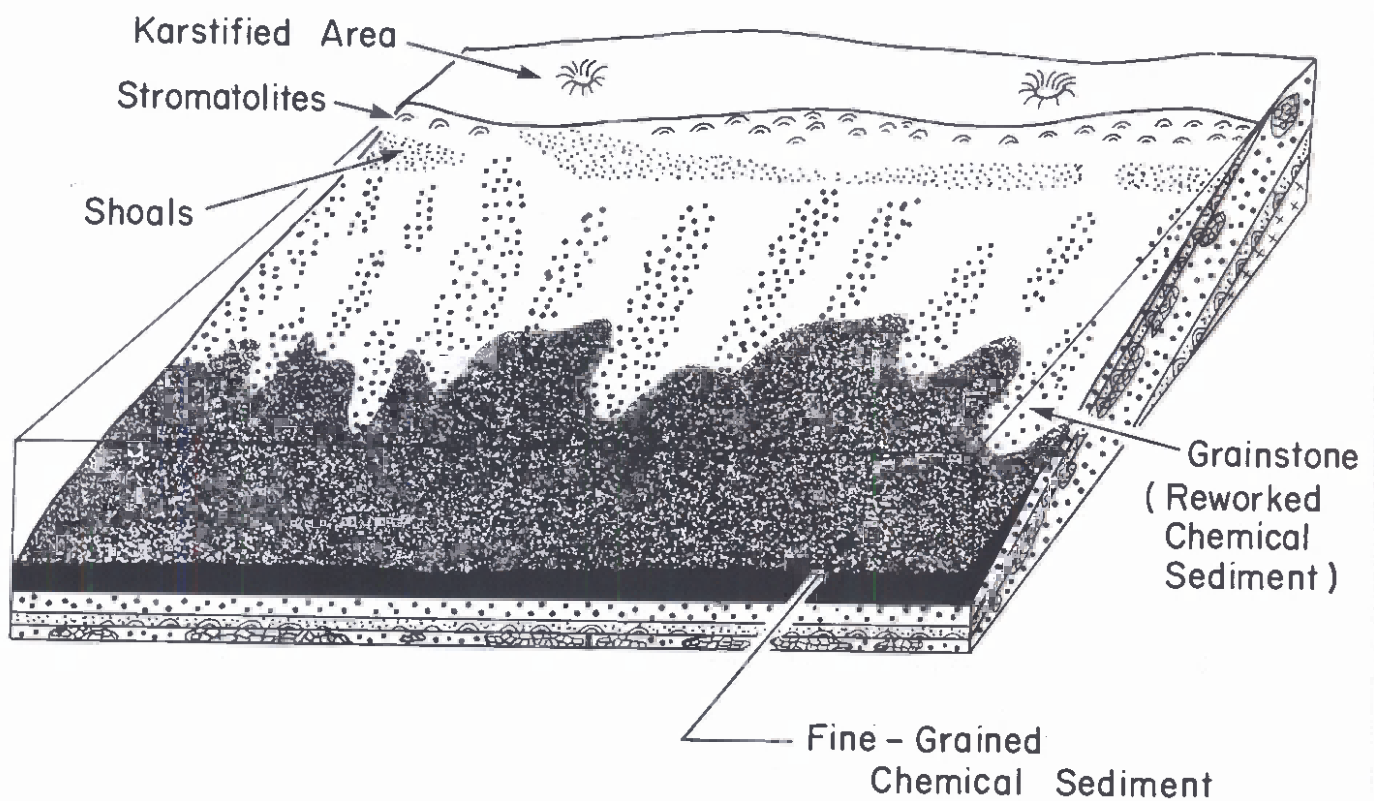
STOP 12

Turn right on the Transcanada and drive across the bridge over the Current River. Proceed up the hill on the other side. A rock cut occurs at the top of the hill and there is a dirt pull-in. Park here.

This grainstone sequence is similar to those examined on the Current River (Figure 19). There may be metre-scale thinning-upwards cycles present here but they are indistinct. Thin, discontinuous chert layers occur at the tops of these possible cycles. In places this chert is very carbon-rich, possibly denoting a slowing of the deposition rate allowing accumulation of organic carbon. This outcrop provides an opportunity to discuss some of the possible process which could drive deposition of this type of asymmetric cycle (Figure 20).

Turn right on the Transcanada and proceed to Oliver Road. Turn left on Oliver Road and turn right at the road leading to the main entrance of Lakehead University.

Figure 20. Generalized depositional model for nonbarred portions of the strand-proximal Gunflint shelf during deposition of upper Gunflint chemical sediments.



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GLACIAL HISTORY AND REGIONAL TILL SAMPLING IN THE ARCHEAN SHEBANDOWAN GREENSTONE BELT

Andy Bajc, Ontario Geological Survey

The Shebandowan Greenstone Belt, host to a variety of precious and base metal deposits, is a bedrock-dominated terrain, typically having thin till cover. Till sampling techniques and resultant gold grain and other heavy mineral results will be examined in an exploration context.

INTRODUCTION

The Shebandowan greenstone belt, which is located immediately north and west of the city of Thunder Bay, has a rich history of mineral exploration and mining spanning well over 100 years. The belt is host to a variety of precious and base metal mineral deposit types. Past producers include the Inco Limited Shebandowan Ni-Cu-PGE mine, the North Coldstream Cu-Au-Ag mine and the Ardeen Au mine which was northern Ontario's first gold producer. Exploration activity within the belt has currently subsided, largely in response to the depressed prices of precious and base metals and the inability of junior mining companies to raise funds for exploration.

In response to this recent decline in mining and mineral exploration activity within the belt, the Ontario Geological Survey initiated a 2 year program of Quaternary geology and drift geochemistry to further evaluate the mineral resource potential of the Shebandowan belt. This was accomplished by detailed (1:50 000 scale) Quaternary geology mapping and regional humus and c-horizon till sampling. Quaternary mapping was undertaken to document the distribution and character of the various Quaternary deposits within the study area and to reconstruct the ice flow history associated with these deposits. This information is essential for the successful implementation of mineral exploration programs utilizing drift prospecting techniques and surficial geochemistry.

During the 1998 field season, Quaternary mapping and regional humus and c-horizon till sampling were conducted over the eastern half of the belt within the area covered by the Sunshine (52A/12) and Kakabeka Falls (52A/5) 1:50 000 scale National Topographic System (NTS) map sheets (Figure 1). The results of this work are summarized in Ontario Geological Survey Open File Report 5993 (Bajc 1999a). Mapping and sampling was extended westward into the Shebandowan (52B/9) and Burchell Lake (52B/10) 1:50 000 scale NTS map areas during the 1999 field season. An OGS preliminary map containing information on the regional distribution of gold grains in till within this area was released during the fall of 1999 (Bajc 1999b). The complete mineralogical and geochemical datasets for the western half of the belt were released in the spring of 2000 (Bajc 2000).

BEDROCK GEOLOGY

A comprehensive summary of the bedrock geology of the study area is contained in papers by Sutcliffe (1991), Williams (1991) and Williams *et al.* (1991) as well as in numerous Ontario Geological Survey maps and reports. The study area contains 3 distinct bedrock domains (Figure 2). The centrally-located Wawa Subprovince, within which the Shebandowan greenstone belt occurs, is fault-bounded to the north by metasedimentary and felsic intrusive rocks of the Quetico Subprovince and unconformably overlain to the south by Paleoproterozoic metasedimentary rocks of the Animikie Group (Gunflint and Rove formations).

Metasedimentary rocks of the Quetico Subprovince consist primarily of turbiditic wackes, arkoses and quartz arenites and their associated paragneisses and migmatites. Post- to syndepositional felsic plutons consisting of feldspar-megacrystic granite, granodiorite to tonalite and monzonite comprise a notable proportion of this domain. Narrow, elongate lenses of tholeiitic,

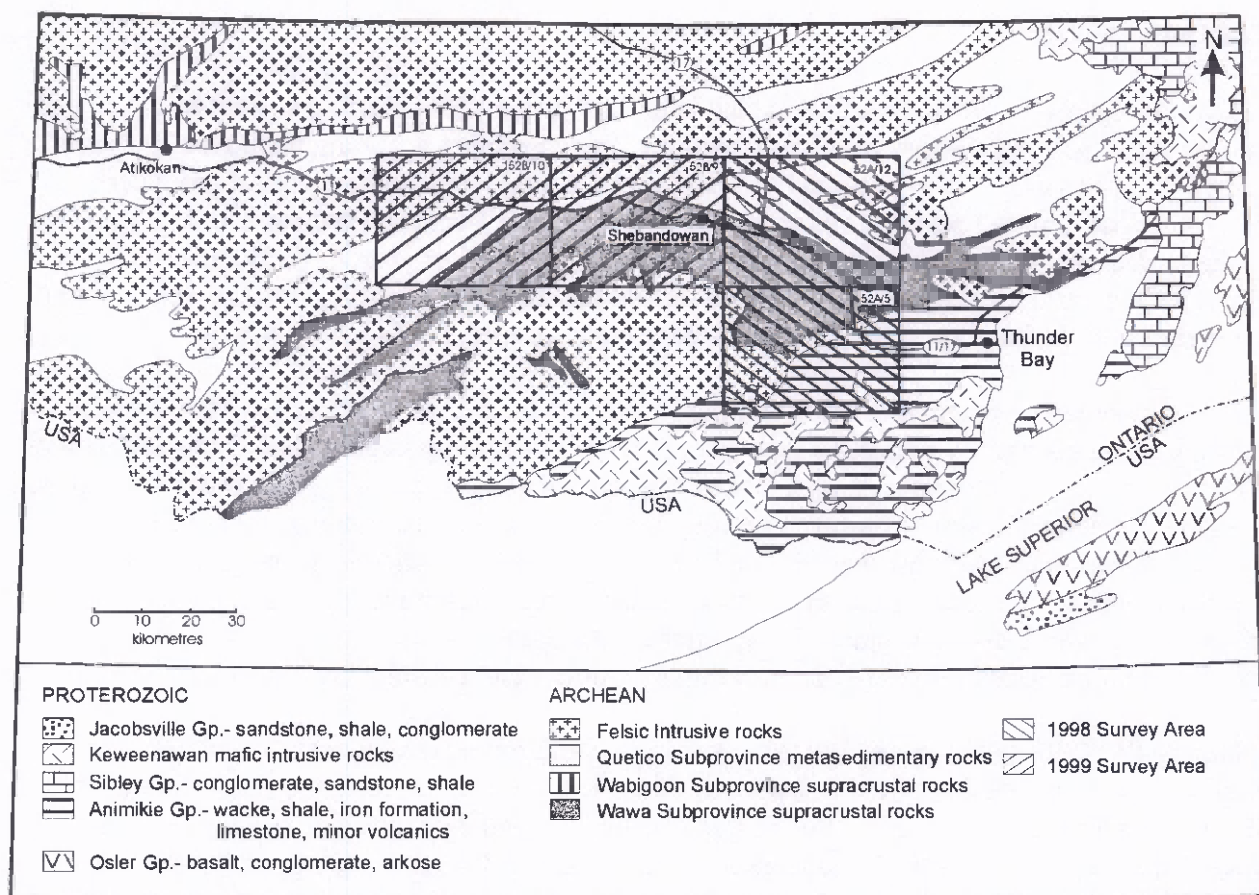


Figure 1. Location map of study area including regional bedrock geology.

mafic metavolcanic rocks and associated gabbroic rocks occur locally within the Quetico Subprovince (Brown 1995, Osmani 1997).

Metasedimentary rocks of the Gunflint and Rove formations (Animikie Group) were deposited within a large basin that extends into Minnesota, Wisconsin and Michigan (Sutcliffe 1991). The Gunflint Formation contains a basal unit of conglomerates that rest unconformably on the Archean basement and are overlain by interbedded argillites, argillite tuffs, cherts, algal cherts, jasper, carbonates, ferruginous carbonates, hematite, magnetite taconite and silicate taconite. The overlying Rove Formation consists of black, locally pyritic shales and grades upward into interbedded black shale and arkosic greywackes. Mesoproterozoic "Logan" diabase sills intrude the Rove Formation and form the resistant cap rocks of large mesas in the southern part of the map area. Silver mineralization occurs locally along normal faults that intersect the contact between Rove Formation shales and Logan diabase sills. Notable silver deposits occur in the southeastern corner of the Kakabeka Falls NTS map area, in the vicinity of Badger and Beaver mountains (Creswel or Rabbit Mountain group of deposits) and Silver Mountain. Accessory minerals in these deposits include: acanthite; pyrite; sphalerite; marcasite; galena; calcite; quartz; barite; and fluorite (Franklin *et al.* 1986).

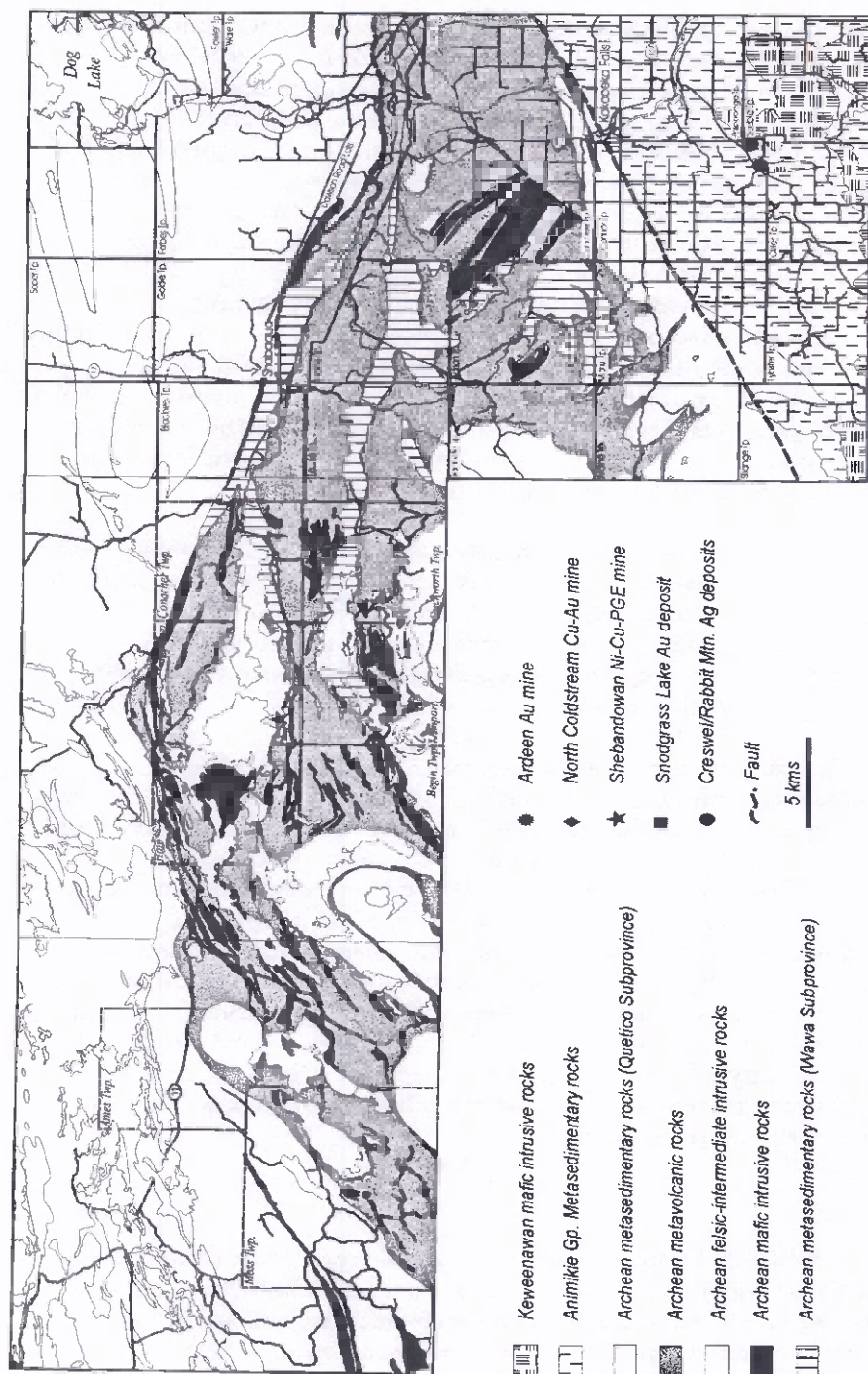


Figure 2. Bedrock geology of the study area.

Supracrustal rocks of the Shebandowan greenstone belt can be subdivided into 2 contrasting packages of metavolcanic and metasedimentary rocks: 1) an older suite of mafic to felsic metavolcanic rocks belonging to the Greenwater assemblage; and 2) a younger suite of metasedimentary and metavolcanic rocks belonging to the Shebandowan assemblage. The Greenwater assemblage consists primarily of tholeiitic basalts with intercalated intermediate to felsic metavolcanic rocks and minor komatiites (Corfu and Stott 1998). Mineralization at the Shebandowan, North Coldstream and Ardeen mines are all contained within Greenwater assemblage rocks. Ni-Cu-Pt-Co mineralization at the Shebandowan mine is hosted by a gabbro-peridotite sill. Cu-Au-Ag mineralization at the North Coldstream mine is hosted by a highly silicified gabbro. Gold mineralization at the Ardeen mine occurs in composite quartz veins developed along the Pele and Ardeen fault systems. Advanced exploration in the form of a decline ramp and bulk sample was undertaken at the Snodgrass Lake Au prospect in central Moss Township during the mid-1980s. Gold mineralization occurs in sheared felsic metavolcanic rocks and in sheared and fractured diorite, gabbro and quartz-feldspar porphyries. The Haines gabbro-anorthosite complex represents one of the larger mafic intrusions within the belt. The intrusion, which is crudely layered, is host to a number of Ni-Cu-PGE occurrences.

The younger, Shebandowan assemblage both unconformably overlies (i.e., fault contact) and intrudes the Greenwater assemblage. It consists of alkalic metavolcanic and intrusive rocks (tuff, breccia, syenite, lamprophyre, quartz and feldspar porphyry, granodiorite and diorite) as well as metasedimentary rocks (arkoses, wackes, conglomerates and oxide-facies iron formation) (Schneiders *et al.* 1998). This assemblage is interpreted as representing an intracratonic basin assemblage similar to the Timiskaming assemblage of the Abitibi subprovince.

Most gold occurrences in the eastern portion of the Shebandowan greenstone belt display a close spatial association with rocks of the Shebandowan assemblage. The style of gold mineralization is not unlike that which occurs in such prolific mining camps as Timmins and Kirkland Lake. Gold mineralization is commonly associated with both large scale and subsidiary structures and structurally-controlled intrusions. Mineralization occurs in sheeted and *en echelon* quartz veins, stockwork zones and vein breccias as well as in stocks, dikes and sills of felsic to mafic composition. Gold is also commonly associated with: sulphide-rich, quartz-veined replacement zones in oxide facies iron formation; silica-flooded, hydrothermally-altered zones; and iron-carbonate and pyrite-rich zones commonly associated with shear and permeable zones. Associated alteration and accessory minerals include: iron carbonate; silica; potassic feldspar; sericite; hematite; chlorite; calcite; pyrite; arsenopyrite; tourmaline; and green mica (Schneiders *et al.* 1998). Notable exploration programs within the current study area are highlighted in the last 5 annual volumes of the Resident Geologist's Report of Activities.

GLACIAL GEOLOGY

The study area is characterized by several contrasting physiographic regions, each defined by a unique suite of overburden conditions. Most of the Burchell Lake and Shebandowan NTS map areas as well as parts of the Sunshine and Kakabeka Falls NTS map areas are characterized as bedrock-dominated terrains. Local accumulations of thick drift occur in morainic belts, glaciofluvial complexes and narrow, structurally-controlled depressions within these uplands.

Thicker deposits of till and/or glaciofluvial and glaciolacustrine sediments occur within 2 isolated basins (Figure 3). A northern lowland, here referred to as the "Kaministikwia basin", occupies the valleys of the Oskondaga, Matawin, Shebandowan and Kaministikwia rivers. A southern basin is centred on the valleys of the Whitefish and Kaministikwia rivers. In the northern basin, thick deposits (several tens of metres) of glaciofluvial and glaciolacustrine sediments drape a rugged Archean bedrock surface producing a gently rolling landscape with sporadic bedrock outcrops. The low relief observed within the southern basin is a reflection of not only the flat-lying Proterozoic sediments that underlie the region, but the extremely thick glaciolacustrine deposits that have infilled deep valleys along the Whitefish and Kaministikwia rivers. Overburden thicknesses along these valleys exceed 60 m in places.

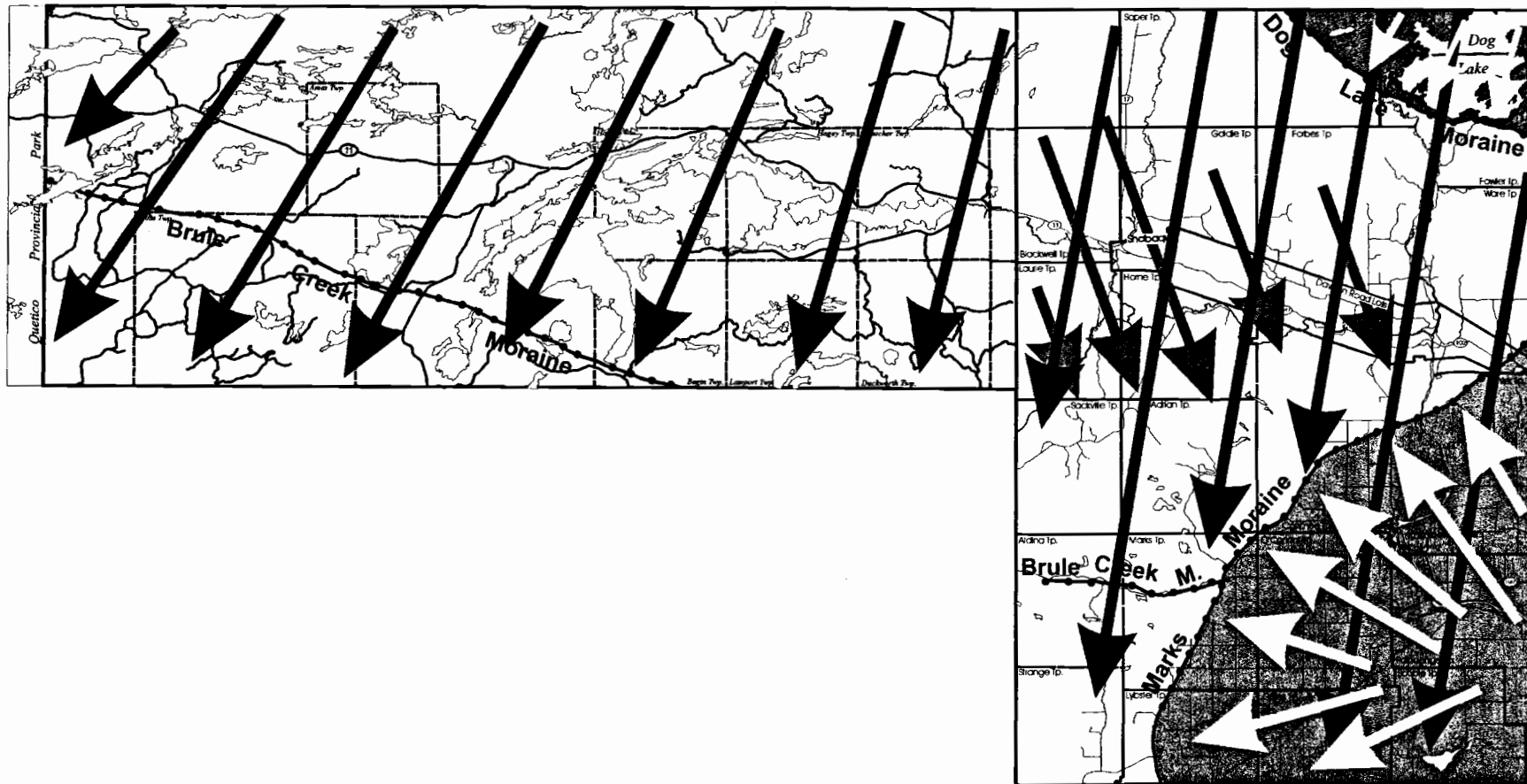
The erosional and depositional features preserved within the study area record a complex history of ice flow events associated with the Wisconsin glacialiation (Figure 4). The entire study area was initially affected by southward flowing "northern" ice. Striae associated with this event generally range between 190° and 220° Az. In the southern half of the Sunshine map area, there is widespread evidence for an older ice flow event towards 160° Az. The age and significance of this event is not known.

Pauses in the general retreat of the "northern" ice lobe resulted in the formation of large, arcuate, recessional moraines in northwestern Ontario. The Brule Creek Moraine, which traverses Aldina and Marks townships in the west-central portion of the Kakabeka Falls NTS map area, represents one of these stillstand positions (Figure 4). This moraine has been traced westward into the Shebandowan and Burchell Lake map areas where it is defined by a discontinuous belt of morainic ridges composed of till and associated ice-contact stratified deposits. West of the study area, the Brule Creek Moraine has been correlated with the Eagle Finlayson Moraine (Zoltai 1965). A large ice-contact delta built off of this moraine along the south shore of Burchell Lake is graded to an upper level of glacial Lake Agassiz. The upper topset surface of the delta occurs at an elevation of 457 to 472 m asl (1500-1550 feet asl). To the east of Marks Township, the Brule Creek Moraine has been overridden and obscured by a younger glacial advance (Figure 4). Isolated buried deposits of ice-contact stratified glaciofluvial deposits of northern provenance and possibly associated with the Brule Creek Moraine occur in south-central Conmee and west-central Oliver townships.

A second major moraine of northern affinity (Dog Lake Moraine) is situated in the northeastern corner of the Sunshine map area in the vicinity of Dog Lake (Figure 4). Unlike the Brule Creek Moraine, this moraine was constructed following a minor readvance of the "northern" ice lobe. Glaciolacustrine sediments underlying till north of the moraine support a readvance. In the vicinity of Dog Lake, the moraine is fronted by a lobate, flat-topped, glaciofluvial plain believed to be deltaic in origin. The elevation of this plain (460 m asl) defines the level of Glacial Lake Kaministikwia, a large glacial lake which occupied the Kaministikwia basin. Distinctive red clays, associated with this glacial lake can be found throughout the basin up to, but not above, this elevation.



Figure 3. Digital elevation model of the Shebandowan greenstone belt and surrounding region.



Glacial Lake Kaministiquia was topographically-supported on its southern flanks by an upland in the northwestern corner of the Kakabeka Falls NTS area (Figure 3) as well as by glacial ice advancing out of the Superior basin and whose terminal position is defined by the Marks moraine (Figure 4). The Marks and Dog Lake moraines are correlative and merge east of the study area along the McKenzie interlobate Moraine. Striae associated with the advance of the Superior lobe range from 340° Az in the northeast to 240° Az in the southwest attesting to the lobate nature of the Superior lobe. This advance is correlated with the Marquette stadial, a late-glacial surge event that resulted in much of the Superior basin being reoccupied by glacial ice about 9.9 ka BP. With the exception of the northeast corner of the Kakabeka Falls area, the Superior lobe advanced into a glacial lake, referred to as "Glacial Lake O'Connor" (Zoltai 1963). Numerous sections along the Whitefish River expose thick sequences of glaciolacustrine rhythmites beneath a fine-textured Superior lobe till. Shorelines of this glacial lake have not been identified.

Till of northern (Patrician) provenance has been observed throughout the entire study area. On Archean terrane, beyond the limits of the Superior lobe, the till is generally less than a few metres thick and has a silty sand texture with low to moderate stone-content. Till cover is significantly more extensive over the east-central part of the Shebandowan NTS map area where thicknesses of 5 to 10 m are not uncommon. Unoxidized till generally has an olive-grey to buff-grey colour. The till often contains a high proportion of clasts that are locally-derived. Faceted and striated clasts within the till indicate deposition by ice flowing towards the south-southwest. Many exposures of this northern till contain remnant clasts of Sibley Group metasediments. These distinctive, reddish-coloured siltstones occur most commonly east of Thunder Bay on the Sibley Peninsula and to a lesser extent, northeast of the study area in the Lake Nipigon basin. Sibley Group clasts in northern till are likely derived from the Nipigon basin, 70 km to the northeast. The non-magnetic heavy mineral concentrates of northern tills are rich in clinopyroxenes derived from Nipigon diabase within this same area.

Northern till is buried beneath variable thicknesses of glaciolacustrine deposits and Superior lobe till within the Whitefish-Kaministiquia lowland. In this area, the till is significantly finer-textured owing its fineness to the underlying shales and argillites of the Rove and Gunflint formations. Stony till, of presumed northern provenance, exposed at water level along Whitewood Creek in east-central O'Connor Township, contains abundant striated and faceted boulders indicating ice flow toward the southwest.

Superior Lobe till is easily recognized by its fine-texture and high proportion of Animikie Group metasedimentary clasts in the pebble fraction. Even in areas underlain by Archean terrane, as occur in the northeast corner of the Kakabeka Falls area, Superior lobe till contains only a small proportion of the underlying rock types in the pebble fraction. This has important implications for programs of mineral exploration utilizing till as a sample medium. The colour of the unweathered till ranges from dark grey, to brown to reddish-brown. Sections of till, containing layers with all 3 colours have been observed in Oliver and Gillies townships and probably reflect the melting out of debris bands derived from different bedrock sources. Superior lobe till also varies markedly in texture. In Oliver Township, near the community of Murillo, Superior lobe till is silty and charged with shales of the Gunflint Formation. The till in this region is fluted and has been observed to reach thicknesses of 7 to 8 m. Similarly, to the south in Scoble and Gillies townships, Superior

Lobe till is silty and charged with clasts of Rove Formation shale and Logan diabase. Along the Whitefish River valley, the till is silty to clayey and contains less than 5% clasts. This is primarily due to the thick deposits of glaciolacustrine silts and clays that were overridden and incorporated into the basal debris layers of the advancing Superior lobe.

OVERVIEW OF FIELDTRIP AGENDA

The landforms, deposits and features we will be visiting today will provide an overview of the important elements of the Quaternary geology of the Shebandowan greenstone belt. An attempt will be made to illustrate how the various Quaternary deposits can be utilized for programs of mineral exploration. Figure 5 shows the locations of the fieldtrip stops. The fieldtrip will begin in the central part of the belt where we will have an opportunity to view deposits and erosional features associated with the advance and retreat of northern ice. Some time will be spent demonstrating sampling techniques, discussing the pros and cons of various sample media, and finally, reviewing strategies for property-scale surficial geochemical surveys. We will proceed easterly into the centre of the northern, Kaministiquia basin where we will have an opportunity to view deposits of glacial Lake Kaministiquia and discuss the glacial lake history of this and other proglacial water bodies within the region. From here, we will proceed south and west where we will have an opportunity to view deposits and landforms associated with the advancing Superior lobe. Stops will include: 1) a section cut into ice-marginal, subaquatic debris flows deposited near the outer limit of the Superior lobe; 2) a Superior lobe ice-contact delta which has been overridden by the advancing Superior lobe; 3) an exposure of Superior lobe till overlying northern till; 4) a panoramic view of the lowland into which the Superior lobe advanced; and 5) a stop to view the morphology of the Marks Moraine. At several of these stops, we will have an opportunity to discuss the suitability of Superior drift as a sample medium for programs of mineral exploration. The trip will conclude by travelling southward into the Whitefish River lowland where we will look at a couple of exposures that illustrate the drastic facies changes that occur in both the Superior and northern lobe tills as you proceed into the lowland. The significance of glaciolacustrine deposits exposed at each of the sites will be discussed as well.

The bus will depart the Lakehead University campus and proceed westward along Highway 102 towards Kaministiquia where it will meet Highway 11/17 at Sistonens Corners. Turn right onto Highway 11/17 and proceed west 16.0 kms. Turn left on a small sideroad situated approximately 150 to 200 m before (east of) a sign on the north side of the highway marking "Bylund Pit Road". Follow the main route on this sideroad for approximately 300 to 400 m. Park vehicle at a large area of stripped outcrop. This is the Bylund Au occurrence.

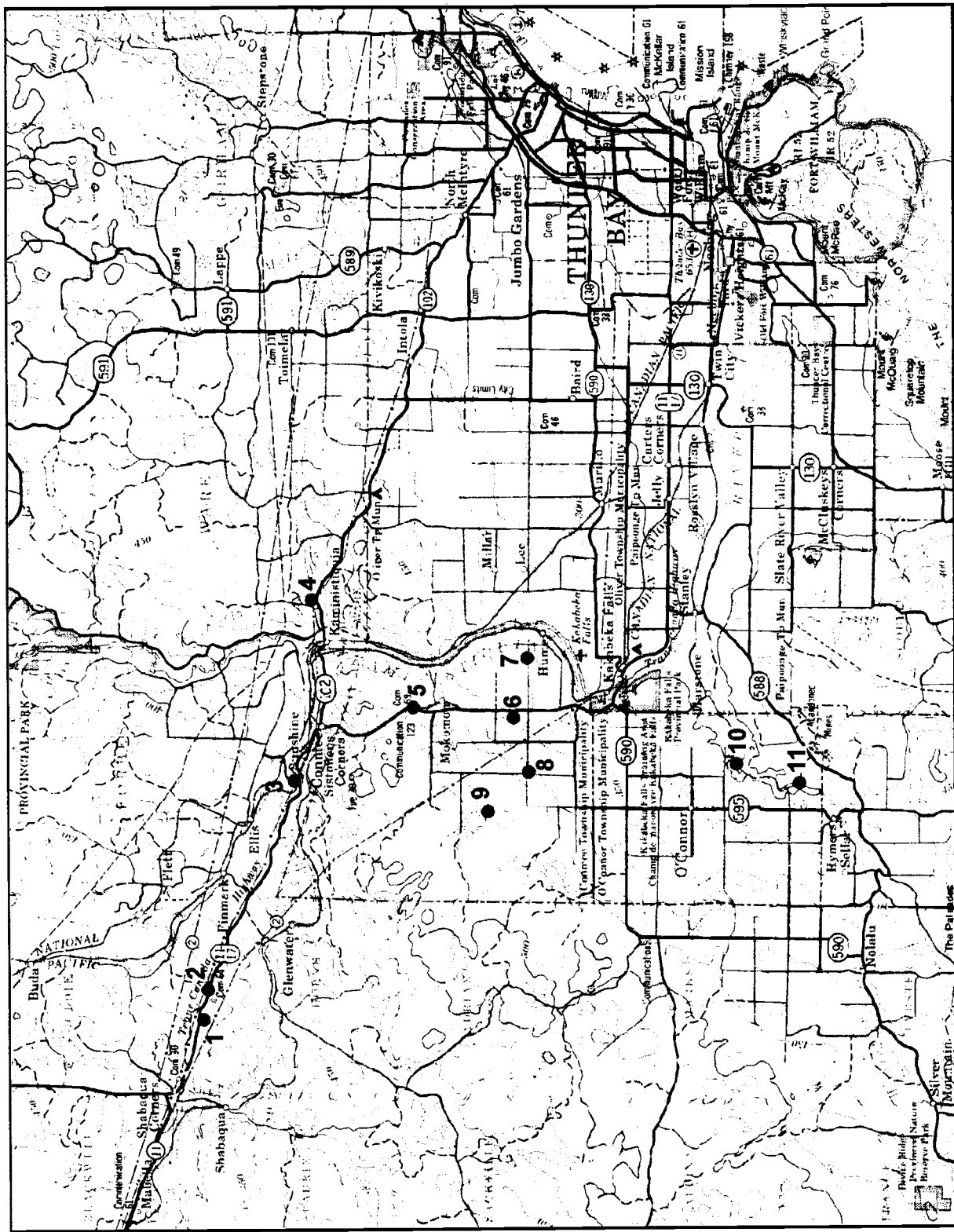


Figure 5. Location map of fieldtrip stops.

Stop 1: Property-Scale Sampling Programs: Bylund Trenches

Regional and property-scale surficial sampling programs have been widely used in the mineral exploration sector for many years. Unfortunately, a significant proportion of the surveys undertaken within glaciated terrain have met with mixed results. Clearly, many of the failures could have been avoided had some effort been put forth in understanding the nature of the Quaternary deposits present within a given area prior to the development of sampling strategies and implementation of sampling programs. At this stop, discussion will focus on the various sampling media that are available to the explorationist for a surficial geochemical perspective, how they are collected and at what density and finally, the analytical techniques employed for their analysis. Trenches exposed adjacent to the Bylund Au occurrence will be used to demonstrate the various media and sampling techniques.

The Bylund Au property is located at the west end of the Dawson Road Lots and is jointly operated by Freewest Resources & Co. Inc. and Greater Lenora Resources Corp. Mineralization is contained within a 125 m wide, east-trending zone of carbonate-altered, mafic to ultramafic metavolcanic rocks transected by a stockwork of northeast-trending quartz-carbonate veins (Schneiders *et al.* 1998). Finely disseminated arsenopyrite is commonly associated with mineralization. The auriferous deformation zone occurs at or near the unconformable contact between Greenwater assemblage metavolcanics to the south and Shebandowan assemblage metasedimentary rocks to the north. Channel sampling over the main Bylund occurrence has returned values of up to 4.1 g/t Au over 14 m. Outcrops exposed in trenches situated both east and west of the main showing have returned values typically in the range of 1 to 2 g/t Au over several metres. Notable intersections include a 0.54 m section, just west of the main showing, assaying 123 g/t Au (C. Bishop, Freewest Resources, personal communication, 1998).

The trenches west of the main Bylund occurrence were opened to investigate Au and As anomalies in soil. They expose a variety of material types. Two facies of till have been recognized; a light, buff-grey, stone-poor, silty sand till of semi-local composition and a stony, silty sand till of local composition. The lower, local till contains abundant clasts of mineralized, iron-carbonate altered sericite schist. It is restricted to the up-ice side of a prominent bedrock ridge where erosive forces beneath the advancing ice were greatest. Compressive subglacial forces resulted in the lower, local till being sheared up into the upper parts of the till section. This till contains anomalous concentrations of gold grains and appears to reflect local mineralization along a major break. The buff-grey till reflects semi-local bedrock sources and was probably derived from the melting out of debris bands in the basal portions of the glacier. Red glaciolacustrine clays overlie the till in low-lying ground to the north. Drift thickness diminishes to less than a metre at the crest of the hill and consists of deeply weathered till. Soil samples collected for mineral exploration from this area returned anomalous Au and As values. The bedrock surface drops off abruptly to the south and is overlain by deposits of different character. Stratified deposits are intimately associated with the till and were probably deposited in a leeward cavity beneath the glacier. These deposits have been modified by glaciolacustrine processes. A thin layer of red clay overlies the till package and is capped by nearshore deposits of sand and gravel. This example clearly demonstrates the complexity and variability that can occur over such a short distance. Clearly, an understanding of the origin of samples collected is required for the successful interpretation of geochemical datasets.

Sample Media

Four distinct media types are commonly sampled for programs of mineral exploration utilizing surficial geochemistry. These include: 1) vegetation; 2) humus; 3) b-horizon soil; and 4) c-horizon till.

Vegetation consists of the leaves, needles, twigs, cones or bark of plants. The premise behind sampling plant tissue for mineral exploration is that the root systems of the plants will absorb metals contained in groundwater, overburden and bedrock at a given sample site. The geochemical response of the plant tissue will therefore provide information on the chemistry of the substrate on which the plant is growing. Vegetation sampling is best utilized in terrains where overburden cover is thin (less than 3-4 m) and locally derived and where root systems are able to penetrate fractures and joints in bedrock. In areas of thick, exotic overburden, vegetation sampling may fail to evaluate local mineral potential. For a regional or property-scale sampling program, it is important to collect samples of the same organ from plants of the same species, size and outer appearance.

Humus is a black to brownish-black, organic-rich material derived from the decomposition of forest litter. It consists of unidentifiable remains of leaves, roots, needles, bark, etc. that have been broken down into a fine-grained, amorphous material by organisms and bacteria. Humus sampling is, in many respects, similar to that of vegetation sampling. Being a first derivative of vegetation, geochemical results should mimic those of vegetation. The same basic rules apply to humus sampling as well. Drastic changes in vegetation type could produce humus with different compositions. Both humus and vegetation are susceptible to contamination from anthropogenic sources of metal. Humus samples collected along Highway 11/17 between the Shebandowan Ni-Cu-PGE mine and the hamlet of Sunshine are strongly contaminated. Hauling of Ni-Cu concentrate along the highway resulted in a corridor a few hundred metres on either side of the highway within which metal values are enriched in humus. The c-horizon till data within this corridor highlights the true geochemical signature for this area.

B-horizon soil is usually the dark, orange-brown material encountered at a depth of 5-10 cm in the soil profile immediately below the humus layer. Often there will be a thin horizon of greyish-coloured mineral matter (Ae horizon) between the humus and B-horizon. This layer (Ae) has been leached of most metals by the downward percolation of groundwaters rich in humic and fulvic acids derived from the decomposition of the overlying organic debris. The B-horizon is the zone of accumulation of clays and iron and manganese oxides derived from the weathering of unstable mineral grains. They are excellent scavengers of metals. B-horizon soils can be developed on parent materials of local to non-local composition. For the successful implementation of b-horizon soil sampling surveys, it is critical that the parent materials be composed of till, that they have similar grain size characteristics and that they be of a local provenance. It is also important that the till cover not exceed 2 to 3 m in thickness.

C-horizon till is the relatively unaltered sediment that underlies the B-horizon. In well-drained upland environments, oxidation can extend down several meters into the profile. Metals contained within sulphide minerals in this zone of weathering are generally liberated by oxidation and

hydromorphically dispersed downwards in the profile. The upper part of the c-horizon will often be enriched in metals compared to the b-horizon. C-horizon till sampling surveys are best undertaken in areas where till thicknesses are less than a few metres. The till will often exhibit a local signature under these conditions. In areas where till cover is thicker, more expensive sampling methods, such as reverse circulation drilling, are required to obtain till samples from the bedrock/overburden interface.

Analytical Methods

Analytical methods employed for programs of mineral exploration utilizing drift prospecting techniques and surficial geochemistry vary depending upon the commodity sought and the sample media collected. For the Shebandowan regional sampling survey, 2 samples of screened c-horizon till (-5 mm) were collected for analysis. A 10 kg sample was collected for heavy mineral concentration and subsequent gold grain, kimberlite indicator mineral (KIM) and metamorphic/magmatic massive sulphide indicator mineral (MMSIM) determinations. Heavy minerals were separated from the -2 mm fraction of these samples by wet gravity tabling, producing a "table concentrate", then further concentrated to >3.2 S.G. using density-dependent settling in methylene iodide. Magnetic minerals were removed from the methylene iodide heavy mineral concentrate (HMC) using an automagnet.

At the tabling stage, a preliminary count of gold grains, including size and shape determinations was undertaken. The gold grain content of the table concentrates was further refined using a secondary panning procedure to obtain more exact grain counts. Only samples containing greater than 10 gold grains at the tabling stage were panned. The size and shape of the panned gold grains were recorded before they were returned to the table concentrate.

A second 2 to 3 kg sample of screened c-horizon till was also collected for -63 micron fraction geochemical determinations. A 30 g split of the -63 micron (silt and clay) size fraction of this till matrix was sent for instrumental neutron activation analysis (INAA). The neutron activation dataset provides excellent information on the Au, As, Sb and REE content of the samples. A second 20 to 25 g split was sent for Pt-Pd-Au determinations by Pb fire assay/ICP-MS finish. A third split was sent for geochemical determinations by ICP-OES and ICP-MS following a standard aqua regia digestion. These methods produce high quality data on the base metal content of the samples. Similar analytical methods should be employed for vegetation, humus and b-horizon sampling programs as well. For humus, the -177 μm fraction is analyzed rather than the -63 μm fraction. Humus samples are briquetted prior to neutron activation analysis. Vegetation samples are either macerated and briquetted or ashed prior to analysis.

Sampling Strategies

The density at which samples are collected is dependent upon the style of mineralization sought. Dispersal trains originating from lode gold type mineralization are significantly smaller than those originating from volcanogenic massive sulphide-type mineralization. The surface area of mineralized rock available for erosion is probably the most important factor that effects the scale of the dispersal train produced. The orientation of the mineralized structure or horizon with respect to

regional ice flow also effects the geometry of the dispersal trains. If regional ice flow is parallel to the strike of mineralization, the resulting dispersal train will be long, narrow and difficult to identify. If regional ice flow is perpendicular to the strike of mineralization, then the resulting dispersal train will be wider. The position of mineralization with respect to regional ice flow is also important. For example, if the mineralized horizon is shear-hosted and situated within a deep ravine perpendicular to ice flow, the advancing glacier may not actively erode the mineralized rock and therefore not produce a dispersal plume from mineralization.

Dispersal trains originating from lode gold deposits of northern Ontario are typically less than 200 to 300 m in length. Sampling programs aimed at locating these types of deposits should therefore be designed to be able to identify a train of this length. A sample spacing of 200 m and preferable 100 m, staggered on 100 m lines, is recommended to recognize a potential target of this type. Once a dispersal train is identified, the density at which samples are collected should be increased to 25 m or 50 m spacing to more clearly define the source of the gold anomaly. Dispersal trains arising from base metal targets can reach several kilometres in length. A sampling program aimed at identifying this type of dispersal train could involve collecting samples initially at 500 m spacing then increasing the density to 100 m in anomalous areas to further refine exploration targets.

Bylund Case Study

High density, property-scale soil sampling was undertaken over and adjacent to the Bylund Au occurrence at the west end of Dawson Road Lots. Fifty-three b-horizon soil samples were collected at 100 m centres on 6 grid lines spaced 100 m apart. The property is characterized as a bedrock-dominated upland with local accumulations of drift (till and stratified deposits) in low-lying areas. Most soils are developed in till, however, a small proportion have stratified sands and pebbly sands as their parent material. These samples were included in the geochemical dataset since most of them were derived from the washing and sorting of the underlying till. Soil samples were not collected from a poorly drained wetland that occurs along the west-central edge of the property. The sample grid is oriented at 150° Az and extends approximately 500 m south of the mineralized zone into unaltered Keewatin mafic metavolcanics. The silt and clay fraction of the soil samples were extracted and geochemically analyzed by instrumental neutron activation analysis and ICP-OES following a standard aqua regia digestion.

Contoured plots of Au and As content derived by neutron activation analysis are presented in Figures 6. Other elements did not show a close spatial association with mineralization. The patterns depicted by Au and As clearly delineate the position of mineralization with little or no evidence for down-ice dispersal beyond a few hundred metres. Similar scales of glacial dispersal have been previously documented at the Matachewan Consolidated Mine in northeastern Ontario (Bajc 1997) as well as at numerous other gold properties within the province. Gold values quickly decline to below detection limit (<2 ppb) within 100 to 200 metres down-ice (south-southwest) from the Bylund occurrence. Arsenic values display a slightly more diffuse pattern.

Elevated Au and As values along L1W and L2W, immediately west of the documented Au zones suggests a possible westerly extension of mineralization. A sample on L1W that returned a

gold value of 1280 ppb and an As value of 88.4 ppm was collected from the fringes of a poorly drained wetland and contained significant amounts of organic matter. This anomaly may have been produced by hydromorphic dispersion from mineralized till and rock on the upland to the north. A similar anomalous Au and As response was obtained, however, from an upland soil 100 metres to the east. These 2 anomalies may suggest a second auriferous structure is present to the south; this warrants further investigation. A strong Au and As anomaly along the central portions of lines 2E and 3E suggest the potential for another, as yet undiscovered, zone of mineralization.

Several conclusions and recommendations can be drawn from the Bylund soil sampling orientation survey.

1. Soil geochemistry is a viable exploration tool for shear-hosted gold in the eastern portion of the Shebandowan greenstone belt. The technique is best suited to upland properties with thin till cover and abundant outcrop.
2. A sample spacing of 100 m staggered on adjacent lines appears to be sufficient for the delineation of potential mineralized zones.
3. Gold and arsenic appear to be the elements best suited for target delineation at the Bylund occurrence. Neutron activation analysis of the silt and clay fraction ($-63\mu\text{m}$) of soil provides excellent contrast between background and anomalies.
4. There is very little evidence in support of down-ice dispersal of gold greater than 100 to 200 m from mineralization. Ultra-high density sampling (i.e. 20 to 25 m spacing) would be required to more clearly define glacial dispersal patterns.

Turn vehicles around and proceed back out to the highway. Turn right, drive 3.1 km east along Highway 11/17 and stop at a rockcut along the sides of the highway.

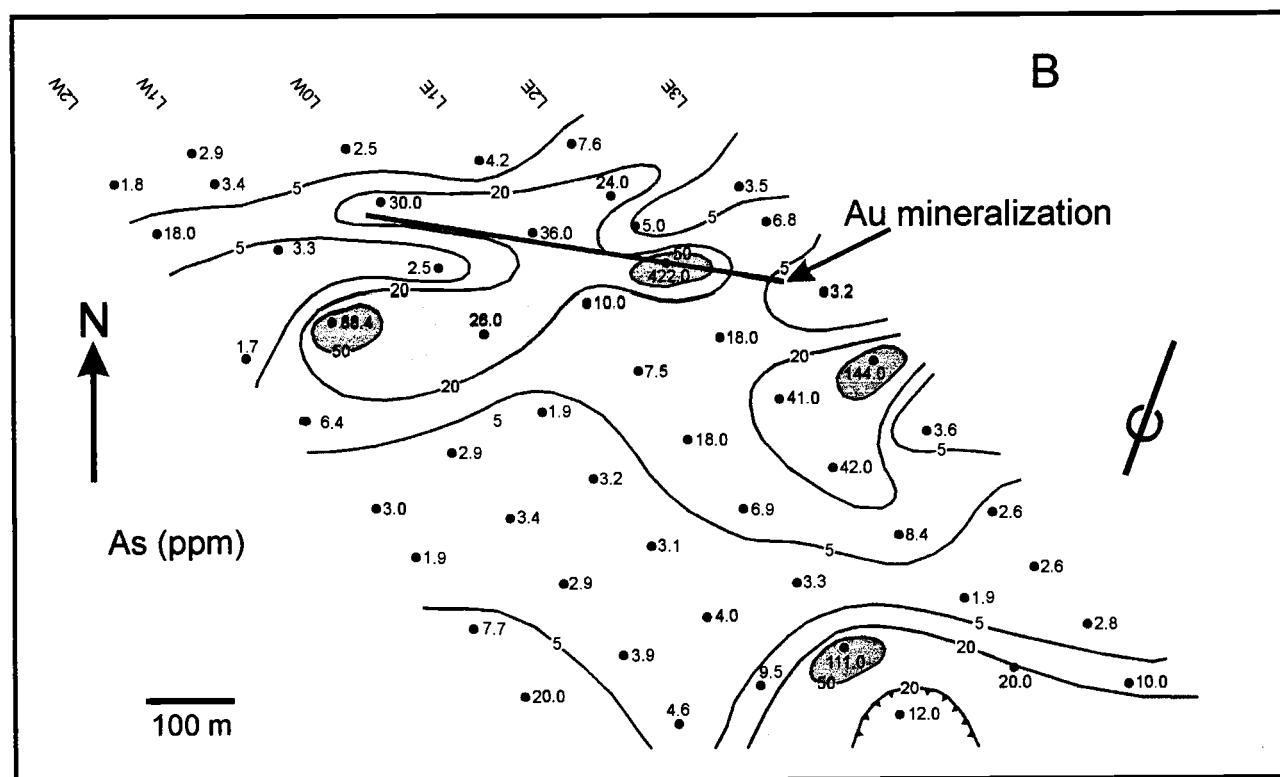
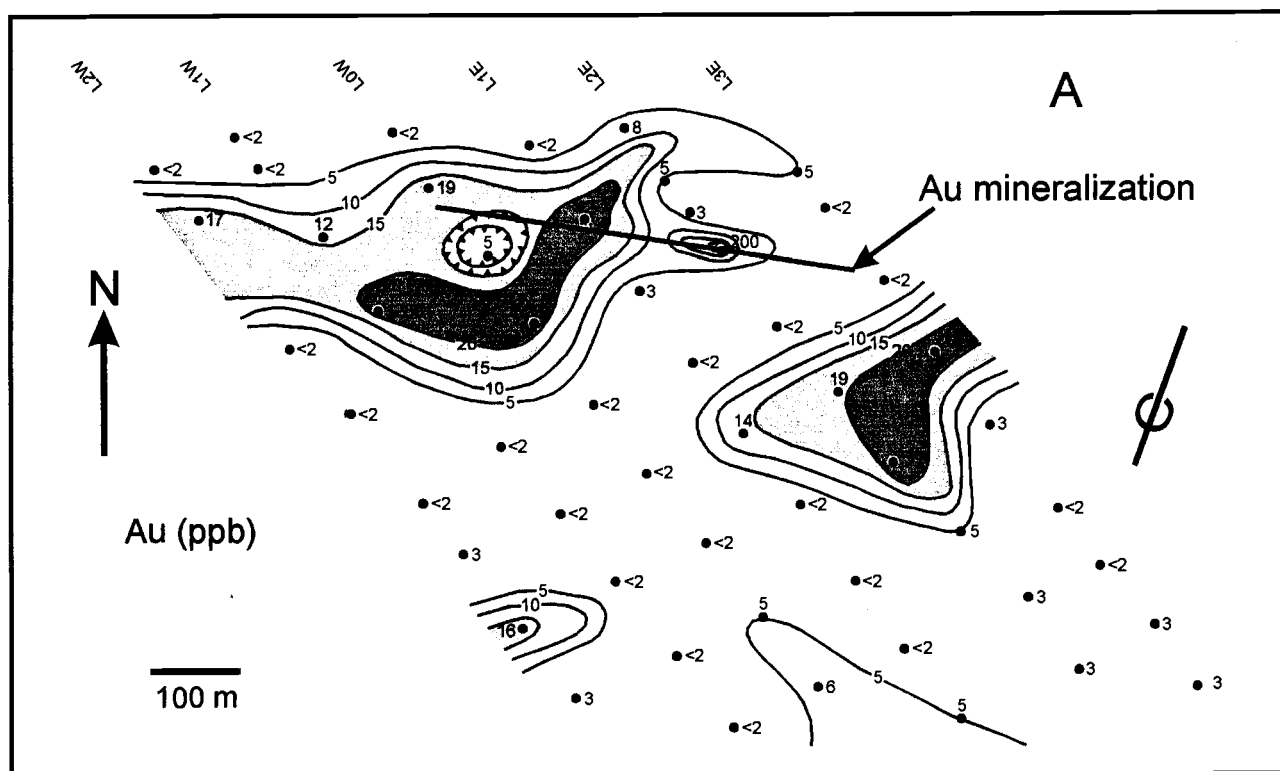


Figure 6. Geochemical response of b-horizon soils over and adjacent to the Bylund Au occurrence, Dawson Road Lots. A: contoured Au data: B: contoured As data.

Stop 2: Crossing Striation Site

At this stop, we will have an opportunity to look at an outcrop of Shebandowan assemblage metasediments that are well striated and flanked by red, glaciolacustrine clay and till. The striation record for the western half of the belt displays a consistent pattern of flow towards the south becoming progressively more westerly towards the west. There is, however, evidence for an older ice flow event towards the southeast in the southwestern corner of the Sunshine map area. The significance of this event is not known. There are several examples of southeasterly-oriented (160° Az) striae crossed by south-southwesterly ($200-210^{\circ}$ Az) striae on the outcrop surface (Figure 7). In most cases, the older striae are preserved on faceted, protected surfaces. This older event is recognized as regionally significant since it has been observed at many sites spanning a 30 km distance and is not apparently related to topography. The till overlying the bedrock surface is believed to be deposited by ice flowing towards the south-southwest (latest ice flow direction). Faceted boulders within the till on the southeast end of the roadcut have striated upper surfaces indicating flow towards the south-southwest. This information is critical for the successful follow-up of till anomalies obtained from this area.



Figure 7. Crossing glacial striae were frequently encountered in the Dawson Road Lots area. In this example, an older set of striae oriented at 160° Az are crossed by a main set oriented at 210° Az.

Proceed 9.8 km east on Highway 11/17 to Sunshine Cross Rd. Turn left (north) onto sideroad and stop by the railway tracks.

Stop 3: Humus Contamination and Implications for Mineral Exploration

At this stop, we will have an opportunity to discuss humus sampling and potential problems that may arise with contamination. This site was chosen for the stop because it is located adjacent to a railway siding that was used by INCO for loading Ni-Cu concentrate from the Shebandowan mine onto railway cars. During the early days of production at the Shebandowan mine site, Ni-Cu concentrate was hauled by dump truck from the mill to the railroad siding at Sunshine, a distance of approximately 50 km. The dispersal of fine particulate sulphides by wind from these trucks resulted in a strong Ni-Cu geochemical anomaly in humus following the Highway 11/17 corridor and extending several hundred metres in from the highway. The Ni-Cu geochemical datasets for the Sunshine and Kakabeka Falls map areas clearly illustrate this contamination (Figure 8). The geochemical datasets for c-horizon till are drastically different from those portrayed by the humus and reflect the primary geochemical signature associated with local variations in bedrock source rocks (Figures 6 and 7). It would appear that there is no signal of contamination within c-horizon till.

All humus sampling programs should consider the potential for problems associated with contamination. Contamination could arise from a variety of sources including road dust from local logging operations, wind-blown tailings dust and airborne fallout from smelters. In the Sudbury area, there is a strong geochemical signature in humus associated with the smelting operations. The element associations closely mimick those of the ore mined in Sudbury. Fine colloidal sulphide particles are dispersed by wind and quickly oxidize in the surficial environment where they are available for adsorption by humic and fulvic acids present within the humus. Should a potential source of contamination be identified within a given area, it is recommended that the explorationist avoid sampling humus and that c-horizon till be considered as the preferred sampling medium for surficial geochemical surveys.

Proceed back out to Highway 11/17. Turn left (east) and drive 3.0 km to the intersection of Highways 11/17 and 102 (Sistonens Corners). Turn left onto Highway 102 and proceed 7.0 kms before turning left into a sand and gravel pit owned and operated by Towland-Hewitson Construction Ltd.

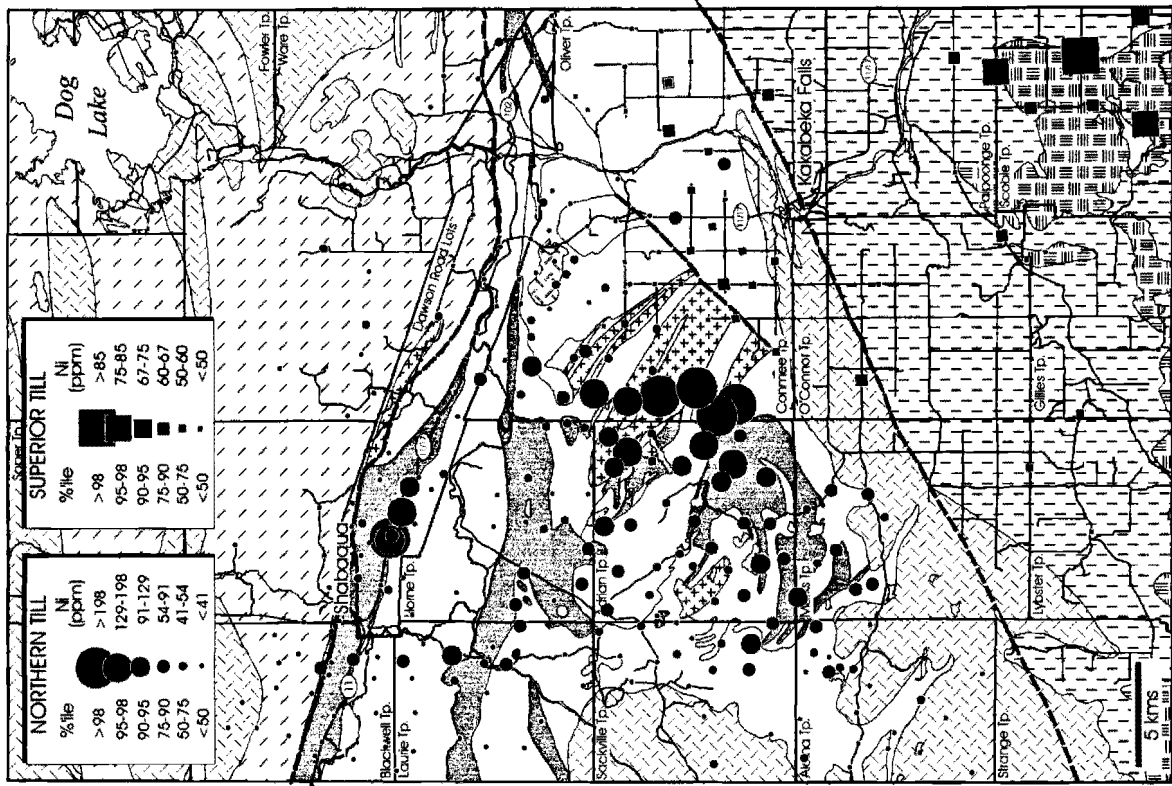
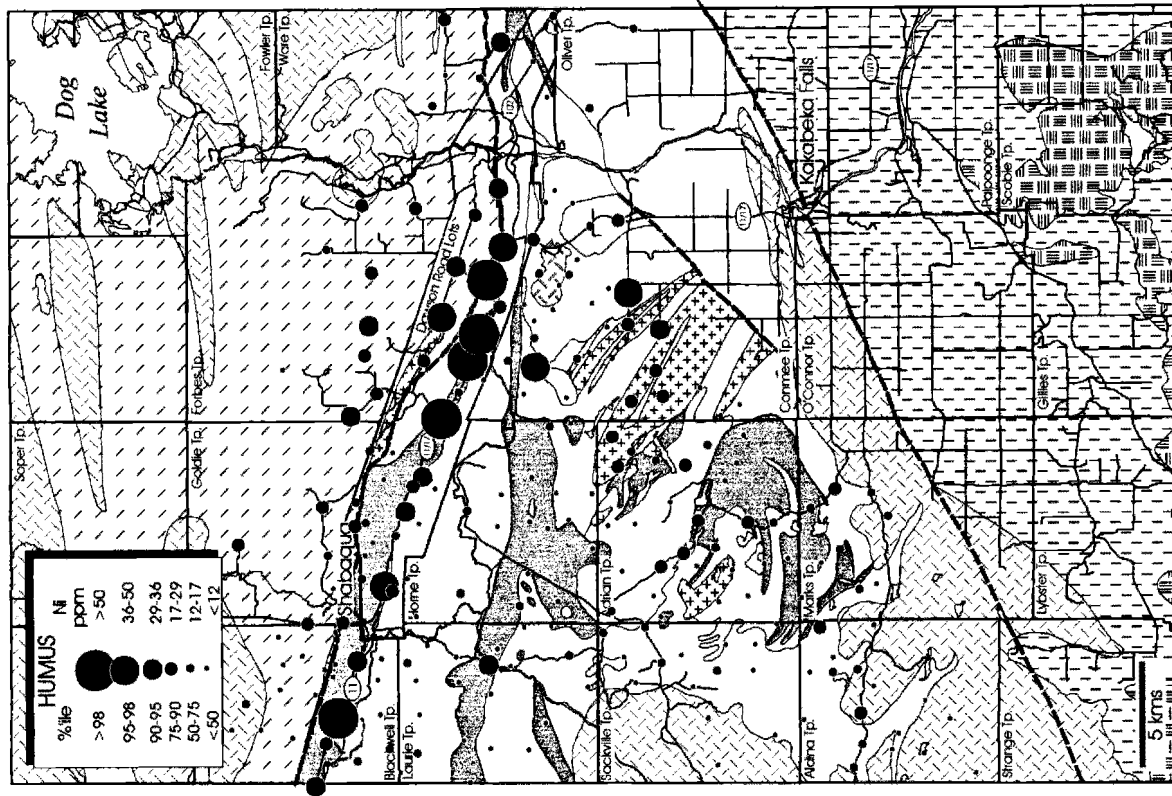


Figure 8. Comparison of humus and c-horizon till geochemical datasets for nickel. Note the strong anthropogenic signature in humus along the Highway 11/17 corridor. This anomaly is attributed to the haulage of Ni-Cu concentrate from the Shebandowan mine to a railway siding near the village of Sunshine.

Stop 4: Glacial Lakes of the Shebandowan Area

Portions of the Shebandowan study area were inundated by glacial lakes. These include: 1) glacial Lake Agassiz in the extreme western part of the study area; 2) glacial Lake Kaministikwia within the Sunshine map area; and 3) glacial lakes O'Connor and high level Superior basin lakes within the Kakabeka Falls map area (Figure 9). The deposits of these glacial lakes are considered a hindrance to programs of mineral exploration utilizing surficial geochemistry in that they conceal and prevent access to deposits of till and the underlying bedrock. The composition of glaciolacustrine deposits does not closely reflect the composition of the underlying bedrock and should therefore not be sampled for mineral exploration purposes. Expensive overburden drilling techniques must be utilized to sample the glacial tills that underlie these deposits.

Glacial Lake Agassiz was the largest of several water bodies which bordered the southern margin of the Laurentide Ice Sheet during Late Wisconsinan glaciation. It inundated an area of approximately 1 million square kilometres, although at no particular time did it occupy this entire area. The lake formed approximately 12.0 ka BP when ice of the Des Moines Lobe retreated north of the continental drainage divide in eastern North Dakota and northwestern Minnesota ponding water within the Red River valley. Progressive expansion of the lake into Manitoba, Ontario and parts of Saskatchewan occurred with further retreat of the ice margin. Glacial Lake Agassiz extended into northwestern Ontario, fronting the retreating ice margin in low-lying areas. Deposits of glacial Lake Agassiz occur in the southwestern corner of the Burchell Lake area along the Obadinaw River valley. The deposits are mainly coarse-textured within this region, however, varved clays have been reported from the banks of Tilly Creek. An ice-contact delta built off of the Brule Creek Moraine south of Burchell Lake grades to an upper level of Lake Agassiz (Herman level) at 457 to 472 m asl. All areas south of this moraine and below this elevation would have been inundated by Lake Agassiz.

Glacial Lake Kaministikwia occupied the valleys of the Oskondaga, Matawin, Shebandowan and Kaministikwia rivers within the Sunshine NTS map area. The lake was bordered on its northeastern and southeastern flanks by glacial ice of the northern and Superior lobes, respectively. High ground bordered the lake to the north, west and southwest. The Dog Lake and Marks moraines mark the positions of the northern and Superior lobes, respectively while Lake Kaministikwia was in existence. An ice-contact delta constructed off of the Dog Lake Moraine just south of Dog Lake and whose upper surface occurs at 460 m asl defines the glacial Lake Kaministikwia water plane. The lake formed within the study area approximately 10.0 ka BP when the northern and Superior lobes readvanced into the study area during a late-glacial surge event (Marquette advance). The extent to which northern and Superior ice retreated prior to this event is not known. Meltwaters initially drained freely southward off of the retreating northern ice margin depositing thick sequences of glaciofluvial sands and gravels within the valleys of the Kaministikwia and Oskondaga rivers. Progressive incision of the fluvial deposits occurred as base levels dropped within the lower reaches of the valleys. Thick sections of glaciofluvial sand and gravel are clearly visible from Highway 102 within the Kaministikwia river valley. These deposits are terraced and draped by distinctive, red clays associated with glacial Lake Kaministikwia. The red clays which can reach thicknesses of several tens of metres, are typically faintly laminated near the base and become progressively more massive upwards. The red pigmentation of the clay is



Figure 9. Digital elevation model of the Thunder Bay area showing the locations of the major moraines and glacial lake basins.

derived from Sibley Group metasedimentary source rocks eroded by the Superior Lobe. Glacial Lake Kaministikwia probably lasted no more than a few hundred years. The lake overflowed westward into the Glacial Lake Agassiz basin via a narrow lowland through Shebandowan Lake. Red clays within the eastern Lake Agassiz basin attest to this influx of sediment from Lake Kaministikwia. Glacial Lake Agassiz must have fallen from a Herman to a lower level prior to the development of Lake Kaministikwia.

Glacial Lake O'Connor occupied the valleys of the Whitefish and Kaministikwia rivers within the Kakabeka Falls map area. The lake fronted the advancing Superior lobe during the Marquette advance and probably drained westward through the Whitefish Lake basin. Shorelines associated with this glacial lake have not been identified. Fine textured glaciolacustrine deposits have however been observed beneath till up to 350 m asl. Deposits of glacial Lake O'Connor occur both above and below till of the Superior lobe. This sequence is typical of an ice advance-retreat cycle within a glacial lake basin. Thick sequences of silty to clayey grey rhythmites, in places exceeding 30 m in thickness, occur beneath fine-textured Superior lobe till within the Whitefish River lowland. Excellent exposures of these deposits occur in large sections along the Whitefish River valley and its tributaries. Grey, rhythmically laminated clays associated with Lake O'Connor overlie Superior lobe till. These deposits have been observed to exceed several metres in thickness. As the Superior lobe continued to withdraw from the Whitefish-Kaministikwia lowland, low-level outlets to the Superior basin were uncovered resulting in the termination of glacial Lake O'Connor. Superior basin waters inundated low-lying areas up to approximately 260 m asl. This represents the approximate level of glacial Lake Minong, the highest water body to occupy the entire perimeter of the Lake Superior basin. Very little is known about the late-glacial history of glacial lakes within this southern basin. Clearly, additional studies are required.

At this fieldtrip stop, we will look at a section of glaciolacustrine deposits associated with glacial Lake Kaministikwia. These red, clay rich deposits are unusual in that they were deposited in approximately 100 m of water less than a few kilometres from the ice margin. The clays are rhythmically laminated at their base, where they overlie glaciofluvial deposits of sand and gravel, and quickly become massive upwards. There is very little indication of a proximal ice margin. That is, there are very few dropstones and few indications of seasonal variability in sediment load.

Proceed back to Highway 11/17 (Sistonens Corners) and turn left (south). Travel 5.2 km south along Highway 11/17 and turn left on Teitto Road. The next stop consists of a road cut on the southeast corner of this intersection.

Stop 5: Subaquatic Flow Till and Glaciolacustrine Sediment of the Superior Lobe

At this site, we begin to see the first direct evidence of the advancing Superior lobe. This 2 m exposure is located less than a few kilometres from the Marks Moraine and consists of well laminated, gritty, pebbly silt and clay overlying massive dark grey silty till. These sediments were likely deposited at the bottom of Lake Kaministikwia as bottom-hugging density currents and as ice-rafted debris (rainout) in water depths of about 50 m. Red Sibley Group siltstones and dark grey shales derived from the Gunflint Formation are clearly recognizable in the pebble fraction and provide direct evidence for a Superior lobe provenance. Red, massive clays associated with glacial

Lake Kaministiquia occur less than a few kilometres to the north of this site. Clearly, the composition of the Superior lobe flow tills and glaciolacustrine deposits do not reflect the composition of the underlying Archean bedrock. This material should not be sampled for mineral exploration purposes.

Turn vehicles around and turn left onto Highway 11/17. Proceed south 5.3 km to Briggs Drive and turn right (east) into Conmee Township sand and gravel pit. Park vehicle at gate if locked.

Stop 6: Overridden Superior Lobe Delta and Conmee Base Metal Float

At this stop, we will be looking at a section of Superior lobe drift. The exposed sequence consists of approximately 6 to 7 m of well sorted, steeply-dipping sand and gravel of Superior provenance overlain by 2 to 3 m of silty Superior lobe till (Figure 10). Clasts in the lower glaciofluvial unit consist primarily of Proterozoic metasedimentary rocks. Numerous cobbles and boulders belonging to the Gunflint Formation and Sibley Group are recognizable. Foreset beds dip steeply to the north and are likely of deltaic origin. The delta was probably built proglacially into an early phase of glacial Lake Kaministiquia. The feature therefore represents a location at which the advancing Superior Lobe stalled prior to reaching its maximum position at the Marks Moraine. The delta is actually located within only 3 km of the Superior lobe limit and occurs at an elevation of about 375 m asl, 85 m below the maximum elevation of Lake Kaministiquia. The delta was overridden by the Superior lobe resulting in the truncation of the foreset unit and removal of the topset beds. Several metres of silty, Superior Lobe subglacial till was deposited on top of the sands and gravels.

Of particular significance is the occurrence of large angular to rounded boulders on the pit floor. The boulders were extracted from the lower glaciofluvial unit and, in some cases, do not appear to have been transported very far. Some of the larger boulders measure several metres across and still display striated surfaces. The boulders are derived from both Archean and Proterozoic source rocks. Several boulders of sulphidized iron formation and massive pyrite of Archean age were discovered in the boulder piles. One of the boulders measured over 1 m in diameter and consisted of massive pyrite and magnetite with 10 to 15% sphalerite disseminated in pyrite-rich sections. Sphalerite was also concentrated along fractures and adjacent to quartz veinlets throughout the rock.

Two samples from the pyrite-rich zones returned values of: 1) 5.13% Zn, 18 ppm Cu, 19 ppm Pb, 260 ppb Au and 0.5 ppm Ag; and 2) 2.85% Zn, 16 ppm Cu, 20 ppm Pb, 245 ppb Au and 0.5 ppm Ag. A sample from the magnetite-rich zone returned values of 850 ppm Zn, 25 ppm Cu, 5 ppm Pb, 25 ppb Au and <0.2 ppm Ag. A second sulphidized iron formation boulder measuring approximately 0.5 m in diameter and consisting almost exclusively of pyrite, returned values of 140 ppm Zn, 8 ppm Cu, 11 ppm Pb, 710 ppb Au and <0.2 ppm Ag.

There are two possible source areas for the boulders. Superior lobe striae in the immediate vicinity of the pit are oriented at 320 to 330° Az. If the boulders were eroded and transported by Superior ice, then there is a 5 km window towards the southeast from which they could have been derived. Proterozoic metasedimentary rocks outcrop beyond the 5 km limit. Alternatively, the



Figure 10. Steeply-dipping foreset beds of a delta constructed along the margin of the advancing Superior lobe (Conmee Township). Silty Superior lobe till caps the sequence.

boulders could have initially been eroded by northern ice from a source to the north-northeast of the pit then remobilized by the Superior lobe.

Exploration work during the early 1900s along the lower reaches of Brule Creek, 4 to 5 km north-northeast of the gravel pit, by B.L. Morrison, the Davis Sulphur Company and General Chemical Company resulted in the discovery of "seven lenticular masses of brecciated, banded iron formation, in which pyrite has replaced a considerable part of the rock" (Carter 1990b). The largest of these masses has a maximum width of 23 m and is 244 m long. Other discoveries include a 21 m wide body of pyrite containing magnetite and pyrrhotite and a 9 m wide by 15 m long zone of magnetite-pyrite-jasper ironstone. It is possible that the boulders found within the gravel pit were derived from this area and that sphalerite was not recognized in the rock. It is not yet clear whether the sulphides indicate proximity to a VMS style zone of mineralization. Further work is required to assess the mineral potential of this area.

Continue south on Highway 11/17 for 0.5 km. Turn left at Holland Road and proceed eastward for 2.6 km to a roadcut at the top of a steep decline.

Stop 7: Superior Lobe Till Overlying Northern Till

The purpose of this stop is to compare and contrast tills of Superior and northern provenance with respect to drift prospecting programs. Exposed in the road cut on the south side of the road is approximately 2 to 3 m of brown, blocky, silty Superior lobe till containing a fairly high concentration of platy clasts derived from the Gunflint Formation, 2.5 km to the southeast. Despite its apparent subglacial origin, this till contains few clasts of Archean supracrustal or intrusive composition. An olive grey, silty sand till of northern provenance is exposed in a few places directly beneath the Superior lobe till. Most of the clasts in this lower till are of an Archean affinity. A quartz-feldspar porphyry is exposed in outcrop further down the slope and probably underlies the northern till in this exposure.

Under discussion is the suitability of Superior lobe till as a sample medium for mineral exploration programs over Archean terrain in the eastern portion of the belt. Compositionally, Superior lobe till overlying Archean bedrock does not reflect local bedrock sources. This is likely attributed to the fact that the Superior lobe was surging into a glacial lake, overriding fine-grained glaciolacustrine deposits and shales. The fine-grained substrates and proglacial lakes facilitated the surging process. Subglacial erosion close to the ice margin was reduced due to a floating ice shelf and reduced subglacial pressures caused by buoyancy.

The background concentrations of gold grains and elements in Superior lobe till are significantly different from those of northern tills. It is therefore advisable to treat datasets containing information from both tills as separate populations for anomaly recognition. Even more importantly, it is recommended that only the lower northern till be sampled in Archean areas affected by the Superior lobe. In areas of thick overburden, it may be necessary to employ expensive drilling techniques to sample the lower till. Alternatively, it may be possible to access the lower till by backhoe trenching off of outcrops in the area. To the south of the Whitefish River lowland, Superior lobe till is exposed on the flanks of the Logan diabase sills. The till in this area appears to reflect local variations in bedrock geology and may be suitable for exploration programs for such commodities as silver or the platinum group elements.

Turn the vehicles around and proceed back to Highway 11/17. Drive straight through Highway 11/17 down Holland Road for 3.2 km. Stop at top of hill adjacent to a microwave tower on the south side of the road. Proceed to the north edge of the sand and gravel pit adjacent to the microwave tower.

Stop 8: Scenic Lookout and Extension of the Brule Creek Moraine

The view afforded at this site provides the fieldtrip participant an excellent view of the lowland into which the Superior Lobe advanced during the Marquette Advance. Thunder Bay is located over 30 km to the east and the Logan diabase sills to the south are approximately 12 km away. Our vantage point is situated approximately 120 m above the Whitefish River lowland to the south and 230 m above Lake Superior to the east. The upland on which we stand clearly affected the advancing Superior Lobe causing it to halt less than 3 km to the northeast of this location.



Figure 11. Oblique aerial view of the Whitefish-Kaministiquia lowland into which the Superior lobe advanced during the Marquette stadial. Mesas of Proterozoic diabase are visible in the distance.

Exposed in the sand and gravel pit before us is a sequence of glaciofluvial sand and gravel of northern provenance which is in turn overlain by a unit of glaciolacustrine sand and finally by approximately 3 m of silty Superior Lobe till. Paleocurrents in the lower glaciofluvial unit suggest flow towards the south-southwest. This is consistent with a northward retreating ice margin. The glaciolacustrine sands are likely associated with an early phase of glacial Lake Kaministiquia.

A number of buried sand and gravel deposits of northern provenance occur in southern and southeastern Conmee Township and adjoining Oliver Township. It is possible that these deposits represent the eastern extension of the Brule Creek Moraine, a significant ice-marginal feature associated with the retreating northern ice lobe. The eastern extension of the Brule Creek Moraine beyond the Marks Moraine was obscured by the advancing Superior lobe.

The next stop is optional depending upon the condition of the road into the site. Continue west another 1.7 km down Holland Road to Sovereign Road. Turn right and drive north 800 m. Turn left on logging road and follow it for approximately 600 m. Turn right at v in road and follow road to the north until you reach a clearcut (~1 km). Park vehicle at crest of hill.

Stop 9: Marks Moraine

This fieldtrip stop is located on the crest of the Marks Moraine, a terminal moraine associated with the advancing Superior lobe. The moraine was constructed approximately 9.9 to 10.0 ka BP, during the Marquette Stadial. It has been traced across the study area from the southwestern corner of Marks Township to the southeastern corner of Ware Township. Mapping is required toward the southwest to determine its position. Where it is best developed, the moraine consists of multiple till ridges within a zone 1 to 2 km wide (Figure 12). It is composed primarily of fine-textured till and therefore displays a relatively low profile. In other areas, the moraine consists of a stacked sequence of flow tills and glaciolacustrine deposits with no apparent geomorphic expression.

The moraine here is asymmetrical in cross-section, displaying a steeper slope on the southeastern, ice-contact side. Fine-textured glaciolacustrine sediments interbedded with Superior lobe flow tills occur beyond the moraine to the northwest. To the southwest are a series of northeast-trending channels that follow the ice margin and terminate in the Kaministiquia basin to the northeast. Paleocurrents of cross-bedded sands within these channels indicate flow towards the Kaministiquia basin.

Drive back out to the intersection of Sovereign Road and Holland Road and proceed south on Sovereign Road for 4.9 km till you reach Highway 590. You have now descended into the Whitefish-Kaministiquia lowland, a vertical drop of over 100 m. Sovereign Road changes into Highway 595 beyond Highway 590. Continue south on Highway 595 for an additional 4.9 km. Turn left on Smith Road and drive for 3.0 km till you reach a bridge crossing Whitewood Creek. Park the vehicle on the right on the far side of the bridge.

Stop 10: Whitewood Creek Section

At this stop, we will be looking at a section along the banks of Whitewood Creek that exposes 2 tills with intervening and superimposed glaciolacustrine deposits. The section has not been studied in detail. It does, however, demonstrate the drastic changes that occur in till character as one proceeds off of the Archean upland to the north into the Whitefish-Kaministiquia lowland. At this stop, we will concentrate on the lower till unit. The steepness of the section makes it difficult to view the upper stratigraphic units. These units will be viewed at the next fieldstop.

At river level is exposed a dark grey, sandy silty subglacial till presumed to be of northern affinity. The till contains numerous faceted and striated clasts supporting its subglacial origin. A stone line or boulder pavement is clearly visible a few metres above water level when the section is well exposed. Striations on the upper surfaces of the boulders indicate ice flow towards the southwest. Most of the clasts are derived from the Gunflint Formation. Shales, wackes, ironstones and oolitic jaspers are commonly seen in the river bed. Only a small proportion of the clasts are derived from Archean terrain to the north. Oxidation of sulphide minerals contained within Gunflint Formation shales has caused a rusty appearance to the till up to the high water line in the section. The character of this lower northern till is quite different from that which is observed over Archean terrain to the north. The till is significantly finer textured, owing its fineness to the texture of the underlying source rocks and displays a much darker colour. It is similar in that it appears to

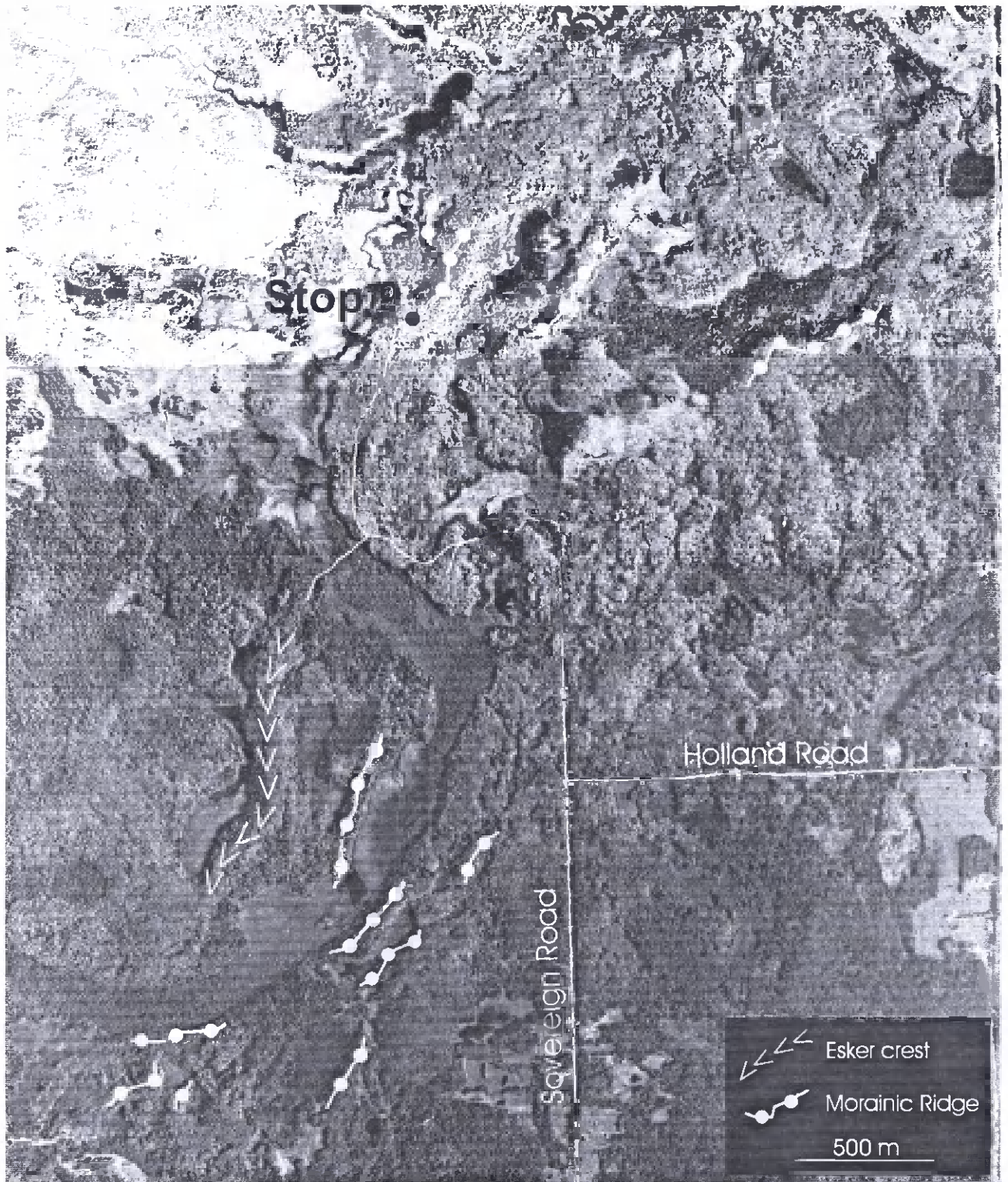


Figure 12. Aerial photo of the area surrounding Stop 9. The morainic ridges form part of the Marks Moraine. The esker ridge was deposited by southward flowing northern ice. Superior lobe till flanks the eastern margin of the esker.

reflect local bedrock sources and therefore is a suitable sampling medium for programs of mineral exploration within this area.

The lower, northern till is overlain by a sequence of silty glaciolacustrine sediments which were likely deposited within glacial Lake O'Connor. These sediments have not been looked at in any detail. There are numerous sections along the valley of the Whitefish River and tributaries that expose Lake O'Connor sediments. These sections require detailed examination to assist with unravelling the history of this glacial lake.

Higher up in the section, immediately above these stratified deposits is a massive silty to clayey diamicton with a few grits and pebbles. This unit is interpreted to be a till derived from the Superior lobe. The till owes its fine texture to the incorporation of large volumes of glaciolacustrine deposits into the base of the advancing Superior lobe. This till barely resembles other examples of Superior lobe till which we have seen earlier in this fieldtrip. The pebbles within this till are easily recognized as derived from the Animikie and Sibley Group metasediments to the northeast and east. This till does not carry a signature of local geology because the ice that deposited it did not have access to the bedrock surface.

The upper, Superior lobe till is overlain by a second sequence of glaciolacustrine deposits. These sediments were probably also deposited within glacial Lake O'Connor as the Superior lobe ice margin retreated eastward out of the lowland.

Turn the vehicles around and go back to Highway 595. Turn left and drive south for 3.3 km. Turn left on Blaikie Rd. and drive east for 1.8 km. The exposure we will be looking at is cut into the walls of the Whitefish River valley and is located on the southwest side of the road.

Stop 11: Whitefish River Section.

At this stop, we will have an opportunity to look at, in detail, a section through glacial lake O'Connor sediments, Superior Lobe till and an upper sequence of glaciolacustrine and fluvial deposits. The lower 3 to 4 m consists of dark grey, well laminated silt and clay likely deposited with glacial Lake O'Connor. The clays are dark grey because their source rocks are the Rove and Gunflint Formations. Glacial Lake O'Connor formed proglacially when the Superior lobe advanced into the Whitefish River valley producing a closed basin supported on its northern and southern flanks by high ground. The lake must have drained westward out through the Whitefish Lake basin.

Eventually the advancing Superior lobe overrode these deposits producing a fine-textured, clayey silt till with few pebbles. This till is very similar to that exposed at near the top of the section at Stop 10. The till is massive, dense, compact and was probably deposited subglacially under active ice. The fineness of the underlying deposits would have prevented the infiltration of subglacial meltwater into the substrate. This would help to promote the development of a thin layer of water at the glacier sole which would have facilitated the rapid advance of ice into the basin. As mentioned earlier, the base of the advancing Superior lobe was isolated from the bedrock surface by

several tens of metres of intervening sediment. As a result, the till deposited by this ice does not display a signature of local bedrock within the Whitefish River valley.

Superior lobe till is abruptly overlain by approximately 2 m of dark grey clay and silty clay. This sediment was deposited proglacially beyond the retreating margin of the Superior lobe. The dark grey colour of this sediment suggests deposition within a closed basin (i.e. a basin not linked with Lake Superior).

This unit is abruptly overlain by approximately 1 m of pebble gravel. The gravel is well-moderately sorted and probably of fluvial character. A low-water phase is implied by this unit. Perhaps the eastward retreat of Superior ice beyond Stanley Hill in southwestern Paipoonge Township resulted in the drop of glacial Lake O'Connor to a lower level. This would have resulted in the influx of coarse sediments from uplands east of the site (Little Pig and Badger Mountains) resulting in the superposition of fluvial sands and gravels over clay.

The fluvial unit is abruptly overlain by reddish-coloured glaciolacustrine deposits of silt, very fine sand and minor clay. The upper surface of this glaciolacustrine plain occurs at approximately 275 m asl, well above the highest elevations reported for Superior basin lakes in the Thunder Bay area. These deposits were likely laid down within a confined basin with a configuration similar to that of glacial Lake O'Connor. The significance of the colour and textural change of the sediment is unclear. Clearly, additional work is required to obtain a better understanding of the glaciolacustrine history of this region.

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GEOLOGY OF THE ARCHEAN STEEP ROCK LAKE - FINLAYSON LAKE GREENSTONE BELT

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Philip Fralick, Lakehead University**

*This 3.0 Ga assemblage consists of ultramafic to felsic
metavolcanic rocks and chemical-clastic sedimentary rocks.
This trip will examine the structural, petrographic,
sedimentological and economic aspects of these rocks.*

LUMBY LAKE TERRAINS

INTRODUCTION

The various Mesoarchean sequences in the Canadian Shield are remarkably similar, being composed of either : 1) a metabasalt-quartz arenite-iron formation association or a 2) metavolcanic-shallow water carbonate association (Thurston and Chivers, 1990; Eriksson et al, 1994). Worldwide examples of these associations typically overlie granitoids or volcanic successions intruded by granitoids and include theoleiitic basalt and komatiite, quartz arenite, iron formation, pelite and/or stromatolitic carbonate. Conglomeratic intervals typically are present at the base of these successions. Included in this association are the Manjeri Formation at the base of the Ngezi Group in Zimbabwe, the Paddy Market Supersequence in the Pilbara Block, and numerous local occurrences in Superior Province (Eriksson et al, 1994; Eriksson et al, 1996). Of the various examples of this association the carbonate dominated Steep Rock assemblage has received the most detailed analysis. Thurston and Chivers (1990) and Eriksson et al (1994, 1996) concluded that associations of this type are formed as a stable platform undergoes rifting.

Data amassed in recent years on these Mesoarchean sequences has, in some cases, led to an elaboration of preexisting tectonic models, while in other cases has caused new models to be proposed. This new data on the Steep Rock-Lumby Lake Terrains is briefly reviewed at the end of this article, with the aim of familiarizing the field trip participants with the new ideas on Mesoarchean tectonics being developed in this area. First we will start with an introduction to and description of the belt.

General Geology

The Steep Rock- Lumby Lake Terrain is located within the District of Rainy River, extending from Atikokan to the northeast, and lying approximately 200km to the west of Thunder Bay, Ontario (Figure 1). **The Steep Rock Greenstone Belt** lies immediately to the north of Atikokan, near the Quetico-Wabigoon subprovince boundary (Figure 2). It was first described by Smyth (1891) as a vertical succession of conglomerate, limestone, ferruginous strata and volcanics. The belt unconformably overlies the 3003-+5Ma (Davis and Jackson, 1988) tonalitic Marmion batholith (Wilks, 1986; Stone et al., 1992). The greenstone Belt comprises the following five Formations defined by Jolliffe (1959), Wilks and Nisbet (1986) (Figure 3) and Stone et al. (1992).

1. Basal quartzo-feldspathic conglomerates and sandstones of the Wagita Formation (up to 150 m thick) representing alluvial fan deposits on fault scarps (Wilks, 1986);
2. Stromatolitic limestones of the Mosher Carbonate Formation (up to 500 m thick) representing deposition in a shallow sea;
3. Iron formation and carbonate-Mn-Fe-rich mudstones of the Jolliffe Ore zone Formation (up to 400 m thick) representing deposition in a subsiding basin (Wilks, 1986);
4. an unusual high-Mg, predominately pyroclastic komatiitic unit named the Dismal Ashrock Formation (50-400 m thick) representing shallow water volcanism; and

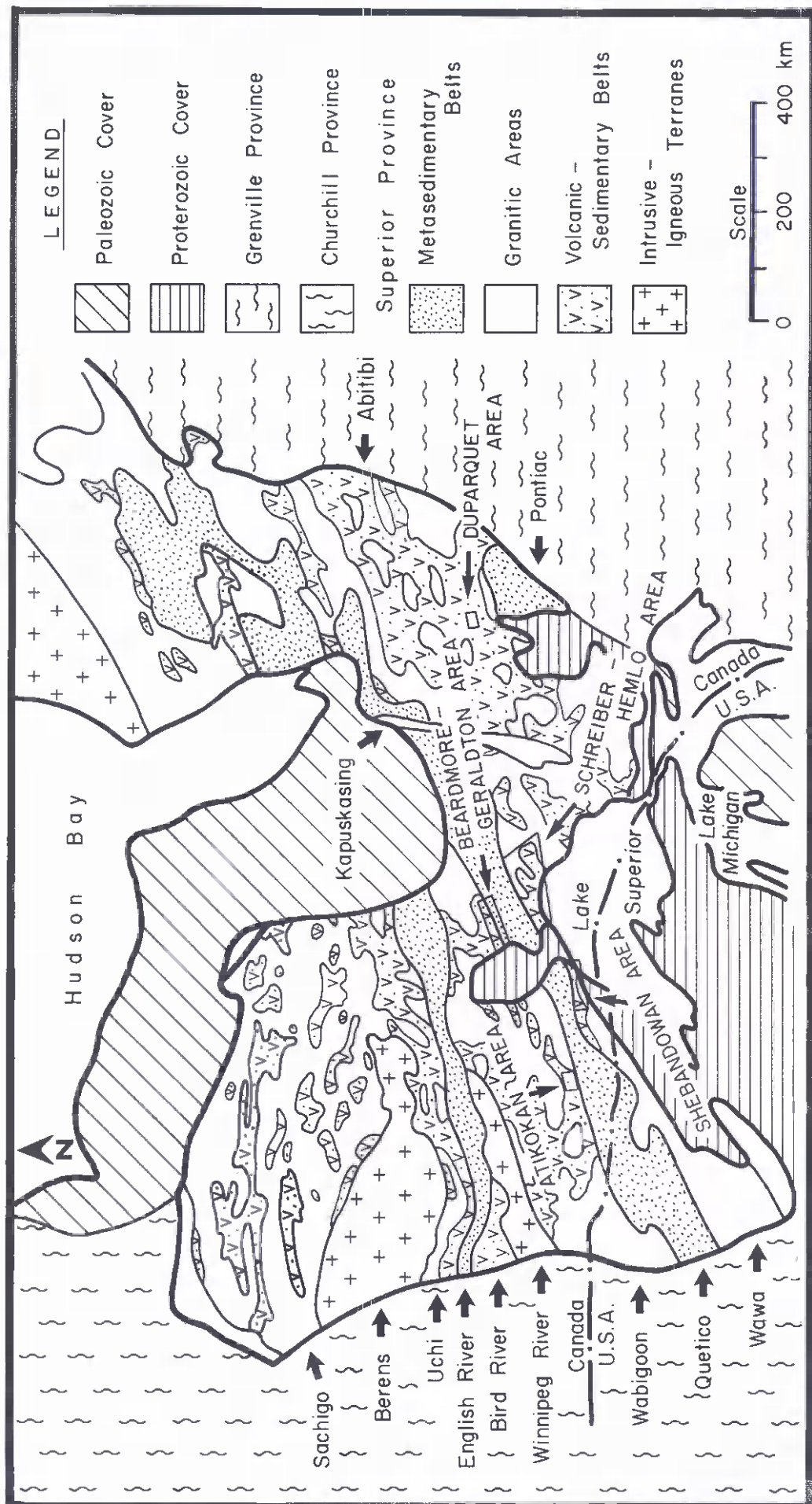


Figure 1. Location of the Atikokan / Steep Rock area in relation to the belt structure of Superior Province.

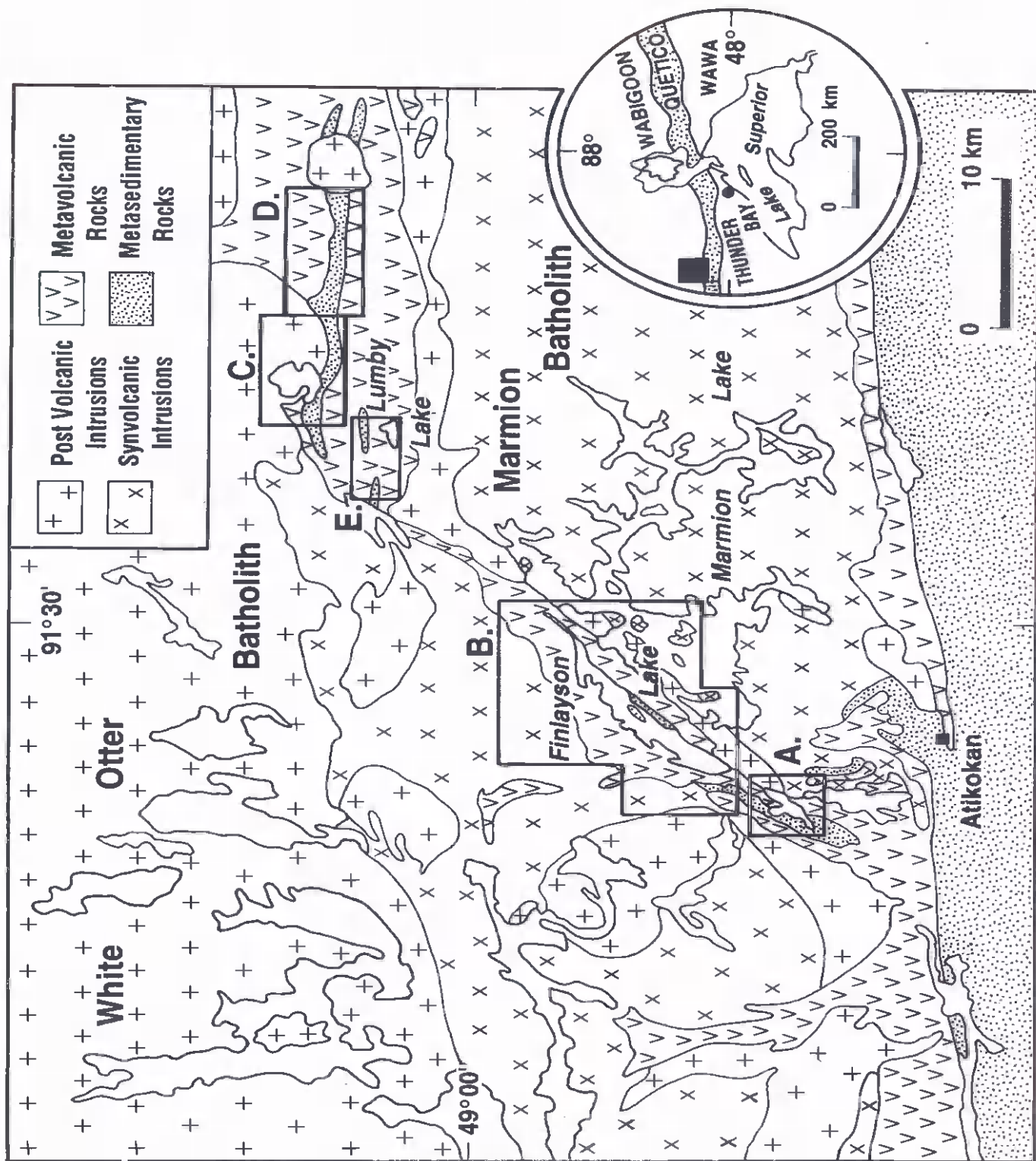


Figure 2.

Geology of the Steep Rock (a and south-east of A), Finlayson Lake (B) and Lumby Lake (C, S, E) areas.

5. a sequence of mafic and minor felsic metavolcanics called the Witch Bay Formation (ca. 5 km thick) representing subaqueous volcanism.

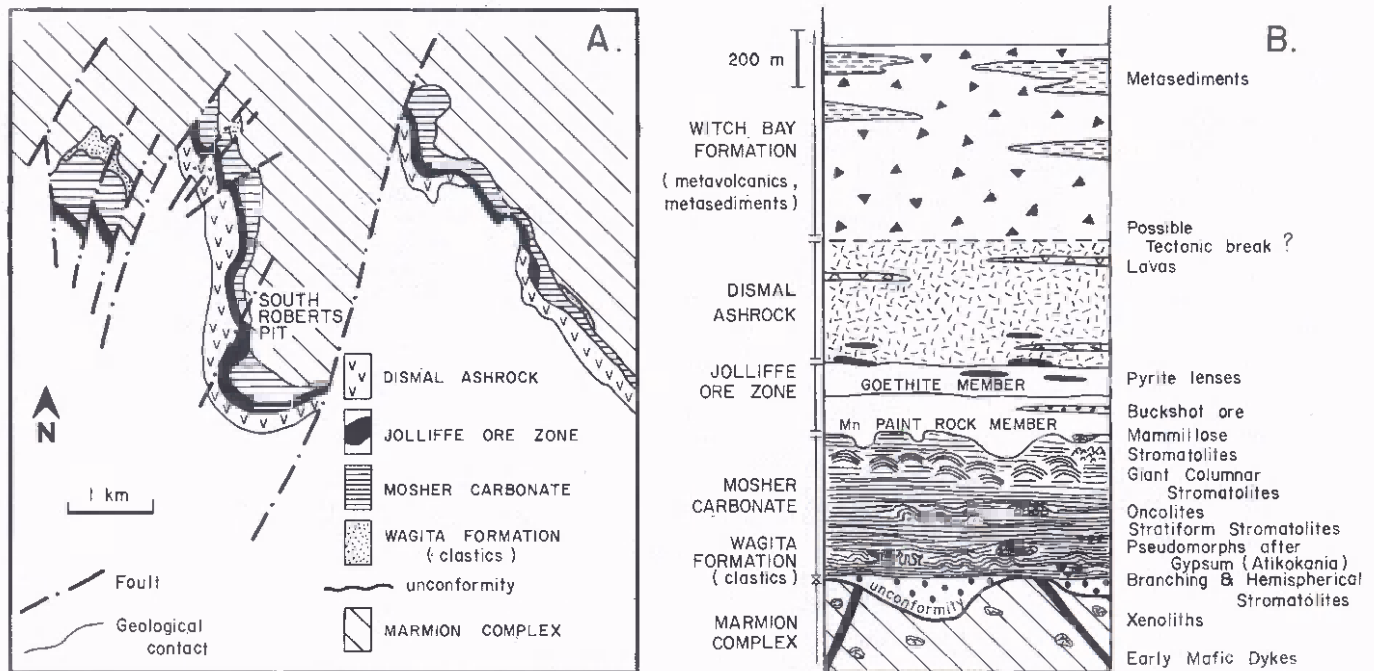


Figure 3. A) Geology of the Steep Rock area. The white area south of the sedimentary units is mostly mafic volcanics. B) Stratigraphy of the Steep Rock succession. (Modified after Wilks and Nisbet, 1986)..

The age of the Steep Rock Group is fairly well constrained. Six detrital zircons from the basal sandstones are ca. 2999 Ma (reported by Fralick and King, 1996 as D. Davis, personal communication). The Dismal Ashrock Formation contains zircons which are 2997 Ma, and these may also to be detrital (D. Davis, personal communication, 1998). Sphenes in the Marmion batholith have been dated at ca. 2950 Ma, possibly reflecting the age of hydrothermal activity that generated the Joliffe Ore zone Formation (D. Davis, persona communication, 1998).

The Finlayson Lake Greenstone Belt joins the Steep Rock Belt to the northwest and extends to the northeast. It is apparently an extension of the Steep Rock sequence, but with the sedimentary assemblage on top of the volcanic pile rather than below it, as is the case with Steep Rock.

The belt is more deformed than the Steep Rock greenstone belt and the rocks are at greenschist to amphibolite facies. The belt is thought to overlie the Marmion batholith (Stone et., 1992) but the contact has been intruded by syn- to post-volcanic tonalite and granodiorite (Stone and Kamineni, 1989). The belt is dominated by sheared, pillowed and massive mafic lava flows and banded amphibolites. Thin units of felsic tuff and metasedimentary rocks are intercalated with the mafic volcanics approximately half way through and at the top of the sequence (Fralick and King, 1996). Six detrital zircons from the upper sedimentary sequence give an age range of 3002 to 2997

Ma (Fralick and King, 1996) and a rhyolite near the top of the volcanic sequence has been dated at 2932 ± 2 Ma (Davis, 1993).

The Lumby Lake Greenstone Belt extends eastwards from the north end of the Finlayson belt (Jackson, 1985a, b). The southern margin of the belt is in faulted contact with the Marmion batholith (Sage et al., 1974). The belt is dominated by mafic volcanic rocks, particularly pillow lavas, but also contains sequences of komatiite flows and pyroclastics, felsic volcanic units and chemical and clastic sediments. The northern half of the belt generally youngs to the south and the southern half to the north, with both halves of the belt dipping steeply. Jackson (1985a,b) suggested that the belt represented a syncline with the fold axis along the sedimentary units at the centre of the belt. This model appears to be oversimplified as there are reversals in younging direction and because the stratigraphy in the two halves of the belt is rather different. We suggest that these opposing sequences may be separated by a tectonic boundary rather than a simple synclinal axis, but do not discount that the two halves of the belt are related.

A felsic tuff unit low in the stratigraphy of the southern half of the belt contains 2999 Ma zircons and a felsic tuff slightly higher in the stratigraphy contains zircons in the range 2999 to 2903 Ma (Davis and Jackson, 1988). However the 2999 Ma component in each of these felsic units is thought to be detrital (D. Davis, personal communication, 1998). This higher felsic unit is also along strike from a ca. 150-m-thick unit (Jackson, 1985a) of sandstones, carbonates and chert thought to have formed in a shallow water environment (lower sedimentary sequence; Fralick and King, 1996).

A thick felsic tuff unit in the northern half of the belt contains zircons that give a maximum U-Pb age of 2973 Ma (D. Davis, personal communication, 1997). An upper sedimentary sequence described by Fralick and King (1996) comprises 500-700 m of chert-rich iron formation and clastic sedimentary rocks that occur along the centre of the greenstone belt. The upper part of the upper sedimentary sequence comprises clastic turbidites with conglomeratic intervals (Fralick and King, 1996).

The mafic and ultramafic volcanic rocks of the Lumby Lake greenstone belt are at amphibolite grade close to the Norway Lake pluton and along the northern margin of the greenstone belt, but otherwise they are generally at greenschist facies. They are, in places, well exposed and, although often sheared, original stratigraphic and textural features are frequently very well preserved, particularly in the northern half of the belt.

Steep Rock Group

Smyth (1891) was the first to describe the Steep Rock succession which he interpreted as a monoclinial assemblage of nine formations that unconformably overlie a granitic basement complex. The unconformable basal contact was confirmed by Lawson (1912), though he reduced Smyth's nine formations to four by eliminating the dubious upper units. Moore (1939), Jolliffe (1966), Shklanka (1972) and Wilks and Nisbet (1988) all redefined the constituents of the Steep Rock Group, with the presently accepted definition being the latter.

Intrusive Igneous Rocks. The Atikokan area straddles an elongate belt of metasediments to the south (Quetico Subprovince) and mixed metavolcanic, gneissic, and felsic plutonic rocks to the north (Wabigoon Subprovince). Pre- to synvolcanic intrusions include mafic tonalite and "old" tonalite

(which predate the Steep Rock Group) in the Marmion batholith and Lefteye stock as well as tonalite and mafic tonalite gneisses of the Dashwa gneiss complex. Syn-to postvolcanic intrusions encompass a broad range of felsic to mafic magmas. "Young" tonalite, which postdates some supracrustal rocks, occupies the Nevison, Righteye, Wasp, Bow, and Hardtack intrusions. The Little Eye stock is composed of hornblende and diorite. Granodiorite and granite occur in the Diversion, Margaret, Bewag, Eye-Dashwa, and White Otter felsic intrusive centres.

"Old" tonalite is so-named because it is the dominant lithology of the basement complex upon which rocks of the Steep Rock Group were deposited. Along northeastern margins of the Steep Rock belt, the "old" tonalite is gradational over distances of several metres to metasandstone and metaconglomerate at the base of the Steep Rock Group. Here the plutonic basement is tan-brown in colour, schistose, highly fractured, and transected by shear zones and abundant metagabbro dykes. The metagabbro dykes are fewer and the "old" tonalite is less intensely foliated in the central Marmion batholith (Marmion Lake area). A sample of this tonalite, collected near the Ontario Hydro generating station, has a U/Pb titanite age of 2953 Ma (Don Davis, Royal Ontario Museum, pers.comm., 1987). Granodiorite of the Margaret and Diversion stocks intrudes the northern and southern margins of the Marmion batholith.

"Old" tonalite is typically medium- to coarse-grained, foliated, and inequigranular to porphyritic. It is composed of plagioclase (40%), quartz (25%), and biotite (20%). Plagioclase and biotite are extensively altered to epidote, sericite, and chlorite in shear zones and adjacent to the Steep Rock belt.

Sedimentary Rocks

The Wagita Formation forms the base of the Steep Rock Group and consists of pebble and cobble conglomerates with angular to subrounded igneous clasts interbedded with poorly sorted sandstones and fine-grained chloritic layers. The clastic sequence varies in thickness from 0 to 150 m and is overlain by carbonates.

The Mosher Carbonates are well banded with the layering accentuated by dark kerogen streaks or millimetre-scale elevated silica-rich ridges at some locations. The laminated carbonates are commonly brecciated where they are spatially associated with fault zones and mafic dykes. Calcite, ankerite and dolomite are present in both the laminated and brecciated units, although the breccias often have an additional, late-stage dolomitic pore-filling cement. Stromatolites are common throughout the carbonate unit, with smaller forms dominating in the lower portions and larger mounds, 3 m in diameter, becoming abundant near the upper contact (Wilks and Nesbit, 1988). The entire carbonate assemblage is up to 500 m in thickness with a very irregular upper contact which may be karstified.

A 100 to 300 m thick succession of manganeseiferous strata termed "paint rock" overlies the Mosher Carbonates. This succession, which forms the lower unit of the Jolliffe Ore Zone, consists of an earthy, poorly consolidated mixture of goethite, hematite, chert, kaolinite, illite, calcite, gibbsite and pyrolusite (Huston 1956). The paint rock is overlain by the upper unit of the Jolliffe Ore Zone, the Goethite Member, which, like the paint rock, is composed of goethite and hematite, but has a lower manganese content (3.8% *versus* 5%, respectively). The Goethite Member, which is 50 to 100 m thick, occurs both as brecciated masses and alternating layers of goethite and chert (McIntosh 1972). The Goethite Member is overlain by mafic to ultramafic pyroclastics and minor thin flows of the Dismal Ashrock.

Wilks and Nisbet (1988) suggest the depositional environment of the basal clastics is a small alluvial fan with a fluvial system that has incised and backfilled channels into the underlying Marmion Complex. As clastic deposition in the area ended, carbonate began to accumulate in a shallow marine setting. The shallow depth is indicated by probable gypsum pseudomorphs, possible desiccated ripples, and polygonal cracks on large stromatolites (Wilks and Nisbet, 1988). Stromatolite morphology may reflect adaptations to various water depths ranging from sabkha to subtidal (Wilks and Nisbet 1988). The karstified surface of the carbonate assemblage suggests an interval of subaerial exposure that caused dissolution of the upper units. It is also possible that acidic solutions moved downward through the sediment pile and dissolved the carbonates. In either case the manganiferous paint rock may represent residual deposits left behind after carbonate dissolution. Little is published concerning the origins of the overlying Jolliffe Ore Zone. The alternating layers of chert and Goethe suggest this unit is a classic iron formation; modified through dissolution or as a paleosol.

Volcanic Rocks The stratigraphy and petrography of the Dismal Ashrock Formation has been described by Schaefer and Morton (1991). Komatiitic lapilli tuff makes up over 80% of the formation, the remainder comprising lenses of komatiitic volcanic breccia, komatiitic volcanoclastic sandstone and siltstone, and mafic to ultramafic lava flows. The rocks are classified as komatiites as defined by Arndt and Nisbet (1982) although spinifex texture is rare. Wilks (1986) recognized possible pseudomorphs after spinifex layering' in one lava flow, and Schaefer and Morton (1991) described micro-spinifex texture in lapilli fragments. The pyroclastic rocks of the Dismal Ashrock Formation are thought to have formed by explosive volcanism caused by ultramafic magma entering water-saturated sediments of the Mosher Carbonate and Jolliffe Ore Zone Formations (Schaefer and Morton, 1991). The top of the Dismal Ashrock Formation is in most places tectonised or invaded by sills but an original stratigraphic relation-ship occurs between the Dismal Ashrock Formation and the Witch Bay Formation near Strawhat Lake (Stone et al., 1992; Tomlinson et al., 1996a; Tomlinson, 1996).

The Witch Bay Formation dominates the Steep Rock greenstone belt and is a sequence of mainly mafic lavas comprising up to 80% pillow lavas. Hyaloclastite and massive lava flows also occur along with rare beds of felsic pyroclastic volcanics. The volcanics have in places been highly folded. Stone et al. (1992) suggested that the original thickness was likely to approach 5 km based on a homoclinal sequence of pillow lavas at the south end of Steep Rock Lake. The Witch Bay Formation is overlain by intermediate to felsic volcanics whose age and contact relationship with the underlying predominantly mafic sequence is unknown.

Finlayson and Lumby Lake Greenstone Belts

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

Volcanic Rocks Volcanic rocks dominate the belts, with mafic lava flows predominating. These commonly have massive lower portions and pillowed upper portions and are rarely capped by brecciated flow tops. Individual flows (ca. 4 m thick) are often separated by thin units of iron

formation and iron rich material is also common between pillows. Variolitic lavas are found in the upper portion of the stratigraphy.

Two komatiite horizons occur in the northern half of the belt. The main komatiitic sequence is 120 m thick and includes numerous komatiite flows, pillowed and hyaloclastite portions of flows, variolitic intervals, subordinate mafic flows and inter-flow sediment comprising chert and iron formation (Tomlinson et al., 196b). Komatiite flows are 4 to 13 m thick with cumulate bases and spinifex-textured tops. This predominantly ultramafic sequence can be traced for ca. 13 km along strike and occurs on either side of the Van Nostrand pluton but it thins to the east. This main komatiite sequence is overlain by further mafic flows and iron formation. An upper komatiitic unit is present east of the Van Nostrand pluton, near the top of the stratigraphy. This upper komatiitic unit varies along strike. To the east it comprises a ca. 20 m thick bedded pyroclastic unit (similar to the Dismal Ashrock Formation) interbedded with thin carbonates and iron formation. West of this is a ca. 20 m thick coarse-grained, unsorted and massive ultramafic pyroclastic-flow unit. Both units overlie a prominent ca. 10-m-thick, chert-rich iron formation and are themselves overlain by further amygdaloidal mafic lavas.

The main sequence in the southern half of the belt is less well known. Fine-grained pillow lavas are common but the exposure is not good enough to establish flow thickness or flow stratigraphy. Some lava units contain plagioclase phenocrysts up to 2 cm across. Poorly exposed komatiite flows of unknown extent occur near the top of the stratigraphy (which were first recorded by Jackson, 1985a). Thin felsic tuff units (10-20 m) occur intermittently in the southern stratigraphic sequence.

Sedimentary Rocks A thin succession of felsic metavolcanic units and metasediments occurs approximately halfway through the volcanic sequence and a thicker assemblage of clastics and chemical sediments caps the assemblage. In the Finlayson Lake area the lower clastic wedge thickens and coarsens to the southwest, and is gradational in this direction to a felsic volcanic assemblage (Figure 4). Beds vary from centimetre-to metre-scale in thickness averaging approximately 70 centimetres. Coarse-grained sand with felsic volcanic-clast pebble bands dominates, except in volcanic proximal areas where the conglomeratic units gain in importance. Clastic mud drapes and mafic tuffs rarely form centimetre-scale interbeds. The sandstones are massive and nongraded. Bed thicknesses decrease to the northeast where the metasedimentary succession is thin and sporadically present.

Little Falls / Finlayson

Lumby

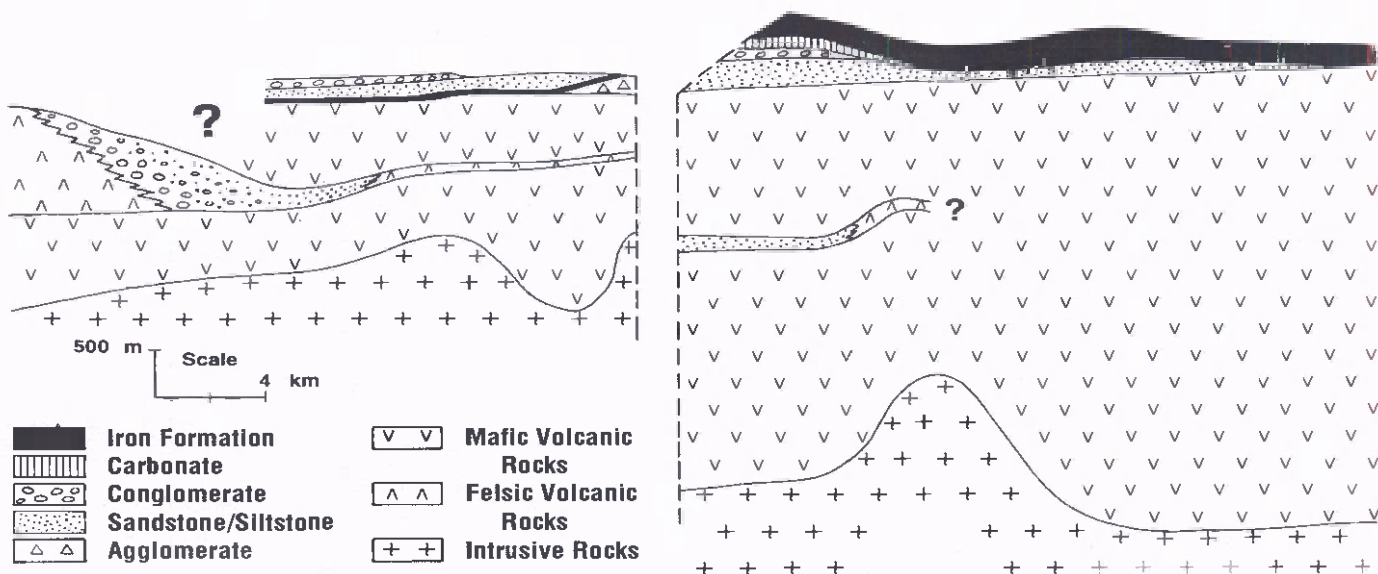


Figure 4. *Reconstruction of the units in the Finlayson and Lumby Lakes area in cross-sectional view. The stratigraphy is discussed in the text.*

These sediments represent a clastic debris apron formed adjacent to, and deriving sediment from, a felsic volcanic centre. Sediment transport was via subaqueous sediment-water dispersions with dispersive pressure probably acting as an important force to keep material in suspension. This requires a high slope, possibly higher than 10 degrees, in the area the flows were generated. Subarial mafic volcanism was also active during this depositional interval.

The lower metasedimentary belt present in the Lumby Lake area is also laterally transitional to felsic volcanics. Sandstones present here are more thinly bedded, averaging 15 centimetres, and do not contain mafic ash interbeds. They are massive, and both graded and nongraded varieties exist. The clastic sequence is intercalated with cherts and carbonates. The cherts are laminated on the centimetre- and subcentimetre-scale and the carbonates are recrystallized with fractures, which may represent original bedding, spaced on the decimeter-scale.

This sequence provides very little information on depositional environment. The sandstones suggest deposition by grain flows and high-density turbidity currents, but this does not imply a water depth. Personal experience indicates that Archean carbonates associated with cherts usually form in shallow water environments, though this remains speculative.

The capping sedimentary sequence present at Finlayson Lake conformably overlies pillowed and massive mafic volcanic rocks in the southern and central area and an agglomeratic pile in the north. Basal cherts and oxide facies iron formation are succeeded upwards by centimetre thick DE turbidites. These gradually thicken and coarsen up sequence to decimeter thick, graded beds composed of medium-grained sand. Above this zone internal, scallop-shaped scours, filled with coarse-grained sand, appear in the beds. These are overlain by pebble and cobble conglomerates with sandstone interbeds and lenses (Figure 5). The coarsening upward trend in this zone is interrupted by abrupt decreases in grainsize to medium-grained turbidites, which again gradationally coarsen upwards to conglomerates.

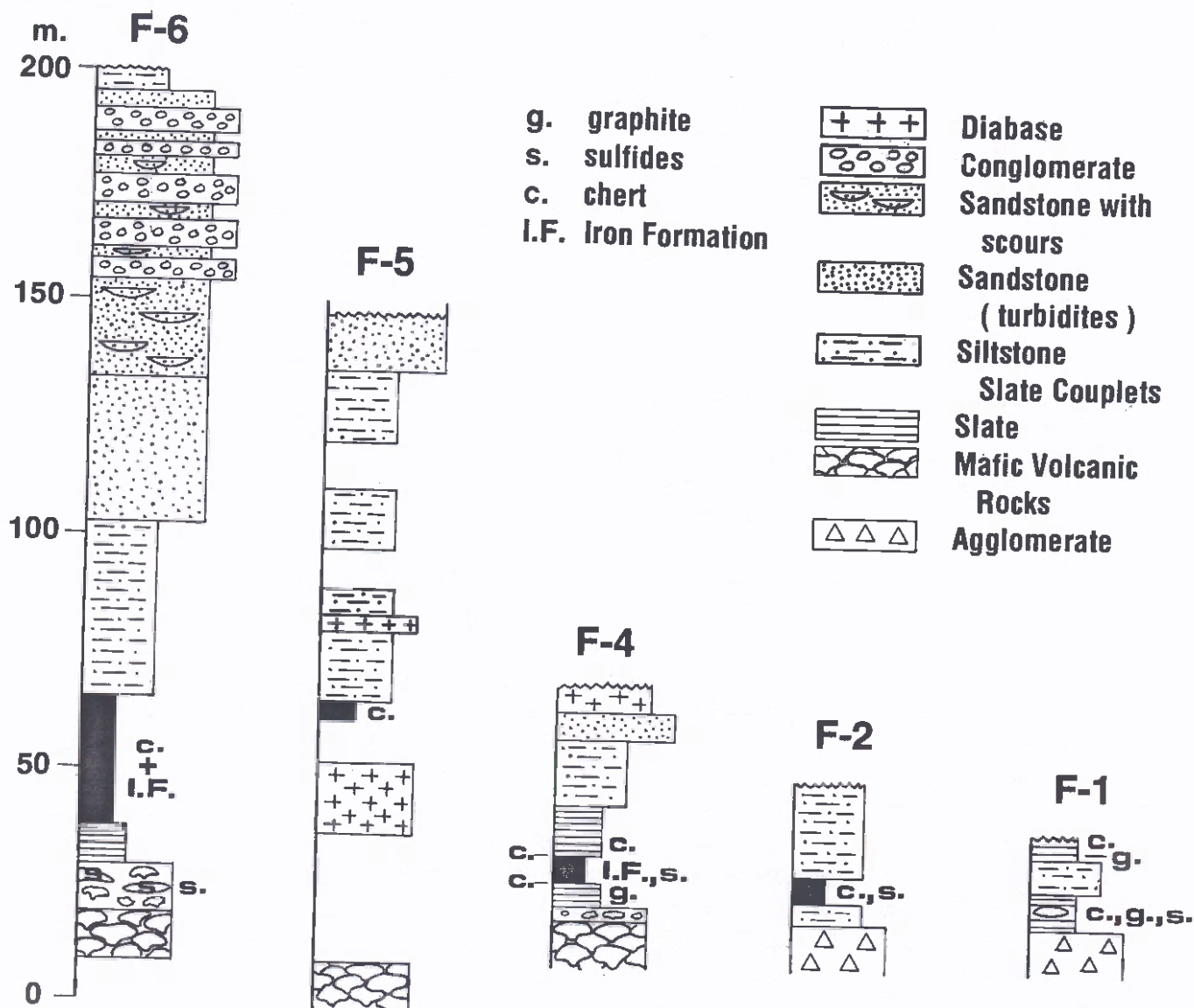


Figure 5. Stratigraphic sections through the upper sedimentary unit on Finlayson Lake. Section F-1 was logged near the north-east end of the lake. The other sections are spaced along a line running approximately 10 km to the south-west and ending with F-6 in the southern portion of the lake. A braided fault system longitudinally cutting the belt in two near its center downsteps the stratigraphy in a north-east direction.

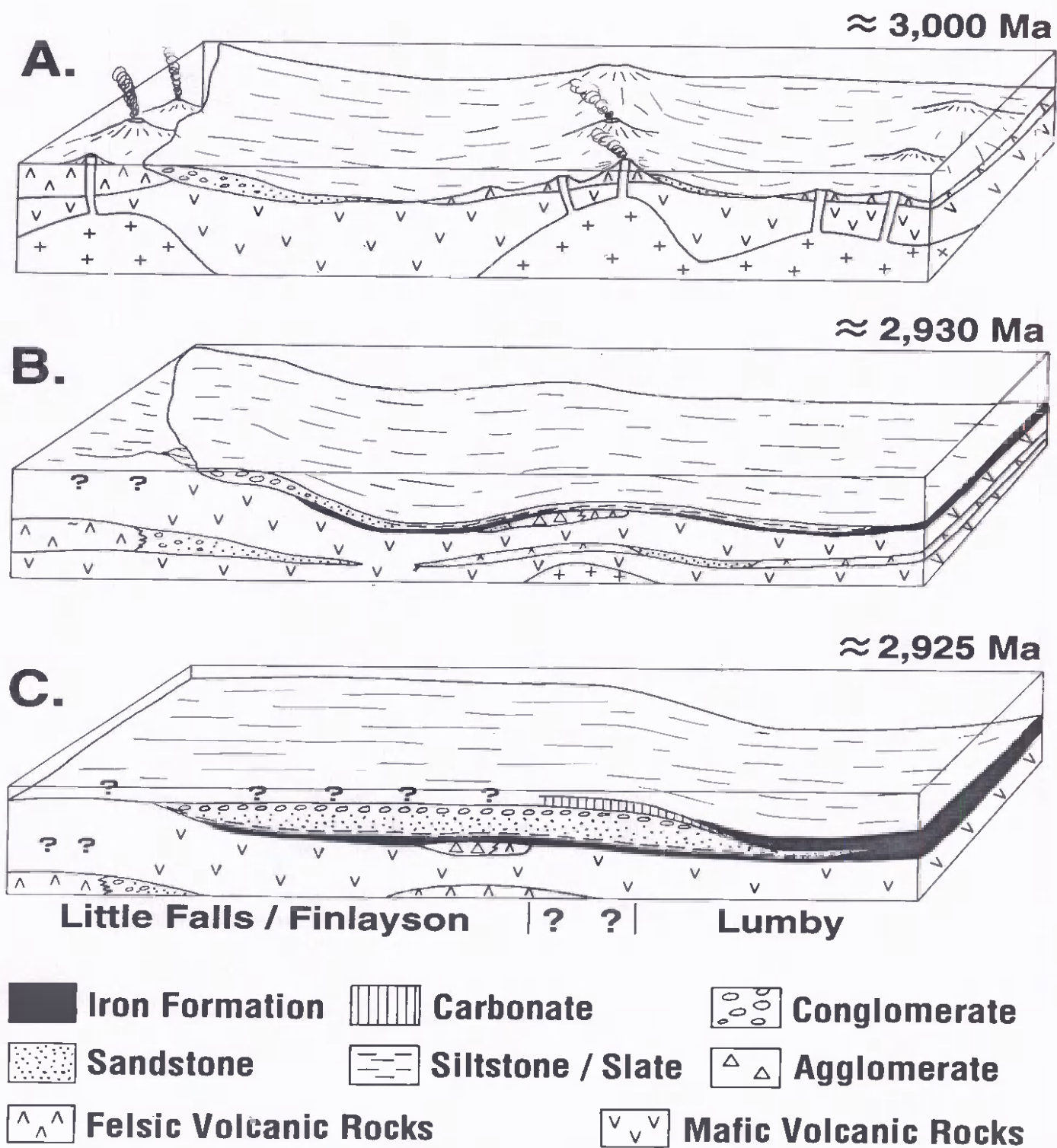


Figure 6. Sequential environmental reconstruction of the depositional environments for the lower sedimentary units (A) and upper sedimentary units (B, C) of the Finlayson and Lumby Lake greenstone belts. Processes are discussed in the text.

This succession represents shoreline progradation (Figure 6). After cessation of mafic volcanism a starved succession of chemical sediments was deposited. The first clastics delivered to the area were clays and silts transported by low-density turbidity flows. This probably represents prodelta debris, though sediment transport via storm currents is possible. Shoreline progradation caused the coarsening upwards succession to form, culminating in deposition in depths effected by wave induced scour. Delta top sands and gravels cap the sediment pile and show the effects of lobe outbuilding and abandonment, with subsequent subsidence induced flooding and a return to deeper water conditions. This is followed by a minor coarsening upward sequence formed by another stage of delta lobe outbuilding in the area and the entire sequence is capped by a final flooding event.

The sedimentary sequence capping the assemblage of mafic volcanic rocks in the Lumby Lake greenstone belt consists of thick sequences of both clastic and chemical units (Figures 5). Way-up directions are ambiguous though it is more likely that the iron formation and chert assemblage overlies the clastic assemblage. A thin zone of chert and iron formation separates the volcanic sequence from the clastic sediments. The clastics consist of decimeter-scale turbidites with metre-scale conglomeratic interbeds in the western portion of the map area. Structural complexities and scarcity of outcrops prohibited ascertaining if a coarsening upwards trend exists. In the west the clastic succession is overlain by thinly laminated carbonates with submillimeter-scale clay drapes commonly developed between laminations. A very chert-rich iron formation assemblage overlies the carbonates in the west, lies on clastics in the central area and directly overlies the volcanic succession in the eastern outcrop area (Figure 4).

Scarcity of outcrop in this belt inhibits a detailed reconstruction of depositional environments. After cessation of volcanic activity a clastic sediment starved interval was followed by deposition via high and low density turbidity currents. Conglomeratic units probably represent the most strand-proximal depositional events. This is followed by the sequence becoming rapidly sediment starved with carbonates deposited in the shallower areas, in the west where the strand-proximal conglomerates are located, and cherty iron formation in deeper areas. With increasing water depth the diachronous contact with the iron formation migrated westward until iron formation was being deposited throughout the entire area.

Structure

The east-trending Quetico-Wabigoon boundary separates turbidites to the south from mainly volcanic rocks to the north and is a first order lithostructural feature in the area. Regional juxtaposition of the contrasting lithologies has been attributed to subduction-related accretion of the Quetico sedimentary prism against the Wabigoon volcanic arc about 2695 Ma ago.

Quetico metasediments have pervasive easterly structural trends and isoclinal, mainly east-facing folds possibly generated by strong dextral transpression after accretion. In contrast, Wabigoon Subprovince shows megascopic oval structures defined by concentrically zoned structural trends in gneisses and supracrustal rocks that envelope felsic plutons. Metavolcanic belts are situated between some oval structures. This structural pattern may have originated due to rise of granites and gneisses in central parts of oval structures attendant with slumping of dense overlying supracrustal rocks into marginal troughs (metavolcanic belts).

The axes of six major folds extend along the central Finlayson belt whereas the Steep Rock belt shows fewer folds and a thick east-younging sequence of pillow lavas that may be a deformed lava cone.

A complex system of oblique-slip and normal faults and metagabbro dykes transect the Marmion batholith. The dip-slip, probably reverse displacement Samuel fault appears to have uplifted a segment of the batholith into the central Steep Rock belt. Displacements of reverse sense are also likely to have occurred on the Atikokan fault, which is conformable with the upper ashrock contact in the Steep Rock Mine area. Clear evidence of a major transcurrent fault in the vicinity of the Quetico-Wabigoon boundary has not been found.

Metamorphism

Supracrustal belts of the Atikokan area are metamorphically zoned. Metavolcanic belts have greenschist assemblages at their cores (chlorite-epidote-plagioclase-calcite±actinolite) and amphibolite assemblages at the margins (hornblende-plagioclase-biotite-ilmenite-sphene). Metamorphic grade does not change abruptly at the Quetico-Wabigoon boundary but increases steadily southward as defined by biotite, Ca-amphibole, garnet, and staurolite isograds. Pressure-temperature conditions in the garnet zone of Quetico metasediments fall in the upper greenschist facies ($P=230\text{--}430\text{ MPa}$, $T=419\text{--}577^\circ\text{C}$).

Except for greenschist assemblages in the Marmion batholith adjacent to the Steep Rock belt, tonalites and gneisses show amphibolite conditions ($P=560\text{ MPa}$, $T=700^\circ\text{C}$).

The high metamorphic grade of gneisses within oval structures is attributed to their having risen from deeper crustal levels, as well as proximity to felsic plutons. Thermal anomalies associated with oval structures and plutons, superimposed on regional burial metamorphism, may be responsible for the metamorphic zonation of these supracrustal belts.

Tectonic Environment

Wilks and Nisbet (1988) were the first to attempt to define the tectonic environment that Mesoarchean strata in the Atikokan area formed in. They believed that a rift, formed by extension of the Marmion block, is the most probable tectonic setting for the Steep Rock Group. This explains the outcrop patterns of the Steep Rock and nearby Lumby Lake successions as having formed on two adjacent tilted blocks. Alternatively, sedimentation may have taken place on a stable divergent continental margin with all or only the lower part of the Wagita Formation representing synrift deposits and the remainder of the succession up to and including the Joliffe Ore Zone reflecting deposition on a post-rift stable shelf. If the overlying Witch Bay volcanics are autochthonous, they record renewed extension.

A study by Thurston and Chivers (1990) put the Steep Rock and Lumby Lake Belts into a group they termed platform sequences. These consist, from base to top, of quartz arenite ± stromatolitic carbonate, oxide facies iron formation and komatiitic to tholeiitic volcanic rocks. They compare these with platform sequences at modern passive margins.

Fralick and King (1995) correlate the sedimentary sequence at the base of the Steep Rock Group with sediments at the top of the Finlayson-Lumby Lake volcanic succession. This implies two periods of mafic volcanism chronologically separated by a sedimentation event. They also stress that the sedimentary sequences record subsidence of an oceanic plateau and the long period of plume driven volcanism is consistent with coupling of the arsenosphere and lithosphere. If they are correct this would negate the operation of modern-style plate tectonics during the early Mesoarchean.

Hollings et al (1998) agree with Fralick and King's (1995) stratigraphic interpretation but advance a more elaborate model to explain the geochemistry of volcanic and intrusive units. They believe that an arc developed, due to subducting lithosphere, which is mainly represented by the deformed and eroded remnants of the intrusive tonalites. Failed subduction of an ocean plateau-ridge complex juxtaposed the arc intrusives with the metavolcanics (Hollings et al, 1998).

Kusky and Hudleston (1999) agree with Hollings et al (1998) that the tonalites represent an eroded arc remnant, but suggest that the sedimentary succession formed a shallow water carbonate platform during arc rifting and the Dismal Ashrock is an allochthonous melange overridden by structurally emplaced arc volcanics of the Witch Bay Formation.

Based on extensive geochemical analysis of volcanic units Tomlinson et al (1999) propose that the unfractionated basaltic komatiites and spinifex-textured high-Mg basalts represent high degree partial melts of the upper mantle. These are geochemically similar to lavas erupted forming the modern Ontong-Java oceanic plateau. Enriched basaltic komatiites represent a deep mantle plume source that has not been mixed with depleted mantle. Volcanism in the Steep Rock, Finlayson and Lumby Lake belts most likely occurred in a continental setting and was probably proceeded by rifting of the continental crust. The stratigraphy of the greenstone belts and evidence of protracted volcanism (ca. 70 Ma) indicates that more than one pulse of plume magmatism may have been involved (Tomlinson et al, 1999).

Obviously opinions differ as to the tectonic setting the succession developed in. The fieldtrip will provide a forum where these ideas can be discussed on the outcrop.

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Day 1 Field Trip Stops

Field stops in the Lumby Lake greenstone belt.

From Thunder Bay travel NW to Upsala on Trans-Canada Highway 17. Approximately 15-16 km further west of Upsala at the Graham logging road, turn left (west) onto the gravel road to Sapawe. This road initially follows the Firesteel River (Provincial Series map 52G/SE, Pakashkan Lake sheet). After approx. 24-25 km you come to a junction where the main gravel Sapawe-Upsala road turns south (left), continue ahead (right hand fork) and after 2 km you come to a bridge crossing Brush Creek (Provincial Series map 52G/SW, Gulliver River sheet). You are at the eastern extent of the greenstone belt and within the northern half of the stratigraphy (Figure 1). This generally east-west bush road runs across most of the length of the Lumby Lake greenstone belt (at a low angle to the strike) before turning south near Lumby Lake and joining the gravel north-south Premiere Lake road (near km 33), which later rejoins the main Sapawe-Upsala road (between km 12 and 13).

Stop 1: UTM zone 15, 643340E 5435920N: BIF, ultramafic fragmentals, mafic lavas.

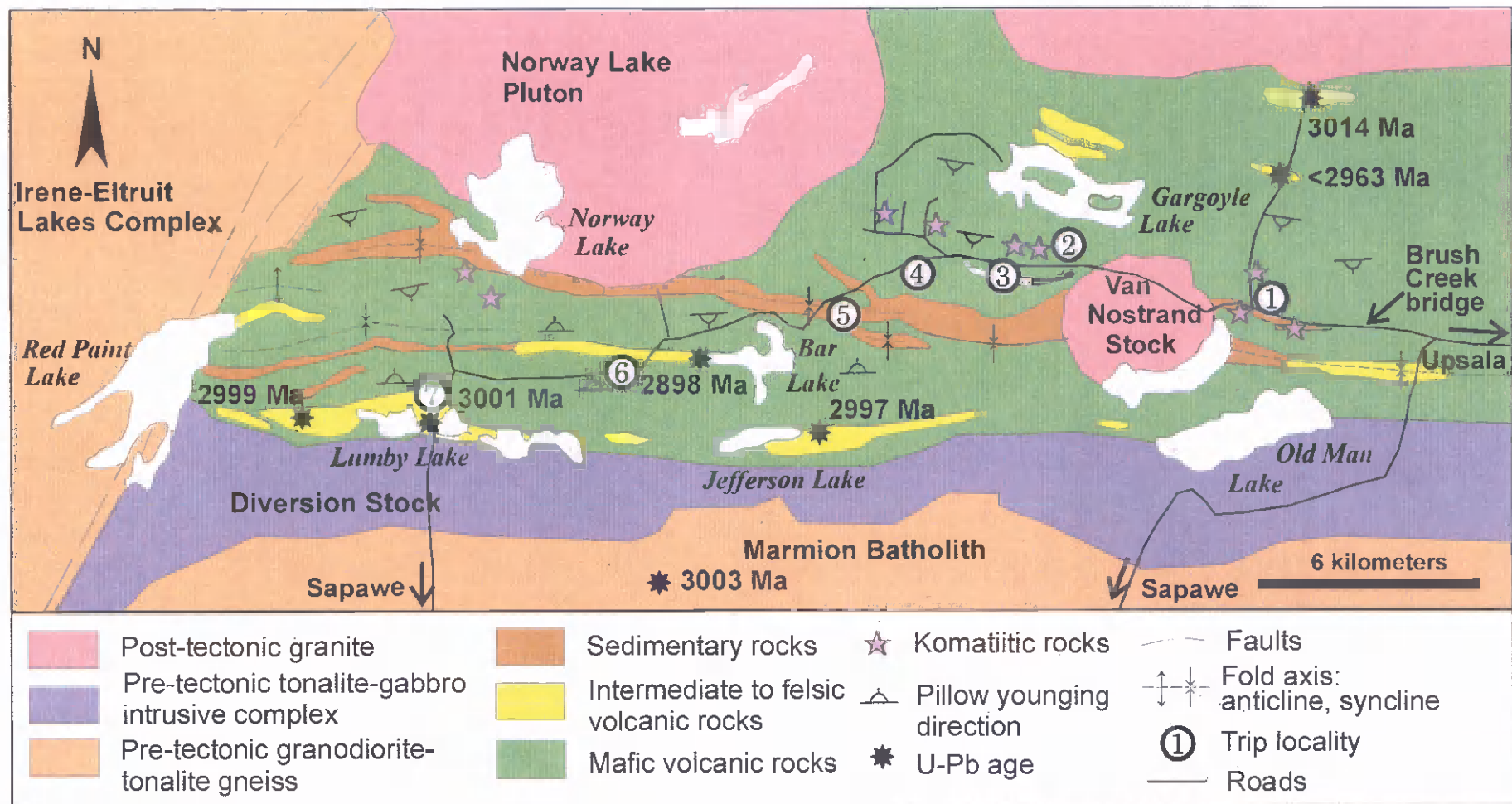
From Brush Creek continue another ~2km to reach the first stop (Figure 1). It is on the left side of the road next to a clearing for parking, and opposite a road heading off on the right hand side. The area was burnt over in 1995 but new trees have been planted and are growing fast. A large outcrop of banded-iron formation is close to the road (~10m thick unit) and then further from the road (south) is an ultramafic fragmental sequence (~20m thick). South of this are mafic amygdaloidal flows, followed by mafic pillow lavas. The ultramafic fragmental sequence is unusual and of debatable origin. It comprises predominantly ultramafic lithic, lapilli-sized fragments within an ultramafic matrix (bulk composition 12 to 21 wt. % MgO, 38-45 wt. % SiO₂). It ranges from ungraded to graded and would be classified un-genetically as a volcanoclastic breccia. In places it shows evidence of hyaloclastite textures and quenched fragments but was it generated by predominantly pyroclastic or autoclastic processes and has it been resedimented?

Geochemically these are Al-depleted komatiites which are enriched in high-field strength elements and light rare earth elements (REE) but depleted in heavy REE. See Tomlinson et al. (1999a) for a detailed description of the chemistry and petrogenesis.

Stop 2: UTM zone 15, 637750E 5437300N: komatiite flows, interflow BIF and chert.

From stop 1 continue a further ~6 km west along the road to a small turn off on the right. Follow this trail for ~200 m then turn left onto another trail and after ~300 m you come to a large, flat, glacially polished outcrop. It may be possible to drive right to the outcrop or the trails may be too muddy. At this stop (Stop 2, Figure 1) there are several large

Figure 1: Geological map of the Lumby Lake greenstone belt showing trip localities
(geochronology from Davis and Jackson, 1985 and Tomlinson et al., 1999b).



komatiite outcrops that have been stripped by a prospector (Ray Bernatchez) and hence the rocks are well exposed with beautifully preserved textures and field relationships visible. The first large outcrop contains komatiite flows, some are thin (<1m thick) with thin pillowed flow tops (sequence youngs to the SW). Spinifex texture is rare on this outcrop. Spherules are common as are ancient cracks that facilitated devitrification. Walk through the bush towards the NW and several outcrops occur along the way, the first on a small scarp shows wonderful spinifex texture. Several small outcrops show banded iron formation (BIF) and evidence of komatiite lava flowing into water-laden sediment. Peperite textures, evidence of autobrecciation, ultramafic pillows etc. are all beautifully preserved. Continue to the NW and a large outcrop of spinifex textured komatiite occurs. Flows are separated by thin units of cherty BIF. A 9m thick flow contains randomly orientated pyroxene crystals (altered to actinolite) which in the top 3m of the flow are spinifex textured with branching acicular blades. Generalized log of these outcrops shown in Figure 2 (from Tomlinson et al., 1996 and 1999b).

Compositionally many of these flows are basaltic komatiites (18-13 wt. % MgO) and spinifex textured high-Mg basalts (9-12 wt. % MgO). Chemically they differ from the ultramafic fragmental units and are Al-undepleted with REE profiles that are generally unfractionated (Tomlinson et al., 1999a).

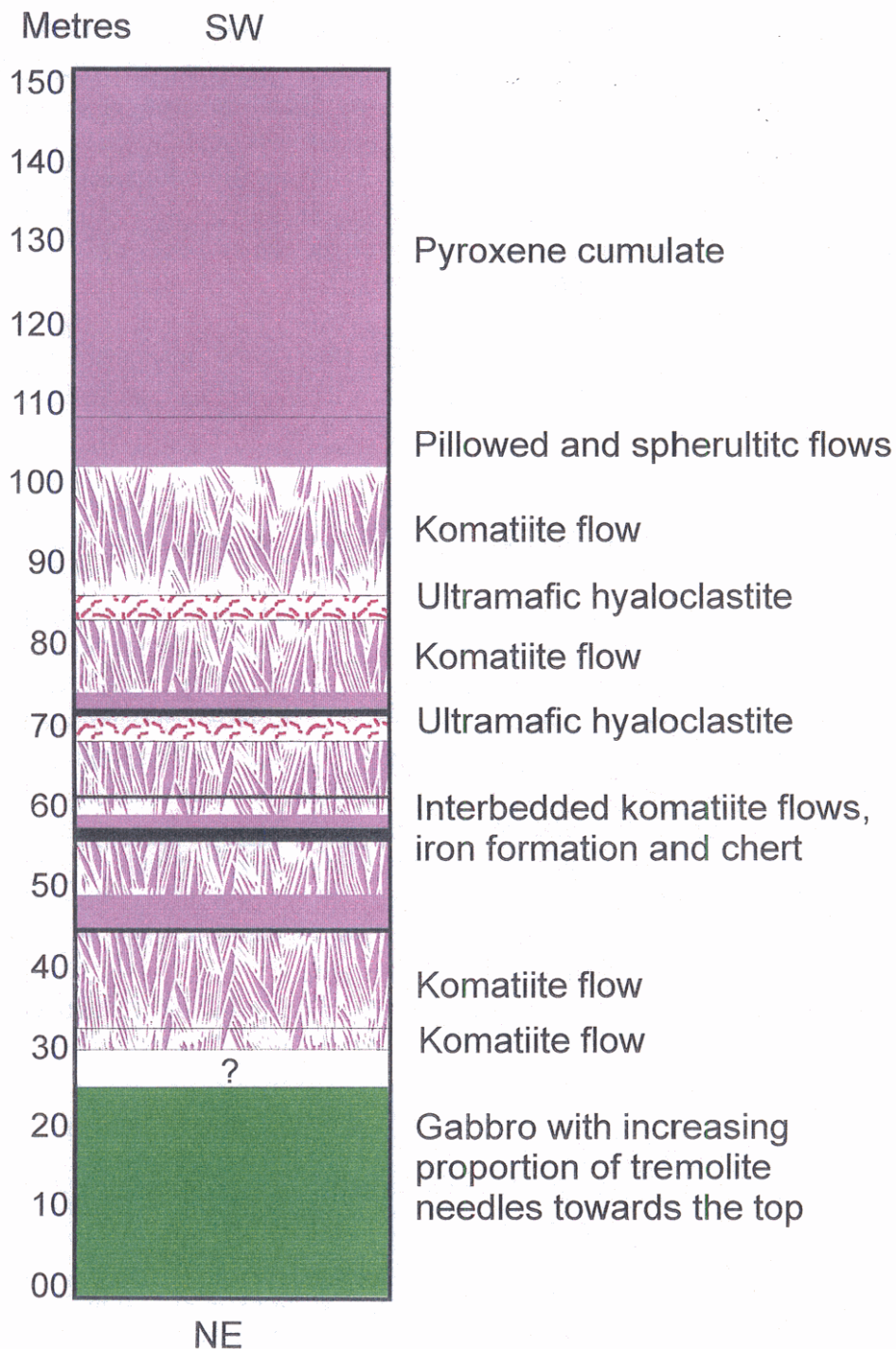
Stops 3 and 4: UTM zone 15, 0637345E 5437130N, and further along the road: Mafic lava flows.

Return to the east-west road that runs across the greenstone belt, reset the odometer and continue driving west. Many outcrops along the road between the komatiite locality and the culverts at Pinecone and Cryderman Lakes (2 large lakes either side of the road after a ~7.5 km stretch) show well preserved mafic lava flows. This lithology (having undergone various states of strain) makes up the majority of the greenstone belt stratigraphy. We will stop at a couple of localities before Pinecone and Cryderman Lakes. In the first (just ~300 m from where we rejoined the road, Stop 3, Figure 1), flow top breccia is overlain by a massive flow base, which is in turn overlain by pillowed lava, which is capped by thin hyaloclastite. The complete flow is 3.8 m thick with 80 cm of pillowed flow. Further along the road several outcrops show undeformed pillows where younging direction is easily identified (any of these outcrops are suitable for stop 4). Pillows are small with convex upward tops and bases that project downwards into the interstices between the underlying pillows.

Stop 5: UTM zone 15, 633266E 5435600N: view.

Continue along the "main" road until reaching the culverts at Pinecone and Cryderman Lakes (Stop 5, Figure 1, ~7-7.5 km from Stop 3), reset the odometer again. These lakes occur at the central east-west axis of the belt within sedimentary units that separate the northern and southern halves of the belt. Unfortunately the sedimentary units are poorly exposed but the view is nice and so is the water if its warm enough for swimming!

Figure 2. Lower komatiite unit, northern assemblage, Lumby Lake greenstone belt (after Tomlinson et al., 1999a).



Stop 6 (if time allows): UTM zone 15, 628130E 5433850N: mafic lavas.

Continue following the road through the belt. The road heads predominantly west and is now cutting the southern stratigraphy at a low angle to the strike. Unlike the northern half of the belt (which youngs southwards), the southern half youngs predominantly to the north. Mafic pillow lavas again make up the majority of the stratigraphy and an outcrop ~ 6 km further on shows good younging to the north (Stop 6, Figure 1).

Stop 7: UTM zone 15, 624000E 5432350N: felsic volcanics at S margin of the belt.

Continue on this "main" road for a further ~6 km as the road eventually turns south into the felsic volcanic sequence at the base of the greenstone belt. Close to Lumby Lake itself a large property has been cleared by Ray Bernatchez and the felsic sequence is well exposed (Stop 7, Figure 1). Felsic volcanic rocks make up a relatively thick sequence here. They have been dated at 3 localities along the southern margin of the belt (including this site) and range from 3001 to 2997 Ma which is within error of the age of the Marmion tonalite. The felsic volcanics are quartz porphyritic and range from fragmental units to massive rhyolite flows.

Continue south and this road rejoins the main gravel logging road down to Sapawe (Provincial Series map 52B/NW, Marmion Lake sheet). A few km south of Sapawe the road rejoins Trans-Canada Highway 11. To the right is Atikokan and to the left is Thunder Bay.

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Day 2. Atikokan area

Day 2 features several unique geologic features spanning 300 My of Archean crustal evolution in the southern Wabigoon and adjacent Quetico Subprovinces. Stops include the early Marmion tonalite batholith, a rare Archean unconformity, the Steep Rock Group (a platform sequence) and contiguous metavolcanic and metasedimentary rocks (Figure 1). An attempt is made to examine the structural relation between these rocks. Most stops are in vicinity of the abandoned Steep Rock Iron Mine from which approximately 100 Mt of ore were removed from 1944 to 1979.

Starting from the junction of Highways 622 and 11B in Atikokan, proceed west on Highway 11B turning north on Mercury or O'Brien Streets. Continue past the Atikokan airport in the direction of the abandoned Steep Rock Iron Mine. After passing under the Canadian National Railway overpass, follow the paved perimeter road along the west side of the mine area. The perimeter road curves around the north end of the mine area and crosses back and forth beneath a power line. Stop where the road crosses the power line for the second time approximately 11 km from the start and walk west to low, rounded outcrops interspersed with grassy areas.

The Marmion batholith -The Marmion batholith is a large oval tonalitic intrusion extending at least 50 km east of Atikokan. Dated at 3003 ± 5 Ma (Davis and Jackson 1988) it is among oldest felsic intrusions in the western Superior Province and together with the pre 2900 Ma Dashwa Gneiss Complex and Finlayson and Lumby greenstone belts (Davis and Jackson 1988) is part of the enigmatic older volcanoplutonic terrane in the central Wabigoon Subprovince.

The Marmion batholith formed a mature erosional surface on which the Steep Rock Group was unconformably deposited at an unknown but probably early stage in tectonic evolution of the area. Although rare in central parts of the Marmion batholith, metagabbro dikes are voluminous near the Steep Rock belt; some dikes seem to predate whereas others postdate the Steep Rock Group. Although the tectonic environment at the time of Steep Rock Group deposition is speculative, the dikes imply extensional conditions, possibly in a rifted continental margin setting (Wilks and Nisbet 1988).

Felsic and mafic varieties of tonalite are mapped in the Marmion batholith near the Steep Rock belt. These are typically medium-grained, weakly foliated and inequigranular to porphyritic rocks composed of essentially plagioclase, quartz and biotite and up to 20% hornblende in the mafic variety. The Marmion batholith is cut by ductile-brittle faults and primary minerals are extensively converted to sericite, carbonate, chlorite and actinolite within a few kilometers of the Steep Rock belt.

Stop 6-1 The Marmion batholith adjacent to the Steep Rock greenstone belt

Exposures in the power line right-of-way are somewhat poor due to an intense brown weathering rind but consist of tonalite interspersed with about 30% metagabbro dikes. Dikes are irregular in shape and are mainly north-northeast trending. All rocks are well foliated to friable and intensely altered to greenschist minerals. A few hundred meters south of this locality the Marmion tonalite is overlain by the Steep Rock Group.

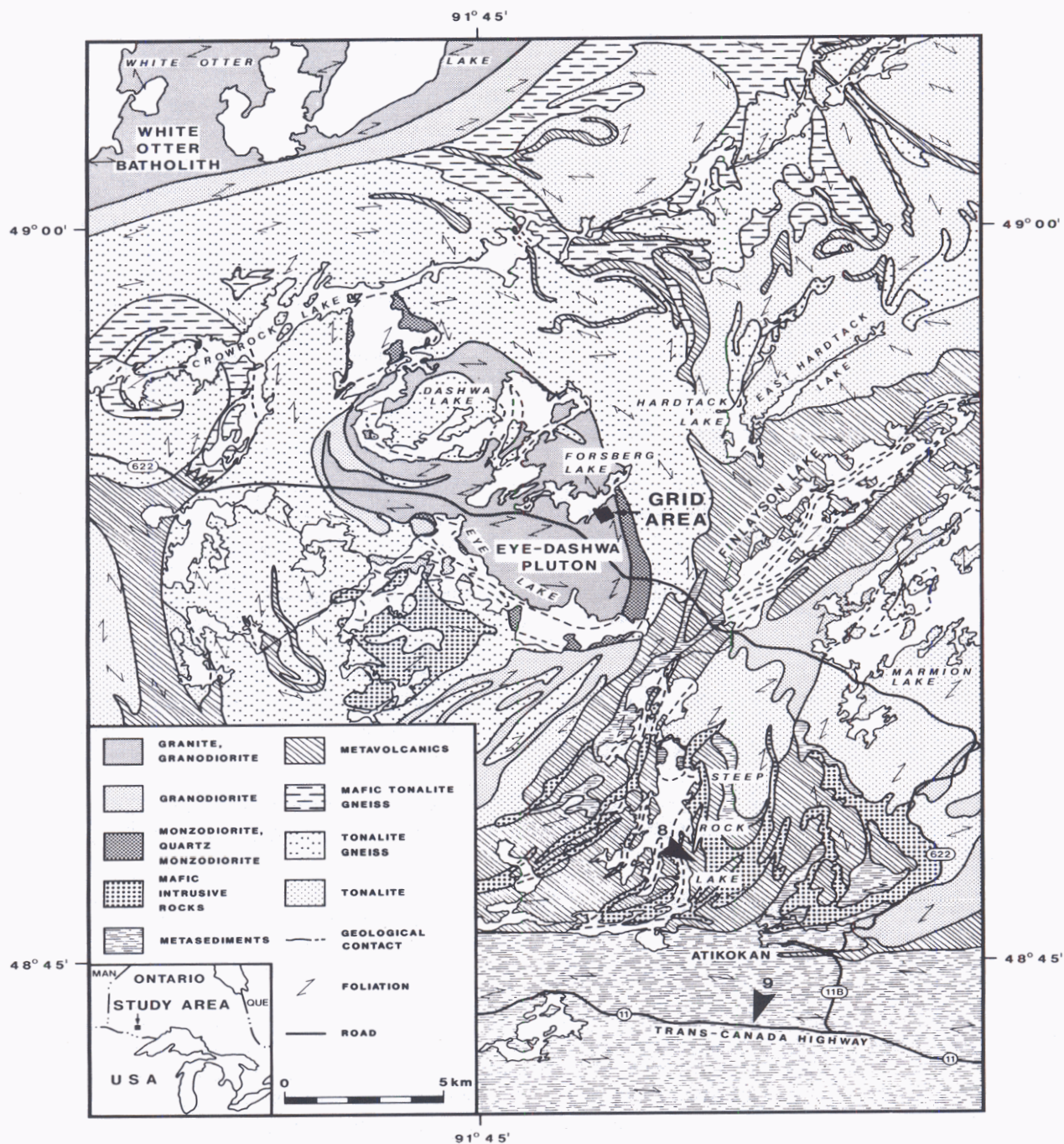


Figure 6-1: Geologic map of the Atikokan area shown field-trip stops 6-8 and 6-9. See Figure 6-3 for other stops.

From the power line, follow the perimeter road south for a distance of about 1 km where the pavement ends at an abandoned railway crossing. Park in the grassy area near the end of the paved perimeter road and walk west along the torn-up railway roadbed.

Unconformity at the base of the Steep Rock Group - Greenstone belts are interspersed with voluminous felsic plutonic rocks throughout the Superior Province and in the majority of places the boundaries between the two are marked by sharp contacts where plutonic rocks intrude and contact metamorphose greenstone belts. Rarely is the base of an Archean greenstone belt preserved although in recent years, an increasing number of localities have been identified where the bases of greenstone successions lie unconformably on felsic plutonic rocks. Most are found in remote northwestern parts of the Superior Province. The Steep Rock Lake locality represents an extremely rare, well-preserved and accessible example of an Archean unconformity and is featured in several stops on this trip.

Stop 6-2 Unconformity at the base of the Steep Rock Group

Outcrops at this locality illustrate the gradational nature of the unconformity at the base of the Steep Rock Group. After walking about 200 m along the torn-up railway line past outcrops of mainly tonalite, you will see a large vertical rock-cut on the east side. Foliated, relatively fine-grained dark metagabbro makes up most of the vertical rock-cut and, at the north end of the vertical rock-cut, is transitional to friable, buff, gritty metasandstone (Wagita Formation). The metasandstone is poorly bedded, but locally shows clasts, and is well foliated, possibly indicating enhanced shearing at the unconformity. The metasandstone can be traced north from the large vertical rock-cut through a series of small exposures to a rounded outcrop of metaconglomerate that lies opposite piles of mine-waste placed on the railway line. Boulders of the conglomerate are mainly tonalite and a dark rock that is probably metagabbro.

Beyond the metaconglomerate, rock-cuts show carbonate breccia, tonalite and metagabbro. Outcrops at the obstructed tunnel entrance are metamorphosed pelite and sandstone.

The section along the railway line passes back and forth through the unconformity at the base of the Steep Rock Group. Rarely is the unconformity sharply defined as the transition from tonalite to sandstone takes place gradually over distances of a few meters.

Return to the vehicle and drive .9 km south along the gravel perimeter road past remains of the head frame of the Hogarth shaft and a branch road (east) to the former concentrator plant area. Turn right and proceed .2 km down the pit access ramp. Turn right at the first side-road and proceed .15 km and park at the outside of a sharp bend overlooking the pit. Walk down the gravel slope following an erosional channel past several rusty, jagged outcrops of tonalite and gabbro. Follow a flagged trail from the base of the gravel slope about 100 m westerly through a wooded area and emerge onto a north-facing slope interspersed with outcrops, boulders and

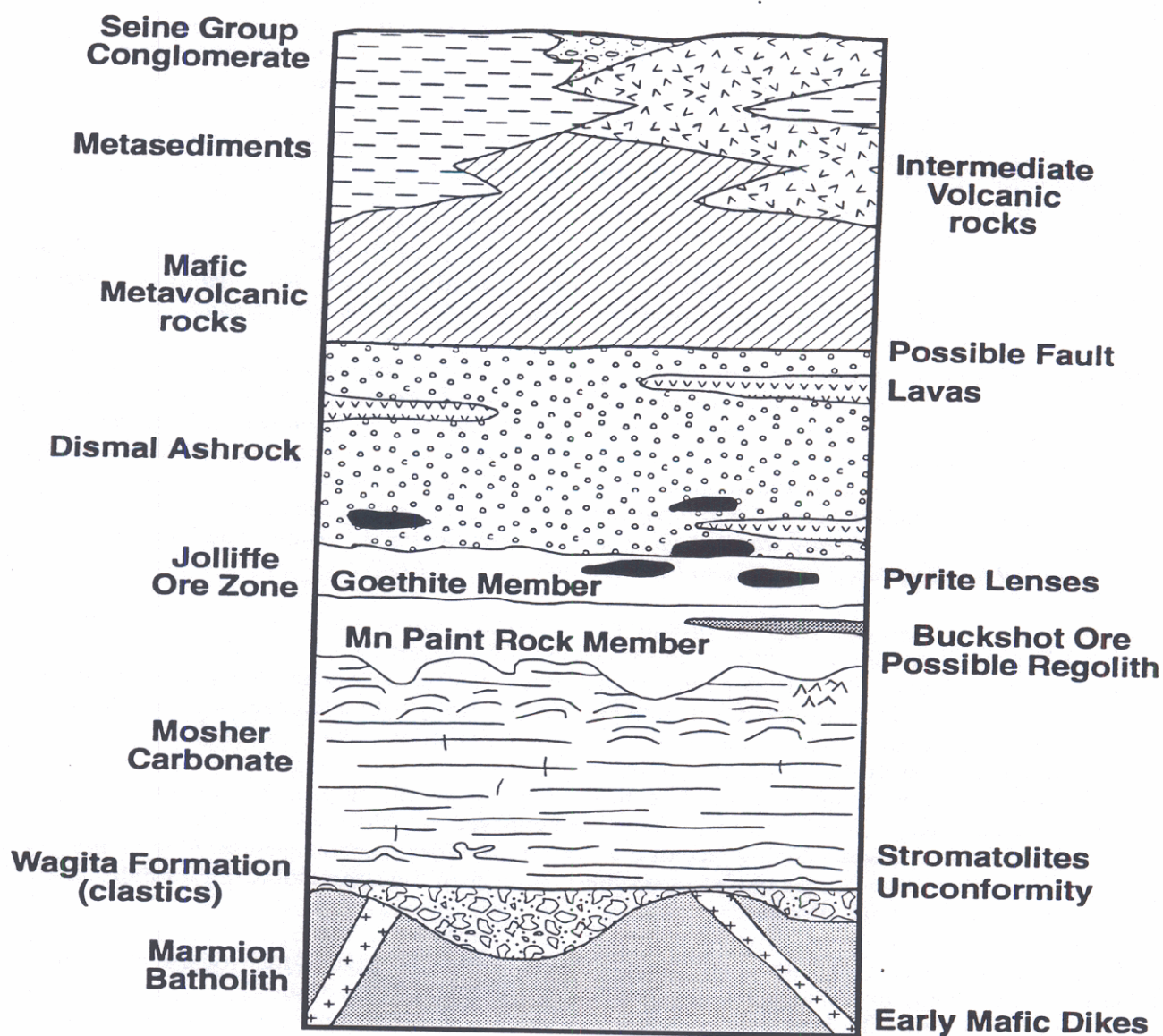


Figure 6-2: Schematic diagram (after Wilks and Nisbet 1988) showing the Steep Rock Group and contiguous supracrustal rocks of the Steep Rock belt. See legend of Figure 6-1. The upper ashrock contact is locally faulted and may represent a depositional time-gap.

small trees.

The Steep Rock Group -The Steep Rock Group was extensively studied over the past century (see reviews in Stone et al. 1992) and although early workers offered diverse structural and stratigraphic interpretations, later studies showed the Group to be a homoclinal succession of 4 formations totalling about 1000 m thickness. These include, from the base up, the Wagita Formation (conglomerate, sandstone), Mosher Carbonate Formation (limestone, dolostone), Jolliffe Ore Zone Formation (goethitic iron formation) and Dismal Ashrock Formation (ultramafic pyroclastic rocks) as shown in Figures 6-2 and 6-3. The lower boundary of the Steep Rock Group is defined by an unconformity at the base of the Wagita formation. Although sheared locally (e.g. Stops 6-2, 6-3) other exposures of the unconformity (e.g. Stop 6-5) show little deformation and preclude tectonic juxtaposition of the Steep Rock Group onto the Marmion batholith. The upper contact of the Dismal Ashrock Formation is highly deformed in the Steep Rock Mine area and is widely cited as the upper boundary of the Steep Rock Group (e.g. Jolliffe 1955).

The Steep Rock Group is cited as a prime example of a platform sequence (Thurston and Chivers 1990); platform sequences typically comprise a succession of quartz arenite and conglomerate, stromatolitic carbonates, iron formation and komatiitic rocks unconformably overlying felsic plutonic rocks. Most are concentrated in the Sachigo Subprovince and are pre 2.8 Ga in age having preceded the mafic, mafic to felsic and "Timiskaming-type" volcanic sequences that make up many greenstone belts. Geologic evidence suggests platform sequences developed in shallow water in a tectonically stable environment, involving initial erosion of widespread older sial followed by rift-related volcanism (Thurston and Chivers 1990).

Stop 6-3 Marmion tonalite; Wagita Formation; Mosher Carbonate Formation; stromatolites

The outcrop at the end of the trail is made up of buff, gritty, poorly bedded Wagita Formation and altered friable tonalite that gives way westerly over a distance of about 10 m to grey, bedded Mosher Carbonate Formation. Note that the contact between clastic metasedimentary rocks and Mosher Carbonate Formation is sharp; rocks are strongly foliated possibly due to shearing at this locality.

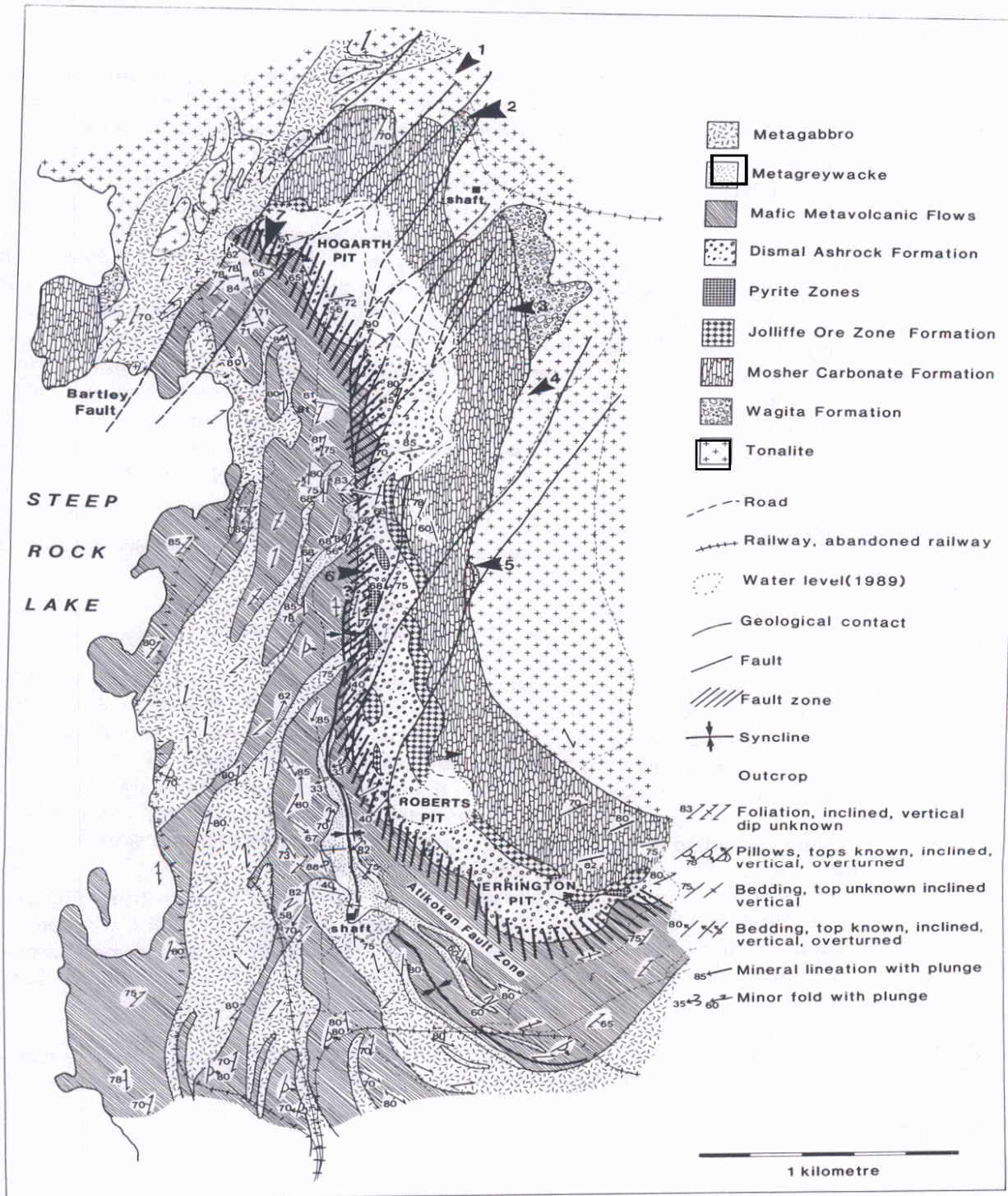
About 10 m north at the low end of the outcrop, spherical and pseudocolumnar stromatolites, typically .1 m in diameter are exposed in bedded Mosher Carbonate Formation. A west-sloping face of the outcrop provides a lateral section showing the concentric domical shape of stromatolites. Apexes of the stromatolite domes indicate a westerly younging direction.

Continue down the slope to a level area and turn west toward the edge of the Hogarth open pit. Observe large domical stromatolites exposed on the west-facing wall of the upper bench. Rather than returning to the vehicles by the way in which you came, proceed northerly from the level area along a steep access ramp to the perimeter road. Walk along the perimeter road to the vehicles and drive .1 km up the hill and park in a level place at the observation area. Walk to the west side of the observation area.

Stop 6-4 Scenic lookout and overview of the Steep Rock Group

The partly flooded Hogarth pit can be seen to the north and the Roberts pit to the extreme

Figure 6-3: Detailed geologic map of the Steep Rock Mine area showing locations of field-trip stops 6-1 to 6-7.



south. Looking west, the west arm of Steep Rock Lake can be seen beyond a concrete retaining dam.

This stop provides an overview of the west-facing Steep Rock Group and adjacent, east-facing mafic metavolcanic sequences in the Mine area (Figure 6-3). The scenic lookout is situated on Marmion tonalite at the base of the Steep Rock Group. The Wagita Formation is not well exposed and can be either absent or else thin at this locality. In either case, the base of the Steep Rock Group extends along the foot of the slope immediately in front of you. Grey outcrops underlying the tree-covered hill 200 to 300 m west, and the rusty jagged bluff to the northwest are Mosher Carbonate Formation. The Jolliffe Ore Zone Formation is almost completely flooded; a few reddish outcrops at the northwest end of the Hogarth pit may be remnants of the ore-zone. Dark grey-greenish outcrops on the west side of the pit near the centre of the mine area are Dismal Ashrock Formation. The sides of large heaps of waste rock on the west side of the mine area also appear to be mainly ashrock. The upper bench road that runs nearly full length of the west side of the mine area approximately marks the contact between Dismal Ashrock Formation and adjacent mafic metavolcanic rocks. East-facing pillow lavas and metagabbro dikes underlie most of the tree-covered ridge beyond the upper bench road. The failed slope at the northwest end of the Hogarth pit is in metagabbro and tonalite.

Proceed back down the side-road, turn right and continue 1 km down the main access ramp. Park at the junction with the first bench road to the left and walk about 300 m south along the bench road.

Stop 6-5: Unconformity; Wagita Formation; Mosher Carbonate Formation

The large bluff east of the bench road is composed of tonalite cut by thick, inclined metagabbro dikes. Near the top and at the south end of a small talus slope about 300 m from the parking area, steep-westerly dipping beds of rusty Mosher Carbonate Formation lie in unconformable contact with tonalite along the face of the bluff. Normally grey crystalline tonalite becomes buff, friable and altered possibly reflecting development of a regolith at the unconformity. Mafic minerals and plagioclase are altered to chlorite, sericite and carbonate however, the rock retains a relict igneous texture shown by unaltered quartz grains dispersed throughout the medium.

The unconformity is marked by a transition over a distance of 1 to 2 m from altered tonalite through gritty sandstone (Wagita Formation) to bedded Mosher Carbonate Formation. Quartz content increases up-section in the sandstone; quartz grains are present in the lower carbonate beds. The unconformity is relatively unfaulted at this locality.

Proceed back up the main access ramp and drive around to the west side of the mine to exposures of Dismal Ashrock in the west wall of the Hogarth pit.

Stop 6-6: Dismal Ashrock Formation

The soft, dull, dark green rock is representative of Dismal Ashrock Formation. The ashrock is composed of dark, poorly sorted, subrounded lapilli and is essentially a lapilli tuff that rarely shows stratification. Some lapilli are zoned possibly due to accretionary growth under subaerial conditions. Although Jolliffe (1955) noted thin lava flows in this unit, the ashrock is dominantly pyroclastic and in this regard is rare among Archean komatiites.

Continue north on the upper bench road along the west side of the mine area. Turn right and proceed .3 km down a pit-access ramp turning right at an ensuing junction and continue .2 km down toward the flooded Hogarth pit. Park by two power poles mounted in cement-filled tires. Walk north along the access road curving up-hill to the west. From the end of the road follow a flagged trail north over heaps of mine waste to the edge of the pit.

Supracrustal assemblages and the structure of greenstone belts - Although originally interpreted as homoclinal successions, geochronology has shown in recent years that many greenstone belts are composites of supracrustal assemblages differing in age by up to 300 Ma (e.g. Corfu and Andrews 1986; Ayres and Corfu 1991). Although assemblages are major structural components of greenstone belts, mapping of assemblages has proven difficult because many contain similar types of volcanic and sedimentary rocks and all have been affected by late metamorphism and deformation. Boundaries between assemblages tend to be poorly exposed and can be either depositional time gaps or faults with large possible displacements. In certain well-studied areas such as the Favourable Lake greenstone belt, assemblage boundaries appear to be mainly thrust faults (Ayres and Corfu 1991).

At the present locality, the boundary between the Dismal Ashrock Formation and contiguous mafic metavolcanic rocks could possibly be an assemblage boundary. Favourable evidence includes the major lithologic and structural transition from the west-facing Steep Rock Group to a thick sequence of east-facing mafic lava flows and gabbro. The boundary is faulted in the mine area (Figure 6-3; Stop 6-7), which implies that the mafic sequence could have been tectonically juxtaposed with the Steep Rock Group. Since neither the Steep Rock Group or metavolcanic rocks of the Steep Rock belt are dated, the depositional time gap at their mutual contact is unknown.

Stop 6-7: Deformation at the contact between the Dismal Ashrock Formation and mafic metavolcanic rocks

The pale green to white outcrops west of the upper bench road are pillowed mafic flows. East of the flooded Hogarth pit are bluffs of rusty Mosher Carbonate Formation. Approximately where the trail ends, the Steep Rock Group curves abruptly west and is cut off by a large metagabbro dike and the Bartley fault at the northwest end of the Hogarth pit. Metagabbro and tonalite make up most of the large bluff beyond the northwest end of the pit; a large rock slide in this area forced the mine to close prematurely in 1979.

Provided that the slope of waste rock to the north is not flooded, carefully move down the

slope. Observe the pale-green to white friable rock that shows good pencil structure in the bench wall to your left. The pale green rock is probably a deformed mafic metavolcanic flow. Moving farther down the slope you pass beside soft, crumbled and crenulated, variably rusty to dark green rock of uncertain type. *Upon reaching the lower bench level, go west along the bench to a point where it is overrun with talus.* The dark-green deformed rocks in the pit wall behind the bench are Dismal Ashrock Formation and metagabbro. A rusty zone, which is possibly a pyrite lense, can be seen high on the pit wall. Several subvertical, north-trending mesoscopic faults cut up through the ashrock and are eroded back into the pit wall.

The area illustrates severe deformation at the upper ashrock contact in the mine area. Mineral lineations plunge down-dip westerly implying a dominant component of dip-slip faulting.

Proceed back up the ramps to the upper bench road and go south on to the paved perimeter road in the direction of Atikokan. Turn right on the road leading to the seaplane base on Steep Rock Lake and continue up a steep hill and around several curves for .4 km. Park in a turn-out on the left side of the road and walk east about 50 m through small trees to low outcrops in an abandoned gravel pit.

Stop 6-8 (optional) Pillowed mafic metavolcanic flows

This outcrop lies within the Canadian Chareleston Gravel Pit that was strip-mined for iron-ore gravel in 1960 to 1965.

Pillowed mafic metavolcanic flows that underlie the western half of the Steep Rock belt are represented in this outcrop. Pillows typically have a pale-green, homogeneous interior and very-fine-grained, dark-green rims. Several varieties of pillows ranging from large "mattress" to small spherical types occur together. The younging direction derived from pillow shapes in this outcrop and elsewhere in the western Steep Rock belt is east. A massive flow occurs at the extreme west end of the outcrop. The rock is calc-alkaline basalt.

Proceed back to Atikokan and south to the junction of Highway 11B and Highway 11. Turn west on Highway 11 and continue for 1.2 kilometers turning north on a forest access road just before Kemuel Lake. Drive about 200 m into a clear-cut and reforested area, park at the first turn-out and walk to a low outcrop on the east side of the road.

Stop 6-9 (optional) Quetico metasedimentary rocks

This outcrop shows several types of bedding and sedimentary features. Thick beds (up to 1. m) of fairly homogeneous metawacke occur at the north side of the outcrop. Bed thickness is generally .05 to .2 m; many beds are graded from a coarse sandy base to a grey, laminated, silty top. Truncated beds and crescentic scour channels filled with coarse material are present. Clasts (up to .2 m) of fine-grained, white chert or possibly

felsic volcanic material fill some scoured channels. Grainsize gradation and scour channels indicate that the local younging direction is south. The outcrop is cut by quartz veins, fractures and small-displacement faults. The mafic mineral is mainly biotite. Metamorphic grade increases south with the appearance of amphibole and garnet about 1. km distant.

End of Day 2. Return to Atikokan or Thunder Bay.

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