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Thirty-First Annual Meeting**

**KENORA, Ontario
1985**



**FIELD TRIP
GUIDEBOOK**

INSTITUTE ON
LAKE SUPERIOR GEOLOGY
31st ANNUAL MEETING
KENORA, 1985

FIELDTRIP GUIDEBOOK

edited by
G.P. Beakhouse

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various aspects of the reproduction of this volume.

COVER PHOTO: Deformed mafic dike cutting 2.85 Ga
gneissic tonalite in the Wabigoon-Winnipeg River
subprovincial interface zone, Keewatin, Ontario

Preface

Kenora, host of the 31st Annual Institute on Lake Superior Geology, is situated at the north end of Lake of the Woods. It is especially appropriate that Kenora host the 1985 meeting on the centenary of Andrew Lawson's first publication on the Lake of the Woods area; he identified many of the problems we are debating today and established the area as a classical Archean terrane. A brief summary of the regional geologic setting of the Kenora area, based on the discussions in the field guides and the references cited therein, follows.

Kenora is located within the southwestern part of the Superior Province of the Canadian shield. The Superior Province has been subdivided into east-west trending subprovinces consisting of greenstone-tonalite terranes (Shebandowan-Wawa, Wabigoon, Uchi and Sachigo Subprovinces), metasedimentary gneiss terranes (Ear Falls - Manigotagan and Quetico Subprovinces) and batholithic terranes (Winnipeg River and Berens River Subprovinces). The town of Kenora lies on the interface between the Winnipeg River and Wabigoon subprovinces.

The western Wabigoon Subprovince comprises a supracrustal assemblage of predominant metavolcanic and subordinate metasedimentary rocks and granitoid intrusions, some of batholithic dimensions. Stratigraphic analysis of the rock sequences within the western Wabigoon Subprovince suggests that there was a consistent evolution through time with initial, quiescent, submarine, mafic, tholeiitic volcanism followed by more silicic, submarine and subaerial, calc-alkaline pyroclastic volcanism and related clastic sedimentation. More than one such cycle may be present. Available high precision U-Pb zircon geochronology suggest that volcanism and plutonism were synchronous and took place over a relatively short time interval between 2755 and 2695 Ma.

The metallogeny of the western Wabigoon Subprovince is diverse. Massive sulphide deposits such as those at South Sturgeon Lake and numerous gold deposits have received the most attention. Exploration for gold mineralization has recently intensified in the wake of a significant discovery at Cameron Lake. Other types of potentially economically significant deposits include disseminated copper and/or molybdenum, rare metal pegmatites, building stone and industrial minerals.

Our perception of the terrane lying to the north of the Wabigoon Subprovince has changed dramatically within the past ten years. This area, referred to as the English River Subprovince, is now known to consist of two different types of terrane for which separate subprovince status has been proposed; a northern supracrustal domain (Ear Falls-Manigotagan Subprovince) and a southern plutonic domain (Winnipeg River Subprovince). The Ear Falls-Manigotagan Subprovince consists predominantly of turbiditic metasedimentary rocks which have been highly metamorphosed and partially melted. The Winnipeg River Subprovince contains three principal rock suites; pre-tectonic gneissic granitoid rocks, pre- to syn-tectonic sodic plutons and late- to post-tectonic potassic plutons. Subordinate, amphibolitic to

granulitic, predominantly metavolcanic, supracrustal inclusions are present in all three suites but are most abundant in the gneissic rocks.

In contrast to the tightly clustered, relatively young ages of the Wabigoon Subprovince, Winnipeg River Subprovince plutonism spanned 600 Ma (2560 to 3168 Ma). Many of the younger rocks appear to represent reworking of older rocks. These relationships have been interpreted in terms of the Winnipeg River Subprovince representing a basement complex that existed prior to Wabigoon Subprovince volcanism and plutonism. The interface between the two subprovince has stratigraphic (unconformable), tectonic, metamorphic and intrusive aspects and defies a simplistic interpretation.

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The Cameron Lake Deposit

Introductory Discussion and Field Guide
31st Annual Institute on Lake Superior Geology
Kenora, Ontario

by

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INTRODUCTION

The Cameron Lake gold deposit is located in northwestern Ontario about 80 km south-southeast of Kenora. It was originally discovered in 1960 by two prospectors working for Noranda Mines (Hunter et al., 1984). Low gold prices at that time coupled with a complex subsurface geometry and poor outcrop exposure discouraged investment and development. Four separate diamond drilling programs were conducted on the property prior to the discovery of a major zone of gold mineralization in 1983 by two junior mining companies, Nuinsco Resources and Lockwood Petroleum. Proven reserves at that time totaled 1,287,000 tons grading 0.154 oz/ton Au, including 807,000 tons grading 0.194 oz/ton Au (Nuinsco Resources Limited, 1983).

REGIONAL GEOLOGIC SETTING AND ITS BEARING ON GOLD MINERALIZATION

Cameron Lake lies in the western portion of the Archean Savant Lake-Crow Lake metavolcanic-metasedimentary belt (Trowell et al., 1980). The region is divided geologically by the northwest-southeast trending, northeast-dipping Pipestone-Cameron Fault (Figure 1). Southwest of the fault lies the north- to east-facing, Kakagi Lake greenstone terrain (KLGT) (Schwerdtner et al., 1979) and to the northeast the south-facing Rowan Lake greenstone terrain (RLGT). The fault is a major zone of deformation and displacement which strikes over 100 km in a northwest-southeast direction. Exposure of the fault in the Cameron Lake area is poor, often being obscured by lakes and low swampy depressions. Correlation of lithologies across the fault, although suspected, has not been possible (Trowell et al., 1980).

The geology of the RLGT is dominated by the Shingwak Lake Anticline. The fold has a east-northeast trending axial surface and plunges steeply towards the southwest (Kaye, 1973). Based on field mapping and lithogeochemical analyses, Blackburn and Hailstone (1984) have shown that the RLGT may be subdivided into the Rowan Lake and Cameron Lake Volcanics. All rocks have been metamorphosed to greenschist-facies assemblages. The Rowan Lake Volcanics form a thick predominantly pillowed, subaqueous mafic flow succession in the core of the Shingwak Lake Anticline. These rocks are overlain by the Cameron Lake Volcanics, a mixed succession of subaqueous pillowed and massive mafic flows, and intermediate to felsic pyroclastics. South of the fold axis top determinations define a steeply dipping homocline facing southward.

Subvolcanic gabbroic sills were emplaced at all levels in the stratigraphy prior to folding. Felsic porphyry sills and dikes, and quartz monzonite stocks intrude all the earlier strata.

Blackburn and Hailstone (1984) have shown that the lower succession (Rowan Lake Volcanics) is tholeiitic and mostly rather magnesium-rich, while the upper succession (Cameron Lake Volcanics) includes both tholeiitic to calc-alkaline varieties. The most significant gold occurrences are within the Cameron Lake Volcanics (Blackburn and Janes, 1983) and tend to be associated with zones of high strain. Several of these occurrences are situated at or near gabbroic-volcanic contacts. Two such gold-bearing, high strain zones, previously unnamed, are referred to

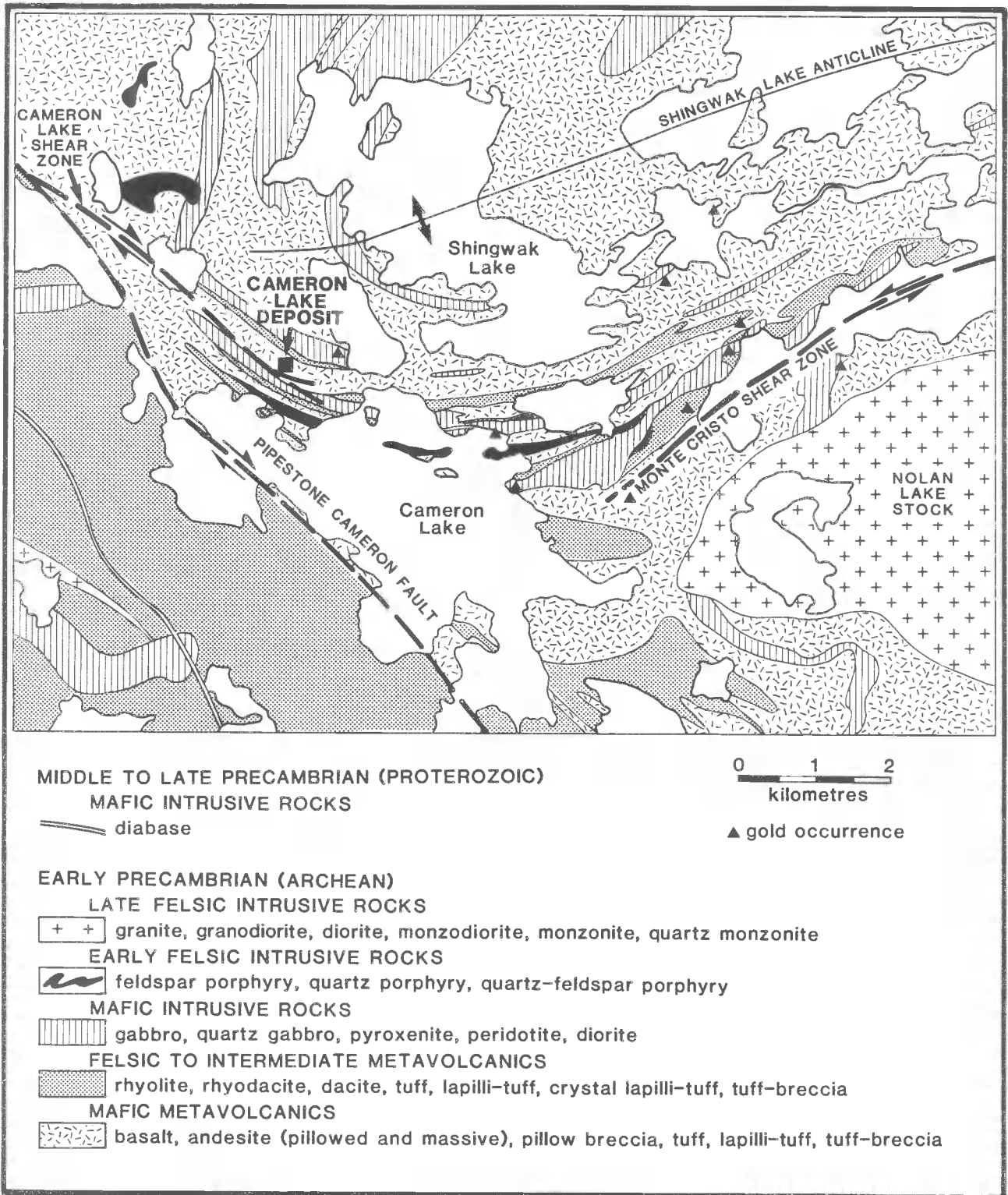


Figure 1: Geology of the Cameron Lake area (modified after Kaye, 1973).

here as the Cameron Lake and Monte Cristo Shear Zones. These are presently the principal targets for exploration.

THE GEOLOGY OF THE CAMERON LAKE GOLD DEPOSIT

The Cameron Lake deposit is situated on the north shore of Cameron Lake (Figure 1) about 700 m northwest of the point of debarkation at the Nuinsco bush camp. The rocks hosting the deposit are composed of mafic volcanics consisting predominantly of fine grained pillowed and massive flows with minor occurrences of pillow breccia interlayered with intermediate to felsic pyroclastic rocks. Two gabbroic sills intrude the intermediate and felsic pyroclastic rocks. Quartz-feldspar porphyry sills and dikes are also present.

The Cameron Lake gold deposit is located within the Cameron Lake Volcanics near the contact with the underlying Rowan Lake Volcanics. It occurs in sheared mafic volcanic rocks adjacent to the highly deformed intrusive contact with a gabbroic sill. The shear zone hosting the Cameron Lake gold deposit is a brittle-ductile, dextral strike-slip system with an attitude of $315/70^{\circ}\text{NE}$ and a maximum strike length of over 1,000 m. Its width varies from 11 m to 60 m.

Two planar fabrics may be seen within the zones of high strain; a cleavage parallel to the shear zone, and a planar mineral alignment (flattening plane foliation) inclined at angles less than 45° to the shear zone. The intersection of these foliations defines a lineation which pitches 85°E within the shear zone. This lineation is perpendicular to the inferred displacement vector which pitches 50°W within the shear plane.

Three distinct veining episodes have been recognized in the carbonatized rocks of the Cameron Lake Shear Zone:

- stage i) early, barren, extensional, quartz-carbonate veins which have been buckled;
- stage ii) a major system of gold-bearing, pyritic, quartz-rich breccia-veins ($315/70^{\circ}$);
- stage iii) a late group of cross cutting en echelon extension veins ($360/90^{\circ}$) which are composed of quartz-carbonate-chlorite-hematite-gold.

The significance of the stage (ii) breccia-veins has been previously underestimated. They represent a specific gold-bearing rock type at Cameron Lake which commonly contains greater than 0.3 oz/ton Au. Pervasively altered carbonate-sericite-pyrite rich phyllonitic rocks which occupy a large volume of the Cameron Lake Shear Zone and envelope the breccia-veins are lower grade. This subdivision of gold-bearing rock types is readily visible in surface exposures of the deposit and diamond drill core.

The breccia-veins are characterized by angular, often elongate fragments of altered basalt which range from 1 mm to greater than 30 cm in size. Breccia fragments may be in contact or separated by a vein matrix (10 to 50%) of quartz, carbonate, and plagioclase. Commonly groups of fragments can be fitted together like pieces in a jigsaw puzzle. Pyrite rarely exceeds 5% and is concentrated within the fragments and forms narrow

envelopes surrounding the breccia-veins. Gold has been seen in hand sample and polished thin section as inclusions within pyrite grains and as free gold within the matrix interstitial to pyrite grains. Minor (less than 1%) chalcopyrite and rutile are also present.

The transition from cleavage development to brittle processes within the Cameron Lake Shear Zone is manifested in the formation of veins and vein-breccias. The change from ductile to brittle behavior may be the result of local increases in hydrothermal fluid pressure which facilitated brittle deformation and vein formation by reducing the effective normal stresses. The phyllonitic rocks formed a sink for the gold-bearing fluids, while the breccia veins formed major conduits enhancing fluid circulation.

The localization of highly anomalous quantities of gold along the Cameron Lake Shear Zone is related to the development of oblique, secondary, bedding-controlled sympathetic shear zones (275/70°N). This is illustrated on a dipping longitudinal section compiled from drilling (Figure 2). Note the 70°W pitch of the gold shoots, and the periodicity of the high grade zones. Figure 3 is a simplified stereographic projection of the Cameron Lake Shear Zone and the oblique, bedding-controlled, sympathetic splays. The pitch of the deposit, as defined by drilling, is coincident with the intersection of the these two shear systems. The relationship between the two shear systems and the deposit is illustrated diagrammatically in Figure 4. It serves to explain the complex subsurface geometry of the gold shoots, as well as the existence of hanging wall and footwall zones on either side of the main zone of gold enrichment. At this stage the periodicity of the high grade gold pods can not be adequately explained.

FIELD TRIP STOPS

At this stop we will examine excavated exposures (Figure 5) of the gold-bearing rocks typical of the Cameron Lake deposit. Emphasis will be placed on the following aspects of the deposit:

- 1) Kinematic indicators useful in the determination of sense of shear (cleavage, lineation, rotation and folding of passive and nonpassive markers, minor folds, enechelon veins);
- 2) Vein paragenesis and its relation to alteration, gold deposition and deformational processes within the shear zone;
- 3) Shear zone geometry and its influence on gold deposition and exploration methods.

STOP A. Discovery Outcrop: This is the locality where the original gold showing was discovered. The outcrop is within the main shear system and consists primarily of pervasively carbonatized pillowed basalt, the least deformed of which indicates an overturned succession topping toward the south. At the southern edge of the exposure is the sheared intrusive contact with the

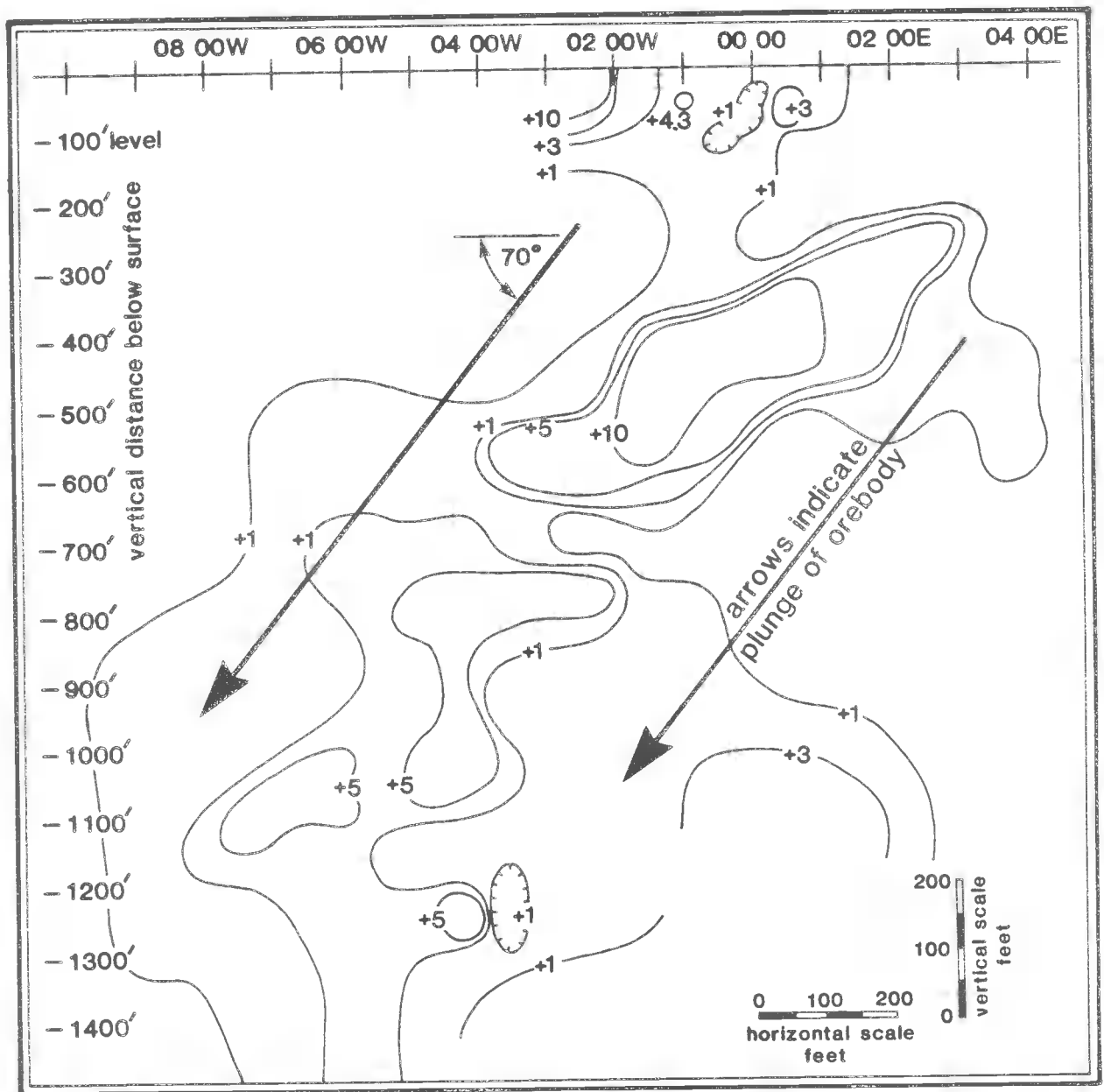


Figure 2: Longitudinal section dipping 65°N from the baseline. Contours represent the value of grade times thickness (eg. 0.2 oz/ton Au over 50 feet = 10). The contours are based on approximately 75 drill hole piercing points.

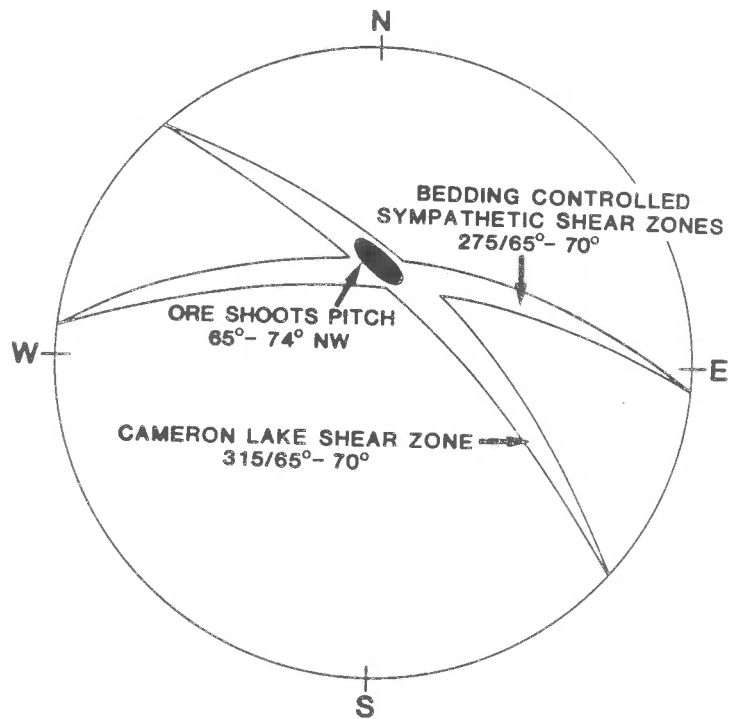


Figure 3: Simplified stereographic projection illustrating the coincidence of the pitch of ore shoots with the line of intersection of the Cameron Lake shear zone and bedding controlled shear zones.

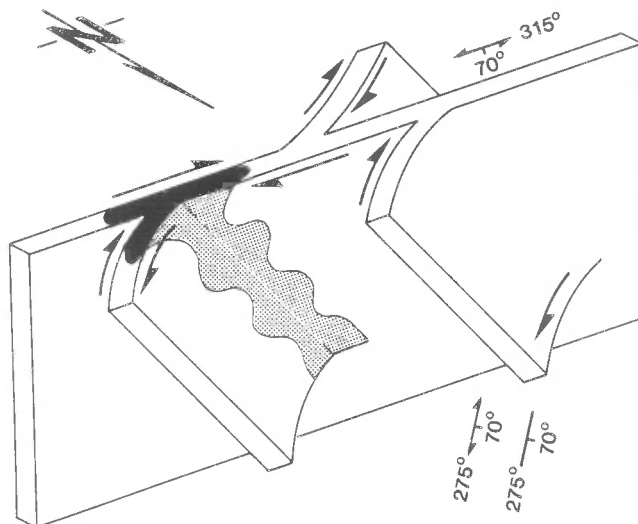


Figure 4: Diagrammatic representation of the spatial relationship between ore shoots and the north-west trending Cameron Lake shear zone and north-east to east trending bedding plane shears.

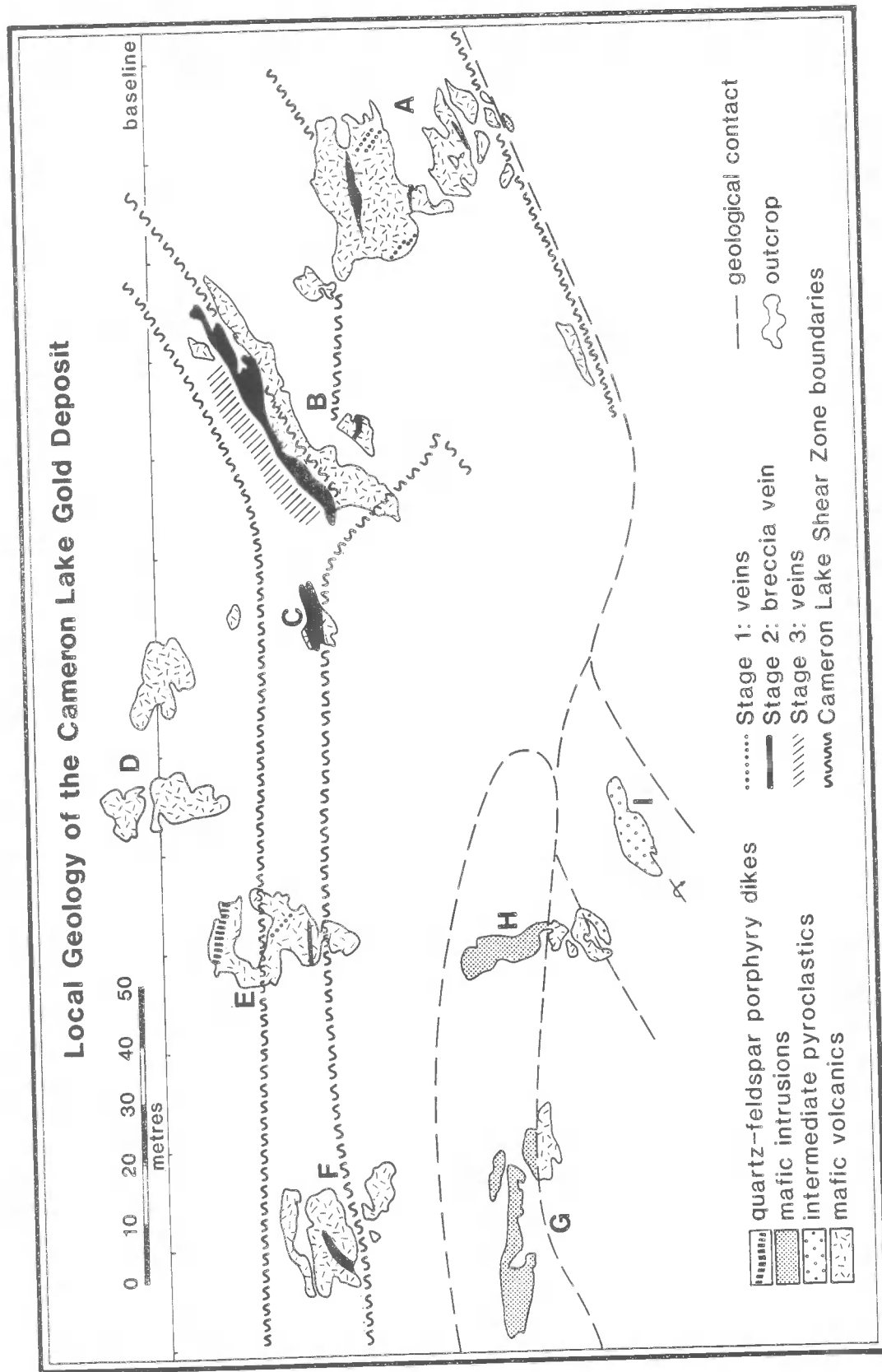


Figure 5: Local geology of the Cameron Lake gold deposit with the locations of outcrops discussed in the text.

footwall gabbro sill. To the north is a thin interflow sedimentary unit (passive marker) which has been folded into the plane of the shear zone. Several early, stage (i) extension veins (nonpassive markers) are present in this exposure. All of these have been shortened and buckled, commonly displaying Z-fold geometry. Three stage (ii) gold-bearing breccia-veins are present, one of which illustrates a distinct brecciated texture and geometric fit of fragments. Note how rapidly these types of veins pinch and swell.

STOP B. North Trench: This outcrop, excavated in 1984, is critical in terms of understanding the geometry of the deposit. At the western end of this exposure the intersection between the main shear system (Cameron Lake Shear Zone) and the bedding-controlled splays may be clearly seen. It is at this intersection where the most dilatant, permeable conditions were developed which facilitated the migration of the gold-bearing hydrothermal fluids. The line of intersection defines the pitch of the gold shoots as defined by drilling for over 1000 m. The large, stage (ii), breccia vein here averaged 0.273 oz/ton Au over 22 channel samples. The breccia-vein is cut by an array of north-striking, late, stage (iii), en echelon extension veins. Note the barren, unaltered bull's nose structure developed east of the confluence of the sympathetic, bedding-plane shear and the main structure.

STOP C. Vic's trench: The extension of the large breccia-vein seen in the last outcrop is exposed in this trench but its strike is now parallel to that of the main shear system. To the south of the breccia-vein occur relatively unaltered and undeformed mafic volcanics and to the north, gold-bearing pervasively altered, phyllonitic rocks.

STOP D. NC-19 trench: This excavated exposure consists of relatively unaltered basalt cut by zones of cleavage which provide a small scale, but barren example of the intersection of two shear systems similar to that seen in the north trench. Note the collar of DDH NC-19, drilled in 1981, which intersected 31.6 feet of 0.327 oz/ton Au (up dip extension of which is exposed in Vic's trench). Two distinct cleavages and the lineation produced by their intersection may be seen in this outcrop.

STOP E. Island Trench: This outcrop consists of pillowed basalt cut by a small quartz-feldspar porphyry dike typical of those intersected during drilling. Although these dikes are commonly altered, they may postdate the formation of the gold deposit. Note the early, stage (i) extension vein which has been buckled and rotated, and the fine-grained siliceous, stage (ii)

breccia-vein enveloped by a pyritic alteration halo.

STOP F. Blood trench: At this location we will examine exposures of weakly altered and cleaved massive and pillowed basalts. This outcrop, interpreted to occur within the main shear system, does not display the pervasive iron staining produced by the weathering of carbonates characteristic of the outcrops to the east. However, even at this distance (approximately 100 feet along strike) from the centre of the deposit, isolated stage (ii) breccia-veins and areas of disseminated pyritic alteration are found suggesting that the potential for additional economic gold occurrences along the Cameron Lake Shear Zone is high.

STOP G. Gabbro trench: In this trench the gabbroic sill which occurs in the footwall to the deposit is exposed. Note the presence of quartz eyes. Carbonate alteration and incipient shearing occur on the northeastern edge of the outcrop indicating proximity to the footwall shear zone and the mafic volcanic hosts to the deposit. On the southeast edge of the outcrop is a contact between the gabbro and mafic volcanics. This exposure is interpreted as a small lobe of gabbro related to the larger (over 100 foot thick) gabbro sill located immediately to the south.

STOP H. Hunter's trench: This exposure of sheared and altered "quartz-eye" gabbro is considered representative of the barren footwall shear zone encountered during drilling. Note the development of a green mica associated with the carbonate alteration. This green mica is tentatively interpreted as a chromium-bearing phyllosilicate resulting from the alteration of the gabbro which contains an unusually high primary chromium content. On the southern margin of the outcrop are contacts with both unaltered pillowed basalt and sheared intermediate pyroclastics.

STOP I. Controversy trench: This outcrop consists of intermediate pyroclastics and fine-grained bedded sediments. Graded bedding indicates an overturned succession facing south. The strike of this unit is the same as the bedding-controlled, sympathetic splay exposed in the north trench which controls the ore shoot geometry. Note the thin mafic dikes intruded along bedding planes.

ACKNOWLEDGEMENTS

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REFERENCES

- Blackburn, C.E. and Janes, D.A., 1983, Gold Deposits in Northwestern Ontario; p.194-210, in *The Geology of Gold in Ontario*, edited by A.C. Colvine, Ontario Geological Survey Miscellaneous Paper 110, 278p.
- Blackburn, C.E. and Hailstone, M.R., 1983, *The Geological Environment of Gold Mineralization, Cameron-Rowan Lakes, N.W. Ontario*; Paper Presented at the Geoscience Research Seminar, December 6-7, 1983, Toronto, Ontario, 13p.
- Hunter, A.D., Curtis, L.W., and Melling, D.R., 1984, *The Cameron Lake Gold Deposit, N.W. Ontario: Pioneering in a Dormant Gold Camp*; (Abstract) CIM Bulletin, Volume 77, Number 863, p.60.
- Kaye, L., 1973, *Rowan Lake Area, District of Kenora*; Ontario Division of Mines, Preliminary Map P.831, Geol. Ser., scale 1 inch to 1/4 mile. Geology 1972.
- Nuinsco Resources Limited, 1983, *Annual Report to Shareholders*; 16p.
- Schwerdtner, W.M., Stone, D., Osadetz, K., Morgan, J. and Stott, G.M., 1979, *Granitoid Complexes and The Archean Tectonic Record in The Southern Part of Northwestern Ontario*; Canadian Journal of Earth Sciences, Volume 16, p.1965-1977.
- Trowell, N.F., Blackburn, C.E. and Edwards, G.R., 1980, *Preliminary Synthesis of the Savant Lake-Crow Lake Metavolcanic-Metasedimentary Belt, Northwestern Ontario, and its Bearing upon Mineral Exploration*; Ontario Geological Survey, Miscellaneous Paper 89, 30p. Accompanied by Chart A.

Geologic Setting and Style
of Gold Mineralization in
the Lake of The Woods Area

Introductory Discussion and Field Guide
31st Annual Institute on Lake Superior Geology
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REGIONAL GEOLOGIC SETTING

INTRODUCTION

Recent geological syntheses (Trowell et al 1980; Blackburn et al., 1985) of the western Wabigoon Subprovince of northwestern Ontario have led to the general conclusion that within the "greenstone" belts there is a general chronologic progression from mafic, platformal volcanic sequences upward into intermediate to felsic, more pyroclastic, volcanic sequences. Clastic and chemical sediments are commonly associated with the upper sequences. These syntheses have in general been carried out east of Lake of the Woods, where continuity of recent 1/4 mile geologic mapping by the Ontario Geological Survey has enabled stratigraphic correlation between individual areas. In the northern part of Lake of the Woods (Figure 1), two areas in which detailed mapping has been done, one in the northwest, and centred on Shoal Lake, and the other in the northeast around Bigstone Bay, are separated by a central area that is presently unmapped. Within each of these two areas the general sequence outlined above has been demonstrated (Blackburn and Janes 1983): correlation of sequences between the two areas has not yet been conclusively demonstrated.

In the northwest, the east-trending Crowduck Lake-Rush Bay fault zone separates two volcanic sequences that can be tentatively correlated, notwithstanding lack of geochronologic information. At Shoal Lake a lower mafic, tholeiitic volcanic sequence occupies the core of the Gull Bay-Bag Bay anticline, and is succeeded upward on either limb by intermediate to felsic pyroclastic sequences. At High Lake, north of the fault zone, a lower mafic sequence is overlain unconformably by a clastic sedimentary sequence (Crowduck Lake Group).

In the northeast, around Bigstone Bay, recent mapping has demonstrated that a lower sequence of mafic, tholeiitic volcanics, that lie in the core of a broad southwest plunging antiform, are overlain by a felsic to intermediate pyroclastic sequence (Ayer et al., 1985). Further to the north in the vicinity of Kenora, a mafic volcanic sequence that is probably correlative with that at Bigstone Bay, has been isoclinally folded about northeast trending axes.

Following are more detailed descriptions of individual sub-areas, all except one of which will be visited on the field trip: the Shoal Lake area is included because of its key situation in understanding Lake of the Woods stratigraphy, and the position of gold emplacement within it.

HIGH LAKE

General geology, adapted from Davies (1965), is illustrated in Figure 2. The oldest rocks, a thick sequence of basaltic flows and mafic subvolcanic intrusions, have been intruded by tonalite, granodiorite, porphyritic granodiorite and by porphyry dikes. The granitic complex extends west for at least 50 km and, especially in Manitoba, the dominant phase is characterized by

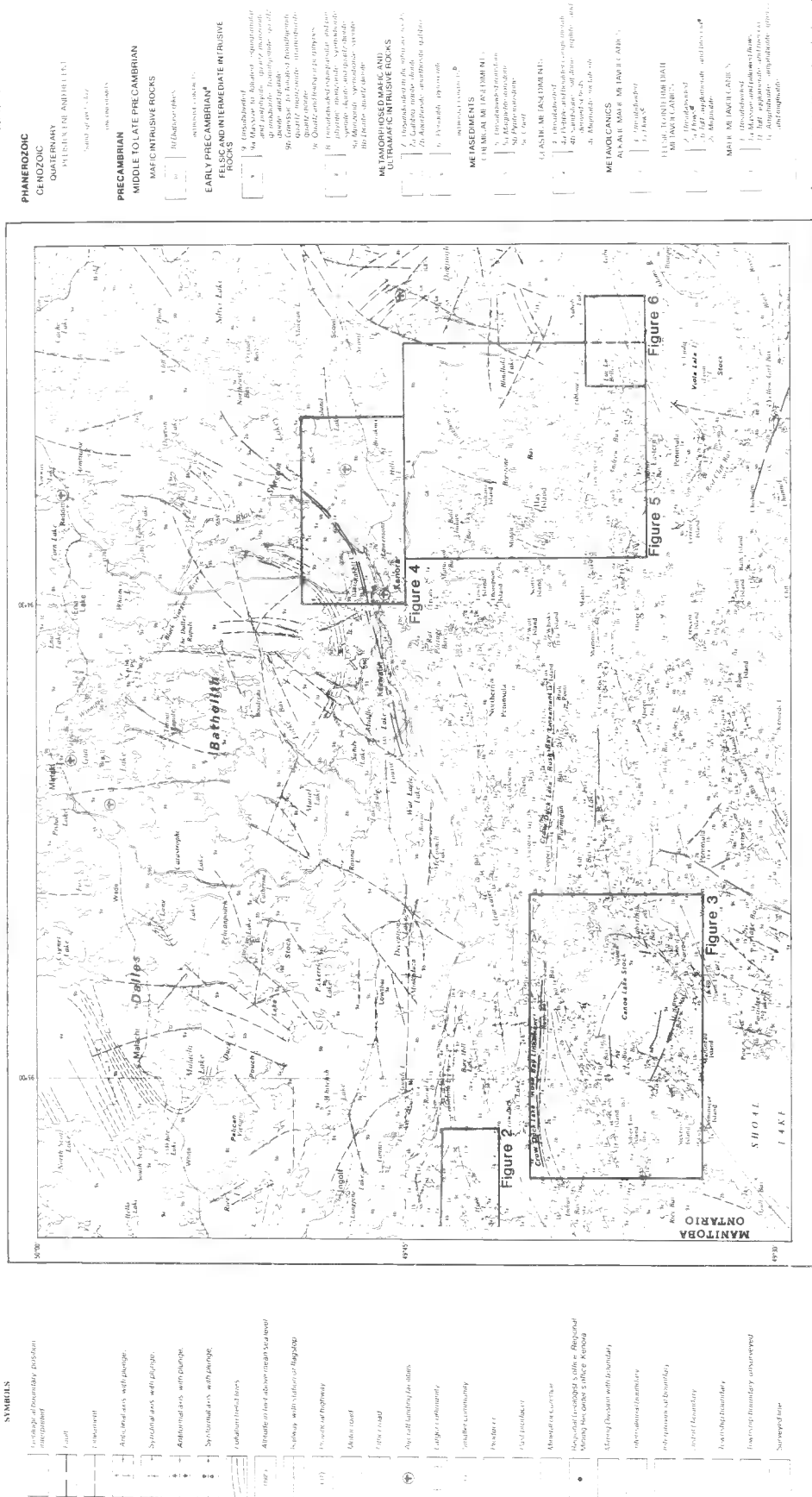


Figure 1: General geology of the Lake of the Woods Area (modified from Blackburn, 1981).

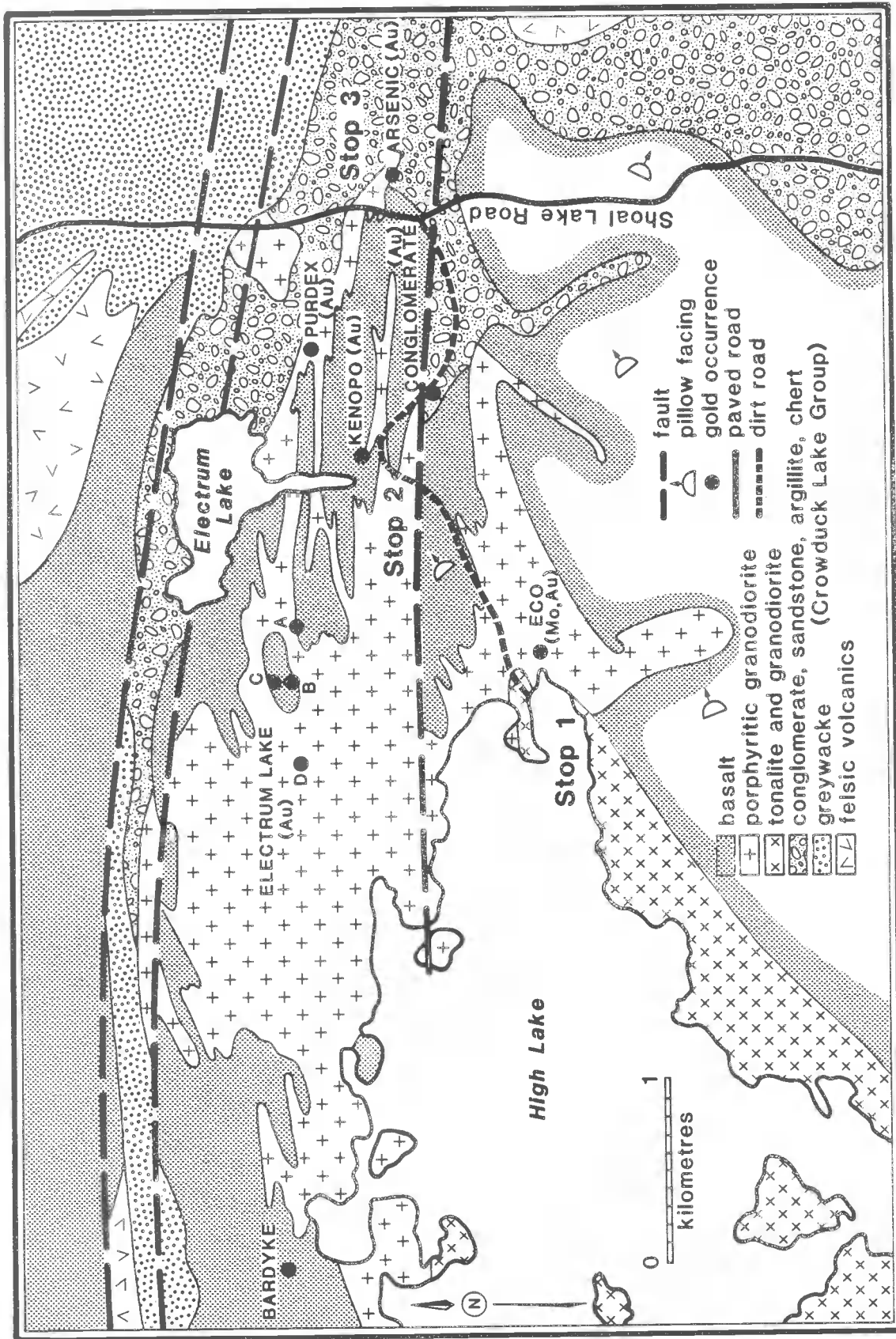


Figure 2: Geology of the High Lake area (adapted from Davies, 1965).

euohedral microcline phenocrysts.

At High Lake, large basaltic blocks are enclosed in and cut by the porphyritic granodiorite and related dikes, and are interpreted to be roof pendants in a subvolcanic intrusion. The non-porphyritic granodiorite locally contains abundant, small, round mafic inclusions; it is typically less deformed than the porphyritic granodiorite, suggesting it is younger, intruding and deforming the porphyritic granodiorite. Sutherland and Colvine (1979) concluded that, although it is possible that the two were intruded in distinct events, the similar composition of both rocks suggests that they were more likely separate pulses of the same intrusive event.

Both the basalt and the porphyry are overlain unconformably by sediments of the Crowduck Lake Group. Conglomerate, sandstone, greywacke, argillite and chert have an apparent thickness of 2000 metres.

Tight folding of both the sediments and felsic volcanics is associated with faulting and shearing that extend into the granodiorite. The over-all pattern of both folds and faults indicates dextral movement along the east-trending faults. Within the porphyritic granodiorite a second set of fractures trends east-northeast.

Mineralization is largely confined to fractures in the porphyritic granodiorite and to shears in both the adjacent basalt and the overlying conglomerate. Molybdenite and chalcopyrite, in trace amounts, are widely distributed in the porphyritic granodiorite; they are more highly concentrated, with quartz, in shear zones adjacent to east-trending faults, and are accompanied by minor gold. Gold is also associated with pyrite, chalcopyrite and pyrrhotite in shears at or near the contact with basalt, and in quartz lenses and irregular masses in porphyry, basalt or conglomerate where there are complex areas of competency contrast. All these mineralization types will be illustrated by field trip stops (Stops 1, 2 and 3). Pedora (1976) has suggested that mineralization is arranged about the southern, non-porphyritic granodiorite phase in a zonal pattern, and that the mineralizing fluid may have originated from the southern phase.

SHOAL LAKE

General geology, adapted from Davies (1983), is illustrated in Figure 3. Tholeiitic basalts, interlayered with mafic and ultramafic sills and minor komatiitic basalts, are overlain by calc-alkaline intermediate to felsic pyroclastics. Overlying these is an upper sequence of mafic flows with some felsic volcanic lenses. A sill-like body of anorthositic gabbro to diorite (the Stevens Island Diorite) lies near the top of the tholeiitic sequence. The thickness of the northwest-facing volcanics is estimated to be 9000 metres.

The Canoe Lake stock of altered quartz diorite terminates, to the southwest, against the tholeiitic basalts, and a major anticline extends southwest from it. The fresher Snowshoe Bay granodiorite intruded the intermediate to felsic sequence; it continues an unknown distance into Manitoba and is of batholithic

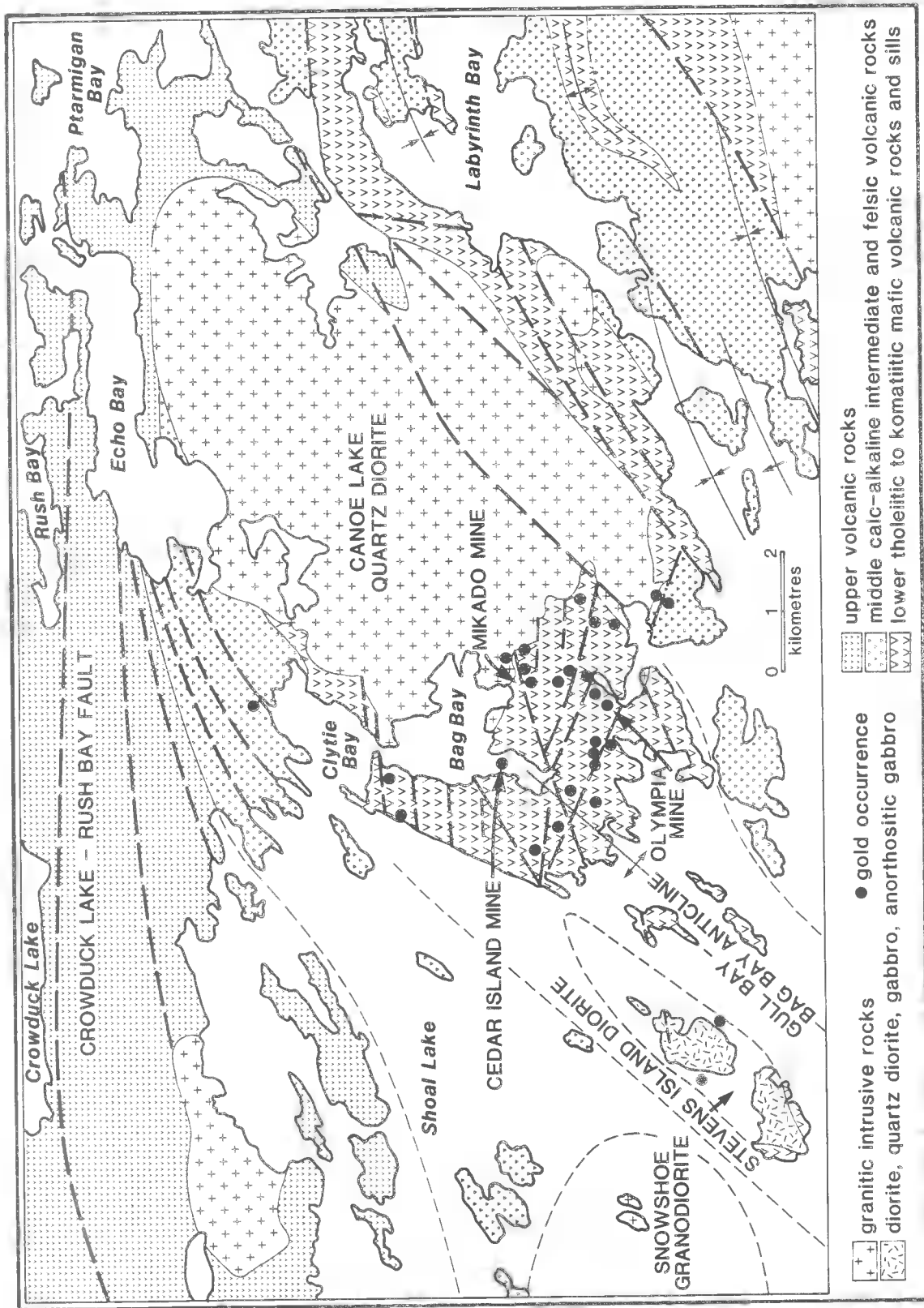


Figure 3: Geology of the Shoal Lake area (modified from Davies and Smith, 1984).

dimensions.

In much of the Shoal Lake area the principal foliation direction is east-northeast. North of the lake the foliation bends east as the Crowduck Lake-Rush Bay dextral fault zone is approached. Within a "shadow zone" southwest of the Canoe Lake stock, the mafic volcanics are not foliated, but a series of east-southeast faults with a large vertical component of movement has offset the volcanic stratigraphy. These faults are interpreted to be related to emplacement of the Canoe Lake stock. Faults with vertical displacement have also been interpreted to coincide with the north and south contacts of the Snowshoe Bay pluton; some faulting parallel to volcanic stratigraphy may also be attributed to emplacement of the pluton.

Most gold occurrences are in the tholeiitic sequence. Those associated with the east-southeast faults typically consist of a chloritic shear zone within which a felsite dike occurs. Silicification, commonly in the form of quartz veins or lenses, is accompanied by pyrite, traces of base-metal sulphides, and rare visible gold. Some gold was produced from fracture zones of this type at the Olympia and Cedar Island Mines.

South-southeast fractures developed near the margin of the Canoe Lake stock following its intrusion. At the Mikado Mine, gold is associated with quartz veins and stringers in such a fracture which crosses basalt and a thick dike of quartz diorite. Pyrite, chalcopyrite, tetradymite and bismuthinite also occur in the quartz.

Fault and shear zones parallel to volcanic stratigraphy are mostly narrow and quartz veining within them is similarly narrow and discontinuous. The fault at the Duport Mine is much wider, and gold occurs with quartz in zones where competency contrast accompanied by brittle fracturing resulted in greater permeability (Smith 1984). There is a strong association of gold with arsenopyrite, but it also occurs as free grains associated with pyrrhotite, pyrite and chalcopyrite.

ECHO, PTARMIGAN AND CLEARWATER BAYS

No recent systematic mapping of this area has been carried out since that of Thomson (1936), but it is possible that most of the volcanic rocks are equivalent to the calc-alkaline felsic sequence and upper mafic sequence of Shoal Lake.

The dominant structural feature is the Crowduck Lake-Rush Bay fault zone in which the volcanic rocks are strongly to intensely foliated. Carbonate is abundant, but only in a few places is there evidence of much mobilization of silica. Where deformation has been greatest, thin quartz veinlets and lenses occur and fine pyrite is disseminated in the schists, but only minor amounts of gold have been found. The zone has been traced to the east at least as far as the south shore of the Northern Peninsula: Ayer (1984) has recently suggested its continuation right across the northern part of Lake of the Woods to Andrew Bay.

Around Clearwater Bay, intermediate to felsic pyroclastics predominate and are interlayered with basalt and overlain by fine clastic sediments. An east-striking synclinal axis lies along

the bay, and a well-developed foliation is within 30 degrees of east. A number of quartz-porphyry dikes also trend east, possibly marking zones of shearing. Quartz veins are mainly in areas of intermediate lapilli-tuff and tuff-breccia, especially in the vicinity of the former Kenricia Mine (Stop 4 on the field trip).

North of the pyroclastics, and Highway 17, foliated basalt outcrops over a width of 1000m. It is in contact with granitic rocks to the north, and may be equivalent to the tholeiitic sequence of Shoal Lake.

KENORA

A wedge-shaped area of volcanic and sedimentary rocks extends northeast from the main body of supracrustal rocks in the vicinity of Kenora (Figure 4). Intensely deformed gneisses lie to the northwest, and to the east are granitoid stocks which may be related to the Dryberry batholith. Tholeiitic basalts are overlain by fine-grained intermediate to felsic pyroclastics, which are in turn overlain by clastic sediments. A gabbroic sill lies near the top of the basaltic sequence.

The principal direction of faulting, the weak to strong foliation, and the trace of fold axes are all approximately parallel to wedge boundaries and converge to the northeast. At the northwest side of the wedge mafic rocks are highly deformed and metamorphosed but elsewhere primary features are largely preserved. An oval stock of porphyritic quartz monzonite, which lies on the trace of the Airport Anticline, is only weakly foliated.

All the known volcanic-hosted gold occurrences except one are in basalt and are associated with quartz veins or silicified shears. Fractured and mineralized felsite dikes which lie in sheared basalt have also been investigated. The mineralized zones trend northeast and are interpreted to be related to movement along near-vertical axial planar shears. Tourmaline and minor sulphides are associated with most quartz veins; chlorite, biotite, carbonate and sulphides are common in the sheared basalt.

Gold occurrences in the Island Lake quartz diorite are associated with shearing. Most of the mineralization is in or near quartz veins which occupy zones of dilatancy. Gold is associated with pyrite, especially along minor fractures in the quartz and the host diorite or quartz diorite. Diike-like bodies of ultra-mylonite lie near mineralized and silicified shears at three occurrences.

Stop 5 on the field trip illustrates the former type, while mineralization at Stop 6 is hosted by the Island Lake quartz diorite.

BIGSTONE BAY

Tholeiitic basalts with an estimated thickness of 8000m underlie Bigstone Bay and are broadly folded about the Hay Island

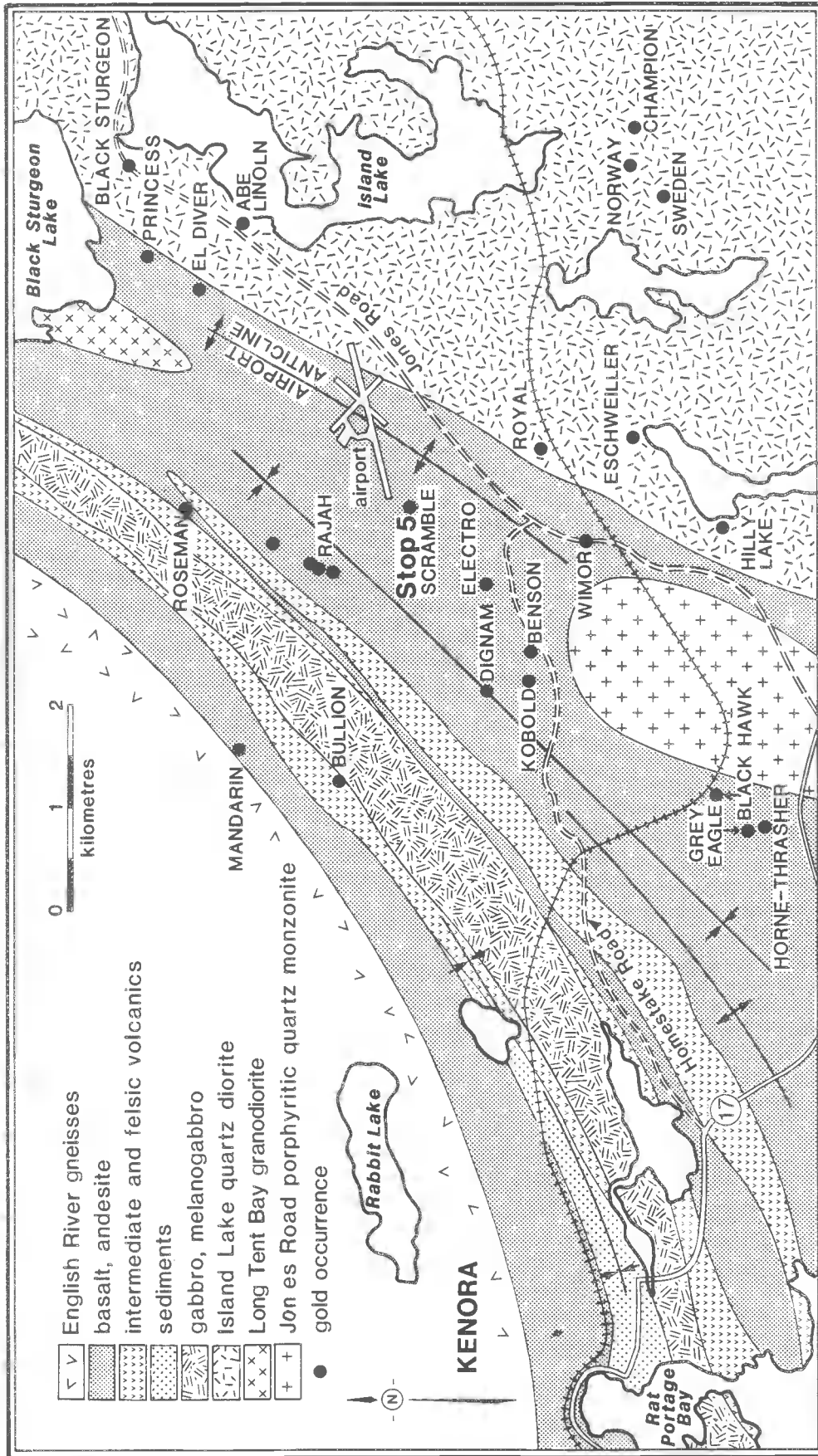


Figure 4: Geology of the Kenora area (modified from King, 1983).

Antiform (Figure 5). Mafic and ultramafic sills or flows, which lie near the top of the sequence, are exposed on the south limb of the fold and on the northwest limb near the fold nose. The mafic rocks are overlain by intermediate to felsic pyroclastics and flows which occupy the core of the Sultana syncline; the syncline and other parallel folds appear to have been refolded about the Hay Island Antiform.

Granodiorite of the Dryberry batholith occupies the core of the antiform. Basalts typically show little change in texture or grain size adjacent to the contact. The Quarry Island stock, which intruded the top of the basalt sequence, similarly shows no contact effects; it consists of medium-grained quartz diorite with a central core of porphyritic granite in the northeast part of Quarry Island and western part of Sultana Island.

On the northwest limb of the antiform, virtually all of the gold occurrences are associated with shear zones parallel to volcanic stratigraphy. Quartz veins with pyrite, many of which contain tourmaline, typically pinch and swell or occur as en-echelon lenses in the chlorite schists. Visible gold is erratically distributed in the quartz. This type will be viewed at Stop 7 on the field trip. On the south limb, the volcanics vary in strike, especially near the granodiorite contact. In the southeast part of Bigstone Bay and extending east to the Islet Lake area, several east-southeast trending shears dip south and show evidence of reverse slip; the shears are carbonatized and contain pyrite and narrow auriferous quartz veins.

Several old workings are in silicified and carbonatized shears which are parallel to foliation in the intermediate pyroclastics. At Middle Island a shaft was sunk on a strongly carbonatized zone in which gold is associated with galena.

The only important gold production in the Bigstone Bay area was at the Sultana Mine on the west side of Sultana Island, which operated from 1891 to 1906. Here gold-bearing veins hosted by sheared quartz diorite and porphyritic granite of the Quarry Island stock were reported to have been most productive at the contact between the two intrusive rocks.

WITCH BAY

Pillowed and massive tholeiitic basalts in the Witch Bay area were intruded by layered sills of peridotite to leucogabbro, and the sequence has been folded about east-trending axes (Figure 6). Intermediate and felsic volcanics overlie the mafic rocks. Granodiorite of the Dryberry batholith cross-cuts this regional fold structure.

Gold occurrences are in carbonatized shear zones in basalt. The gold is hosted by silicified and pyritiferous schist or by quartz veins containing up to 20 percent sulphides. At the Wendigo mine and Stella occurrence, the mineralized shears lie stratigraphically below the peridotitic base of a sill. The Witch Bay occurrence appears to be at a similar stratigraphic level. The shearing is believed to be related to competency contrast between the basalt and the sill. Two of these deposits will be seen on the field trip at Stops 8 and 9.

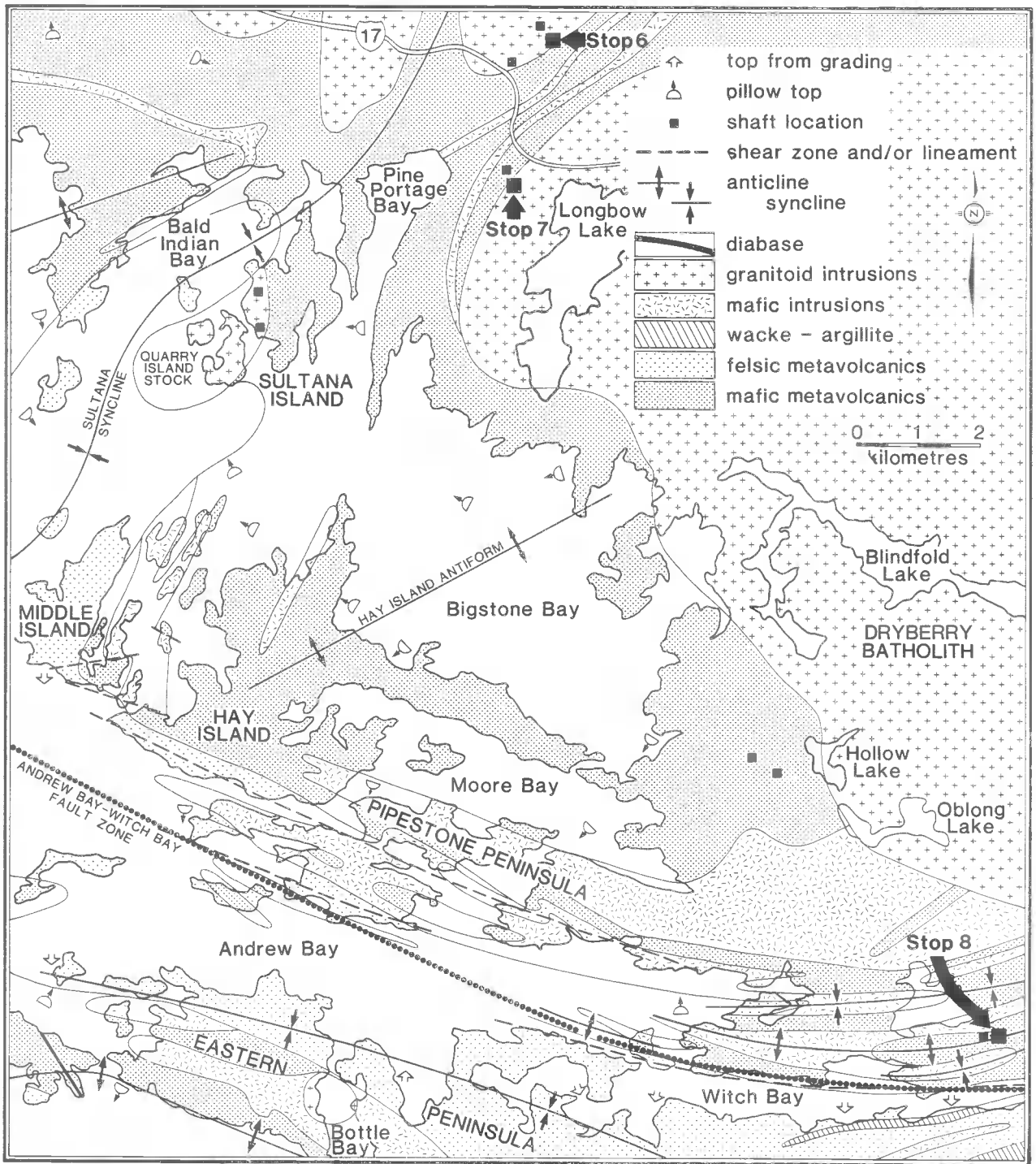


Figure 5: Geology of the Bigstone Bay area (modified from Ayer, 1984).

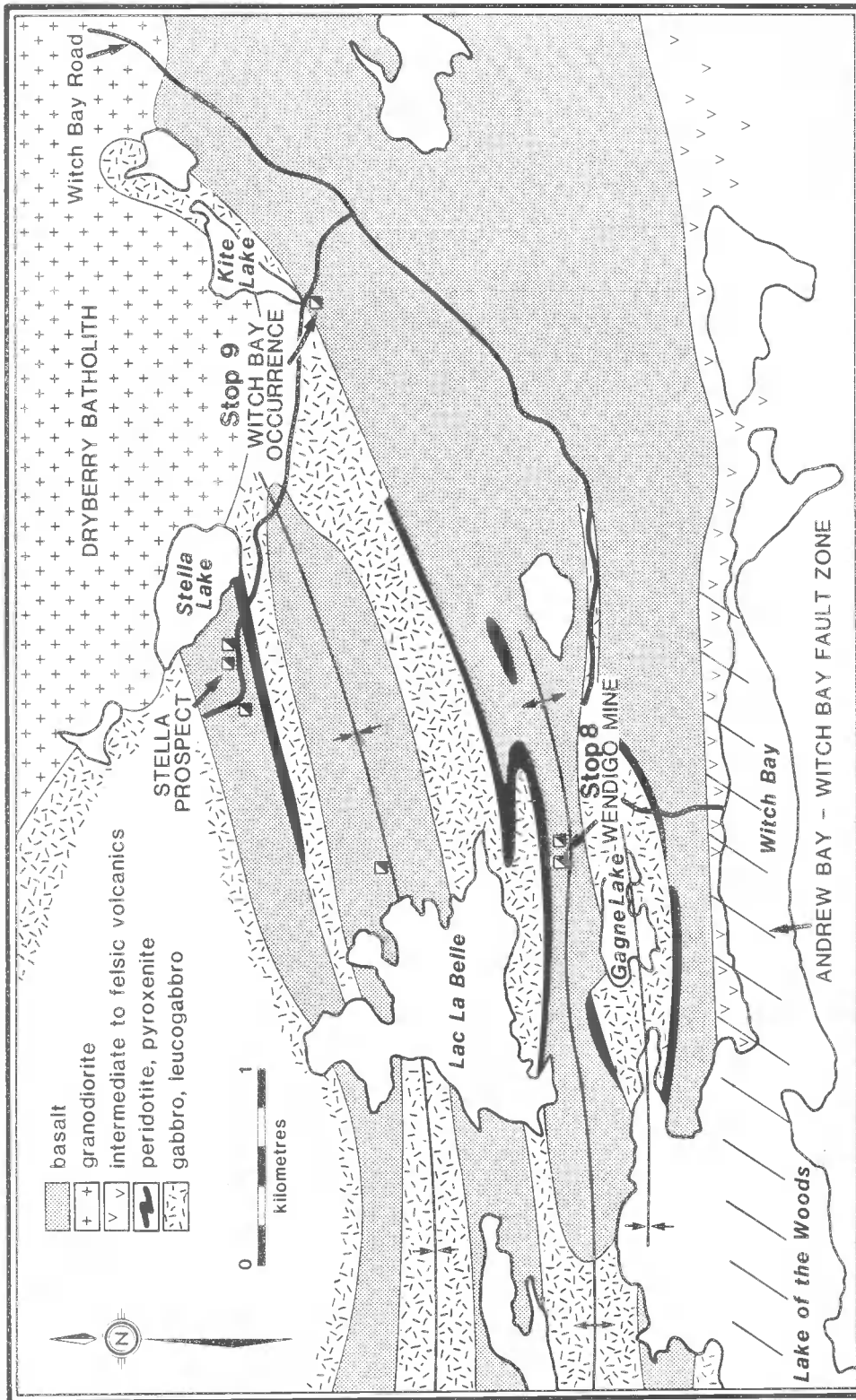


Figure 6: Geology of the Witch Bay area (modified from Davies and Smith, 1984).

The Wendigo Mine produced gold and silver from 1936 to 1943 and copper during part of that period. The average gold content of milled ore was 0.33 ounces per ton. The main quartz vein is in a carbonatized chloritic shear zone which strikes easterly and dips steeply north. The vein was up to 75 cm wide but averaged 30 cm, and contained up to 50 percent sulphides (pyrite, pyrrhotite and chalcopyrite) as stringers and small masses which are concentrated at edges of veins.

CONTROLS ON MINERALIZATION

Recent work by Davies and Smith (1984) has shown that in the general Lake of the Woods area:

1. The majority of gold occurrences are hosted by tholeiitic basalts that occur stratigraphically below felsic volcanic sequences.
2. The major past-producing mines (eg. Wendigo) and the recently active Duport property occur near the top of this tholeiitic sequence, where preliminary data indicate a higher iron magnesium ratio.
3. All of the occurrences are associated with faults or shear zones. Most gold production has come from shears which are nearly parallel to volcanic stratigraphy.
4. Hydrothermal solutions were rich in silica and carbon dioxide, and most were enriched in sulphur, potassium and boron.
5. Gold may have been carried as a sulphur complex in these solutions: its precipitation may have been enhanced by combining sulphur with iron. If such conditions prevailed, iron-rich basalts may have been the preferred site for gold deposition.

ACKNOWLEDGEMENTS

Thanks are due to: Bernard Guarnera of Boise Cascade Corporation, for permission to work on the Scramble property and for the release of geological maps, and to Mike Parr for stimulating discussion of the epigenetic versus syngenetic evidence at the Scramble; numerous students from the University of Manitoba who under the direction of Lorne Ayres and Bill Brisbin, have cleared outcrop over many years at the Conglomerate Zone unconformity; and to Mike Hailstone for his enthusiastic work with a Wajax pump during preparation for this field trip.

THE FIELD TRIP

INTRODUCTION

The field trip route utilizes existing roads and highways, and is designed to examine major aspects of Lake of the Woods gold occurrences without going onto the lake itself.

On the afternoon of Day 1, the trip commences near the Manitoba border, to view several occurrences within and marginal to the High Lake stock. It then proceeds east along Highway 17 (Trans Canada Highway) towards Kenora, with one stop along the way at the Kenricia Mine, one of the more recent past-producers in the area.

On the morning of Day 2, we will examine a number of occurrences within the lowermost mafic volcanics of the Kenora area. On the afternoon of Day 2, occurrences are viewed near Witch Bay, including the former-producing Wendigo Mine. These occurrences lie in the upper part of the lower mafic volcanics, near the contact with the overlying calc-alkaline, predominantly pyroclastic suite.

DAY 1

Proceed west from McLeod Park (Husky the Muskie) in Kenora along Highway 17 to the Shoal Lake Road turn off, a distance of 44 km. The Trans Canada Highway runs parallel to, and in general just to the south of, the boundary between the Wabigoon and English River Subprovinces.

Turn left (south) on the Shoal Lake road. Proceed 2.8 km to the High Lake road, a gravel road on the right. From this point proceed 3.2 km along the gravel road to the Evenlode molybdenite property of Eco Explorations Ltd. at the east end of High Lake.

STOP 1 EVENLODE MOLYBDENITE-GOLD MINE (Figure 7)

At the Evenlode property a number of molybdenite-bearing quartz veins occur within ductile shear zones in the porphyritic phase of the High Lake subvolcanic intrusion. The veins contain low to trace amounts of gold.

Four veins of molybdenite-bearing quartz have been identified on this property. The major surface and underground work has been done on the Main, or No. 1 vein, the most recent work being by Eco Molybdenite Mines Ltd. from 1980 to 1982. The headframe is still standing, but the mill was removed from the site in 1984.

"The main vein is associated with a strong shear zone striking about N.80°W, and dipping about 80°N in its central portion, but more shallow to the east and west. Obliteration of feldspar phenocrysts, development of sericite, and minor silicification occur adjacent to the shear zone in which quartz veins, lenses and

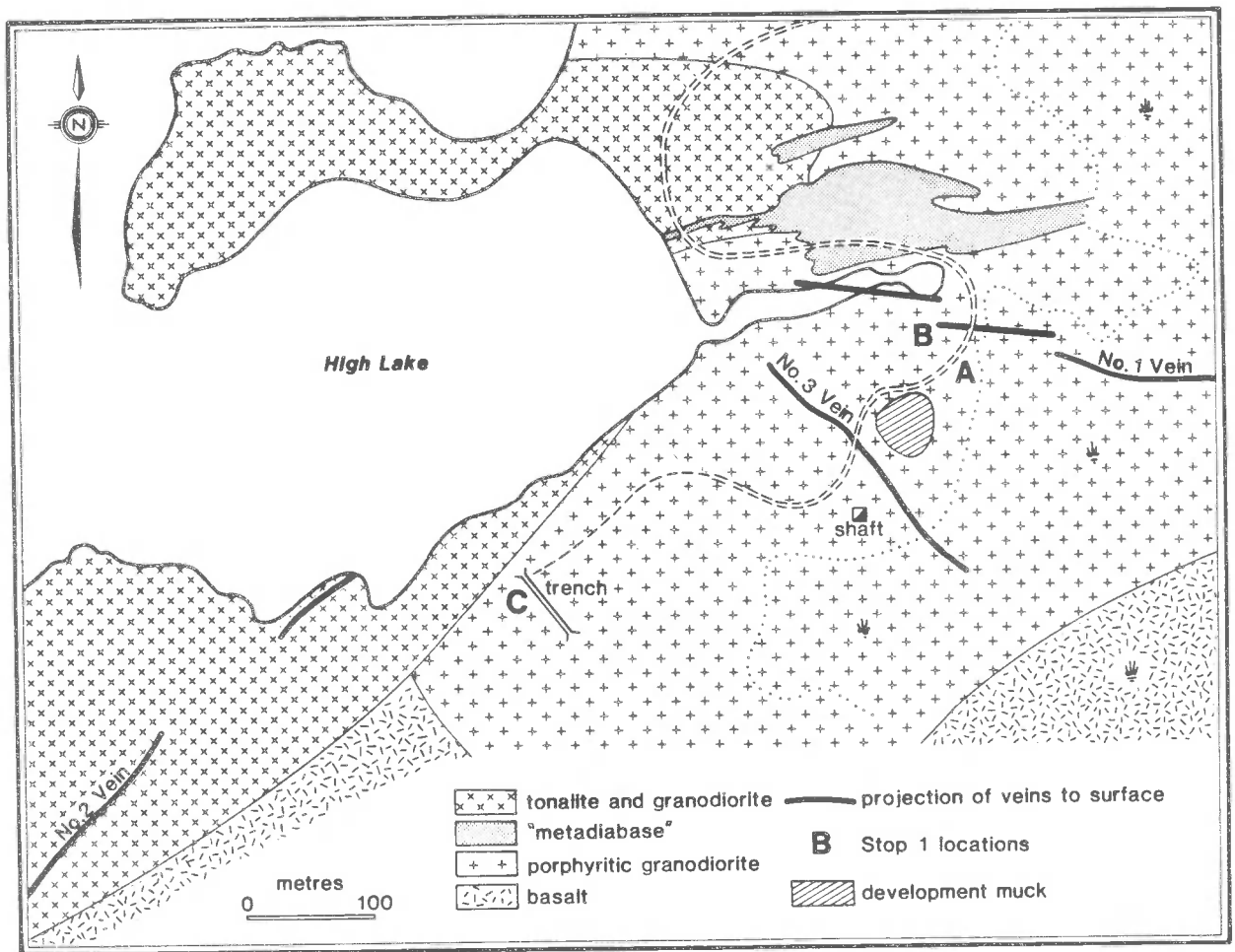


Figure 7: Geology of the Evenlode Molybdenite-Gold Mine (after company plans, Assessment Files Kenora).

stringers are found. The shear zone has been traced for about 2300 feet, the eastern 1600 feet being that portion containing mineralized quartz.....'Drill-indicated ore' over a length about 1200 feet...., across an average width of 4.7 feet, and to an average depth of 508 feet, has been calculated at about 126,000 tons grading 0.68 percent molybdenite" (Davies, 1965, p.51). Bulk sampling gave a higher grade (0.87 percent molybdenite) and 0.015 ounces of gold per ton (about 0.5 grams of gold per tonne).

The No. 2 vein system, to the southwest, trends northeast and consists of one dominant quartz vein and several subsidiary veins within a shear zone 30-80 feet wide. The dominant vein, which has been traced for 700 feet, "is reported to average 0.47 percent molybdenite over a 2.2 foot width, with gold averaging .02-.05 ounces per ton" (Davies, 1965, p.51). The vein is poorly exposed and will not be visited. The No. 3 vein strikes northwest and dips 45° to 80° northeast, with an estimated grade of 1-2 percent molybdenite. A shaft was sunk near this vein and muck from it covers the vein completely.

Locality A An outcrop of unaltered granodiorite, typical of the porphyry phase of the High Lake intrusion, occurs close to the road at the entrance to the property. The rock is characterized by rounded quartz and subhedral plagioclase phenocrysts in an aphanitic matrix.

Locality B The main vein is exposed in a trench through which a small stream drains. Molybdenite-bearing quartz occurs in a silicified shear zone in which the destruction of feldspar phenocrysts was accompanied by sericitization.

Locality C A fourth vein, between the No. 2 and No. 3 veins, has been uncovered in the north part of a 60 m trench. The 'vein' strikes about N10°W. dips steeply west and consists of a shear zone in which there are narrow, discontinuous quartz veins. Pyrrhotite is associated with the molybdenite. Assay data for the vein are not available.

From the Evenlode property return 2.2 km along the High Lake road. The road passes first through porphyritic granodiorite and then through basalt which is cut by porphyry. At the northernmost point in the road, a small amount of gold and silver was recovered from irregular zones of quartz exposed in pits just north of the road. The surface geology here demonstrates the complexity of the basalt-porphyry relationship. Stop near the gate; outcrop to the north is porphyry. From the parking point walk east about 120 metres to a flagged trail, and south on the trail to Stop 2.

STOP 2 THE BASE OF THE CROWDUCK LAKE GROUP: THE
UNCONFORMITY AND THE CONGLOMERATE ZONE GOLD
PROSPECT

At this stop a rather spectacular exposure of the unconformity at the base of the Crowduck Lake Group is in close proximity to a minor but structurally significant gold occurrence, the Conglomerate Showing.

Locality A Basalt, and porphyry dikes related to the High Lake intrusion, are present in the western part of this outcrop area and are overlain unconformably by sediments of the Crowduck Lake Group. The sketch map by Beakhouse (this volume, p.86) illustrates three sedimentary units that, over-all, show coarsening upward, from cherty siltstones and thinly bedded sandstones at the base, through pebbly sandstone into conglomerate. In addition, massive to thickly bedded sandstones occur in the north part of the outcrop area. Relationships between these units are complicated by syn-depositional erosion and by the original highly irregular surface upon which deposition occurred. Noteworthy features at the surface of the unconformity include; a regolith, most notably developed above a porphyry dike; possible evidence of syn-depositional faulting at the southern contact of the porphyry with basalts; local provenance of basaltic clasts in a lens of conglomerate.

The conglomerate is characteristically heterolithic; porphyry clasts are predominant and conspicuously larger than other types, which include mafic and felsic volcanics, chert and iron formation.

Soft-sediment deformation structures in the finer sedimentary rocks include slump folds, load casts, and flame structures.

Foliation is at high angles to bedding, and parallel to clast elongation and fold axial planes. Minor folds show geometry similar to regional-scale folding.

Quartz and quartz-tourmaline veinlets fill fractures in some porphyry clasts. The veinlets are oblique to foliation and commonly project into the conglomerate matrix, clearly indicating their late emplacement.

From this point we return to the road, walk west to the gate and south 20m through the bush to the open outcrop area.

Locality B CONGLOMERATE PROSPECT (Figure 8)

An embayment of the basal conglomerate at this location has an average width of about 15m and is exposed over a length of 95m. The conglomerate is enclosed to the south, west and north by foliated basalt and consists of subrounded clasts in a strongly foliated

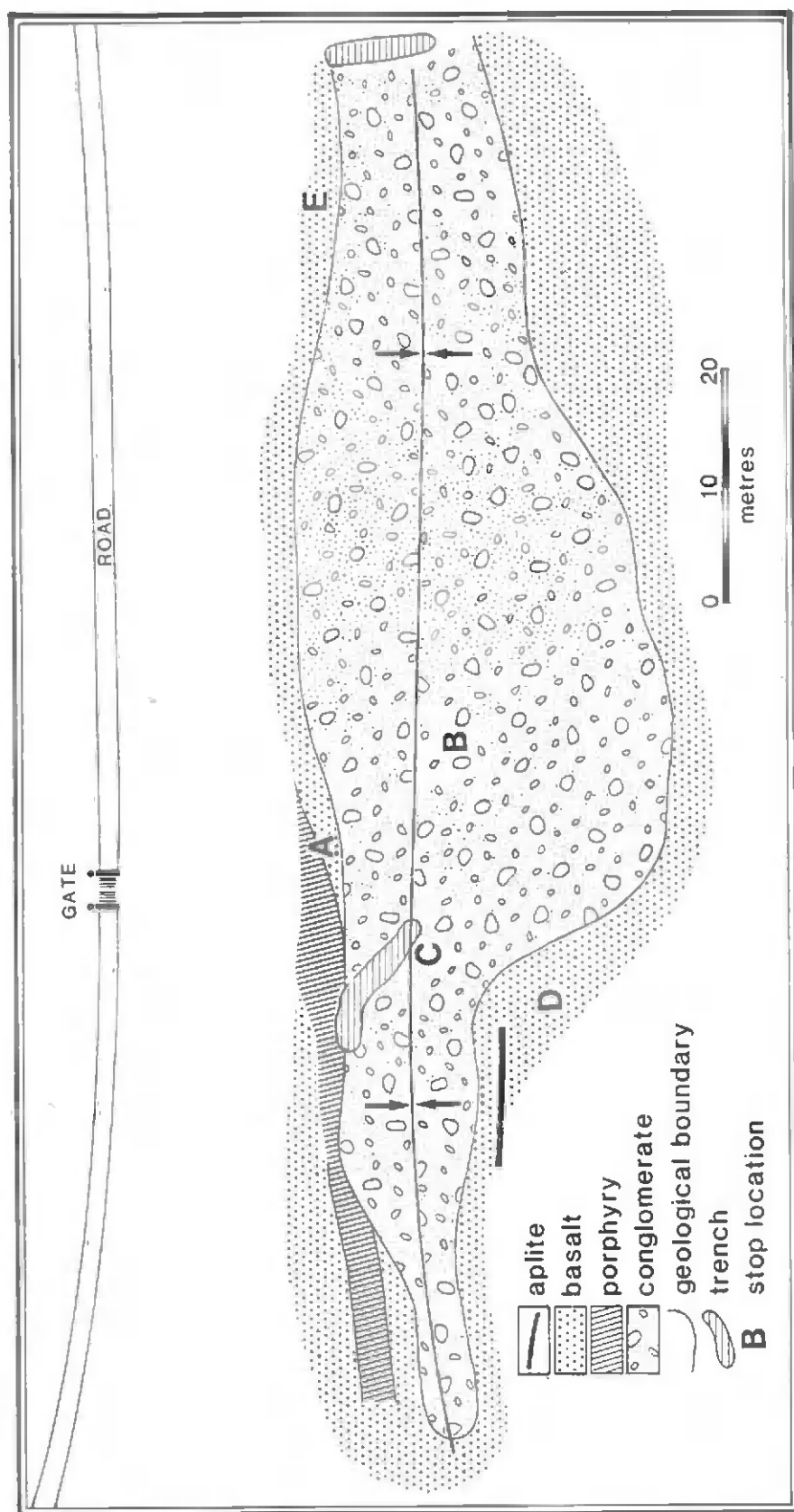


Figure 8: Geology of the Conglomerate Zone gold prospect (adapted from Davies, 1965).

dark wacke matrix. Clasts are mostly quartz porphyry and granodiorite, with lesser amounts of feldspar porphyry and mafic volcanics, all of which may have been locally derived. There is no distinct bedding, although clast size in general increases to the west. Clast elongation, which is slightly greater near contacts with basalt, is parallel to the east-trending foliation; average width: length: depth ratios are about 1:2.5:4.

The geometric and spatial relationships between the conglomerate and basalt, combined with the strong east-trending foliation, suggest that the embayment of sediments may have been modified by tight folding about a synclinal axis plunging 65° in an easterly direction.

Quartz is mainly present as lenses, thin veins and irregular masses which cut across foliation, especially in the vicinity of the pit. Quartz also occurs in tension fractures in a few granitic clasts and in a porphyry dike in schistose basalt at the north edge of the outcrop. Pyrite is present in pockets in the irregular masses of quartz and as disseminations in quartz veins and along slip and foliation planes, especially in the conglomerate matrix. Minor tourmaline is also associated with the quartz.

The pit was the source of a 76 ton bulk sample from which an average gold recovery of 0.14 oz. per ton was obtained in 1939. Extensive drilling and channel sampling was undertaken in 1944, but only 11 core samples of greater than 1.5m length had gold assays exceeding 0.08 oz per ton. The erratic nature of the mineralization is demonstrated by a 1.8m intersection which assayed 1.48 ounces of gold per ton (Davies, 1965, p.35).

The presence of gold in quartz which cuts across foliation, indicates the late- or post-tectonic timing of mineralization.

Return to the Shoal Lake road, turn left (north), and proceed about 150m. Walk east along a trail across outcrops of conglomerate to the Arsenic Zone.

STOP 3 ARSENIC ZONE GOLD PROSPECT (Figure 9)

The Arsenic Zone is so called not because arsenic is a major component but because, of nine zones explored by Electrum Lake Gold Mines Ltd. around 1960, it was the only one in which arsenopyrite was identified. At this occurrence, gold is associated with quartz that contains tourmaline, pyrite, arsenopyrite and minor pyrrhotite. The quartz is in irregular veins which occur in a complex of porphyry and felsic pyroclastics or flows, near the north side of a prominent hill of conglomerate. As at the previous stop, the occurrence is close to the unconformity at the base of the Crowduck

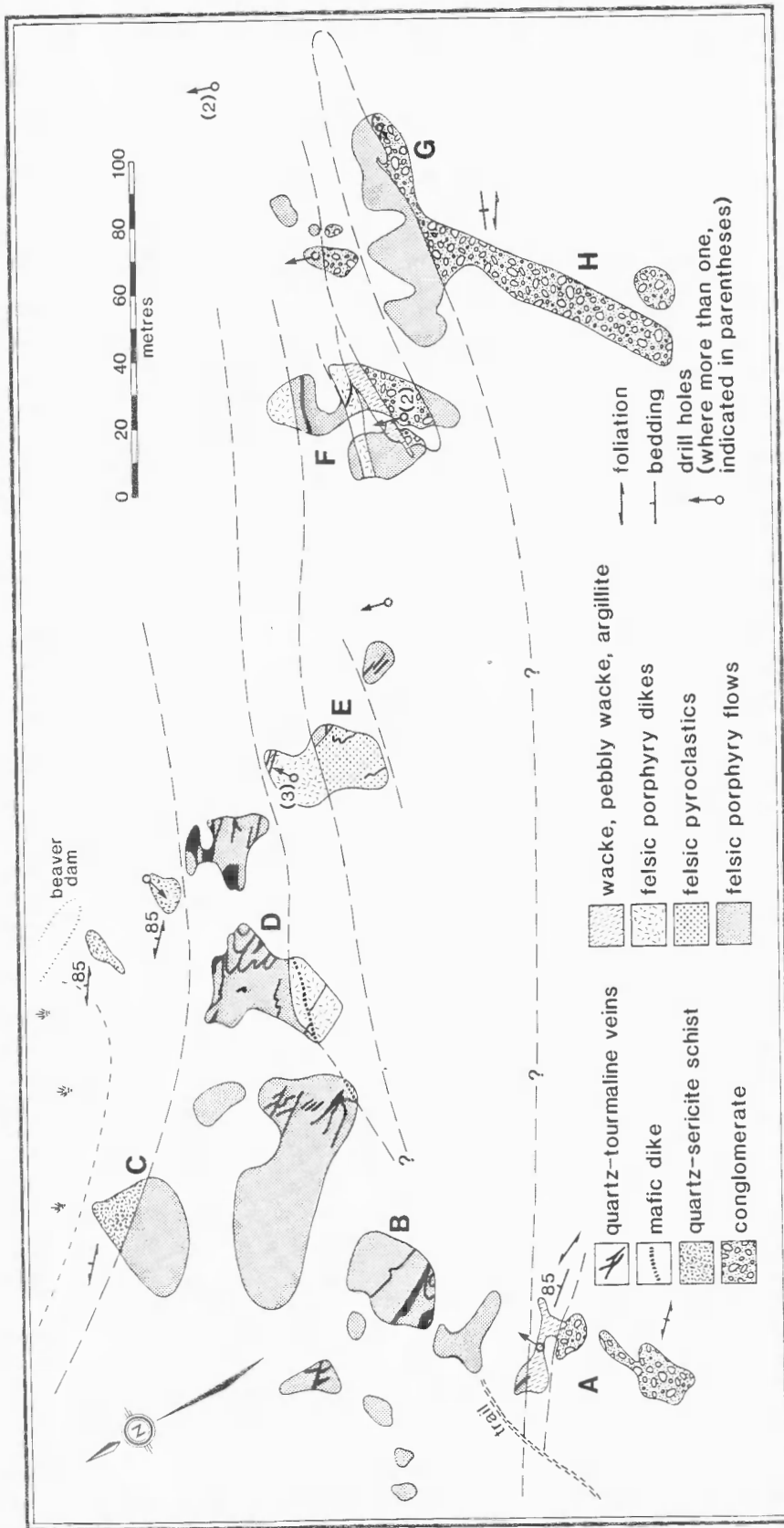


Figure 9: Geology of the Arsenic Zone gold prospect.

Lake Group.

The relationship between the conglomerate and the felsic volcanic and sub-volcanic rocks is difficult to determine. In a previous interpretation (Davies, 1965) it was suggested that the felsic rocks were all porphyry that occurred as an east-trending dike, and younger than greywackes that were identified north and south of it. The conglomerate in turn was considered to be younger than the porphyry dike. It was also noted (Davies, 1965) that this interpretation was not satisfactory, since the greywacke appeared to grade into the conglomerate.

Our recent mapping has shown that the "porphyry dike" consists of at least two phases of porphyry, and a pyroclastic unit. The contact between the main porphyry and the conglomerate appears to be an unconformity, along which shearing has occurred (Localities A and G). This main phase of the porphyry is similar to porphyry identified elsewhere as part of the High Lake stock. A small outcrop of pyroclastics (Locality E) is probably associated with this main porphyry phase. Elsewhere in the High Lake area similar transitions from felsic porphyry into pyroclastics of the same composition support the interpretation that this porphyry is a sub-volcanic intrusion. Dikes of porphyry clearly cut the main-phase porphyry (Locality F), but the timing of their emplacement relative to deposition of the conglomerate is not clear.

The highly laminated, fine grained grey rocks on the north side of the porphyry (Locality C), formerly identified as greywacke (Davies, 1965), are here considered to be a mylonite zone. The mylonite may underlie the swampy area to the north. This mylonite zone is probably associated with the dominant easterly trending faults.

The gold-bearing quartz veins occupy fractures that clearly post-date: the main phase porphyry within which they are dominantly developed (Localities B, D and F); the sedimentary rocks, within which some pebbles and boulders contain late fractures that have been filled by quartz veins (Localities A and H); and the late porphyry dikes (Localities D, E and F). Their relative age with respect to the mylonite is unknown.

Geometry of the quartz-tourmaline veins is complex, and has not been analyzed in detail. However, there appear to be two dominant trends within the main porphyry, one northerly, the other southeasterly. These dominant trends are especially evident at Locality D. Within the conglomerate, quartz-filled fractures in pebbles and boulders show consistent orientation within individual outcrops, but this varies from outcrop to outcrop (see Localities A and H).

In summary, gold at the Arsenic Zone is contained within quartz-tourmaline veins that were emplaced into

fractures developed by brittle failure in felsic porphyry and associated rocks, marginal to an east-southeast-trending mylonite zone.

Locality A Crowduck Lake polymictic conglomerate, containing clasts of predominant porphyry, minor dacite, and rare basalt and chert, is in contact with porphyry. A narrow unit of inter-bedded sandstone and conglomerate with 1-2 cm pebbles lies immediately adjacent to the porphyry. The contact with the porphyry is sharp and sheared. A quartz vein occupying a fracture in the porphyry extends a few centimetres into the sediments. Larger conglomerate clasts at this locality show quartz-filled tension fractures that extend into the matrix, indicating their late development.

Locality B Quartz-tourmaline veins with associated pyrite, pyrrhotite, and minor arsenopyrite occupy a shear zone trending 140° in porphyry. Other veins occur in minor fractures.

Locality C Mylonitized porphyry underlies the north slope of the outcrop ridge. Strike is about 130° , dipping steeply northeast.

Locality D Abundant quartz-tourmaline veins, showing two prominent orientations, one approximately north, the other southeast, occur within two phases of porphyry. A very narrow mafic unit, possibly a dike, parallels and in part follows the contact. Grab samples obtained from the southeast end of these outcrops contain up to 10 oz per ton Au.

Locality E Contact trending 105° between a porphyry dike and pyroclastic phase of the early porphyry. A number of clasts in the pyroclastic unit are elongated (138°) parallel to foliation in the conglomerates and the mylonite zone. Note quartz veins both in the porphyry and the pyroclastics.

Locality F Two late porphyry dikes intrude the main-phase porphyry. The highly irregular nature of the unconformity is seen here, with a deep embayment of both the conglomerate and a basal wacke to argillite unit into a valley in the porphyry.

Locality G The unconformity between the main-phase porphyry and Crowduck Lake conglomerates is sheared, but clasts of porphyry derived in place are identifiable in the basal part of the conglomerate. Note the sigmoidal, quartz-filled tension fractures in the contact zone.

Locality H Weak bedding in the Crowduck Lake conglomerate is seen at a very shallow angle to

foliation. Narrow quartz veins cut conglomerate in two dominant trends, one at 085°, another at 130°. Unlike Locality A, where clasts are large, and quartz veins are dominantly confined to the clasts, the clast size here is smaller and veins cut clasts and matrix.

From Stop 3 proceed north on Shoal Lake road to Highway 17. At the highway, turn right (east) and continue for about 30 km to the Kendall Inlet Road. For the first 11 km the highway is approximately parallel to regional foliation and passes through volcanic rocks (mainly felsic pyroclastics). The highway crosses granitic rocks for the next 10 km and for the remainder it is in well-foliated basalts and intermediate to felsic tuff-breccia. Turn right (south) on the Kendall Inlet Road and proceed south for about 2 km to the vicinity of the old Kenricia Mine.

STOP 4. KENRICIA GOLD MINE (Figure 10)

Gold-bearing quartz veins at this stop were discovered in the latter part of the 19th century. Three shafts were sunk and at that time the property was known as the Three Ladies Mine. In the mid 1930's substantial tonnages of gold ore were outlined by drilling. Development proceeded on three levels and a mill was constructed. During a one-year period, 2533 ounces of gold and 521 ounces of silver were recovered, but both the grade of the ore and the recovery were below expectations. The mine closed in 1940.

The area is mainly underlain by intermediate to felsic lapilli-tuff and tuff-breccia, having a combined thickness of 200 to 3000 m, and these are overlain by at least 800 m of fine pyroclastics and related sediments. An east-trending syncline extends through Clearwater Bay and the entire section from the boundary of the Wabigoon Subprovince to Clearwater Bay, which includes about 1000 m of foliated basalt at the base, is believed to be south facing.

The principal lithologies on the peninsula where the veins are located are tuff, lapilli-tuff, and tuff-breccia, containing dacitic clasts within a dark matrix. Foliation is moderate to strong, trending east and dipping near vertical. Clasts are aligned within the foliation plane.

Seven quartz veins have been discovered, although most of the work was concentrated on the No. 1 and No. 3 veins. All of the veins contain carbonate and most contain black tourmaline. Mineralization consists of minor pyrite and chalcopyrite with traces of galena and native gold. Thomson (1937) has described three generations of quartz in the No. 3 vein; quartz-carbonate veins are cut by sugary, faintly bluish quartz in which tourmaline and sulphides occur along thin

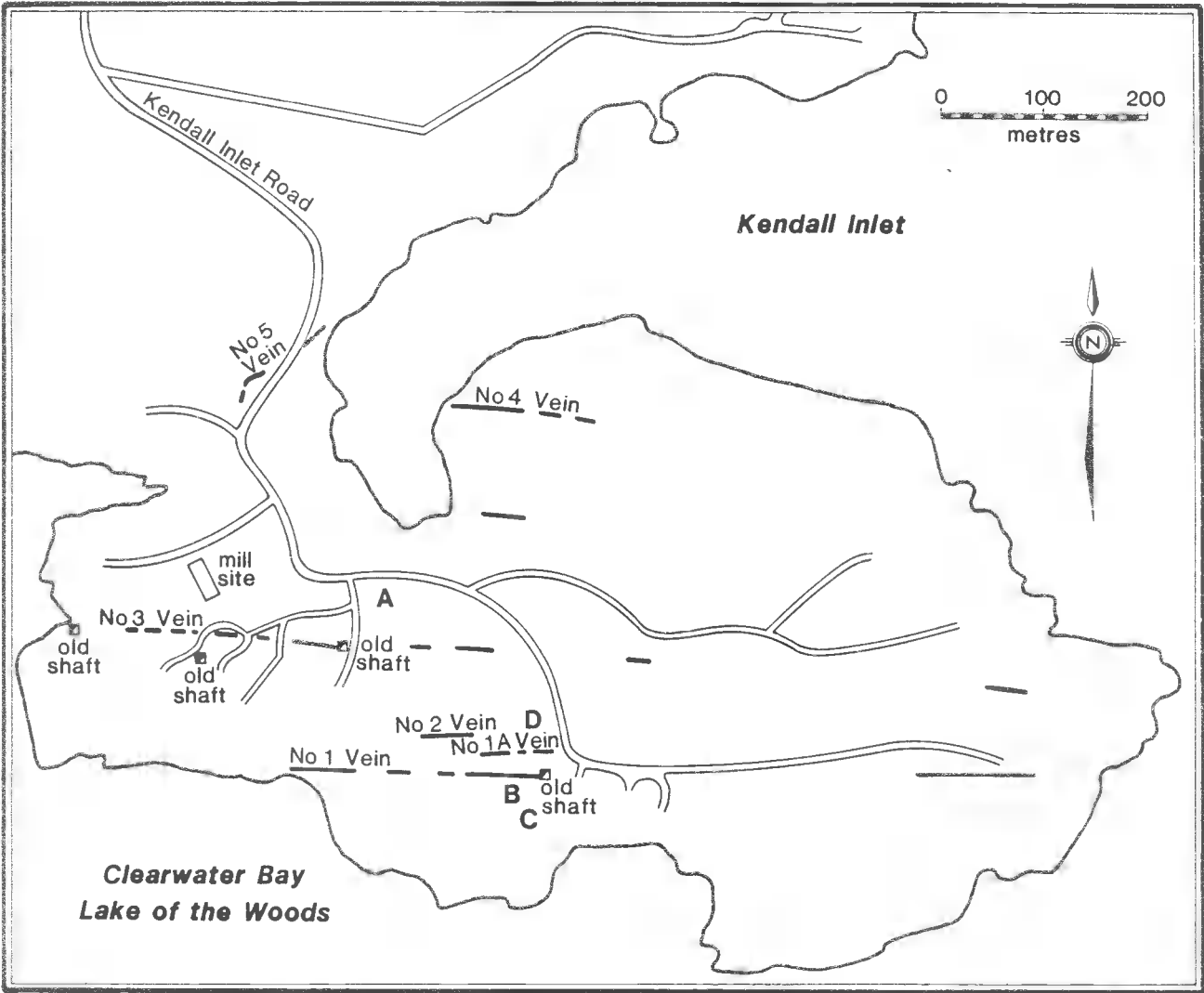


Figure 10: Location of gold-bearing veins at the Kenricia Gold Mine.

fractures, and these are cut by glassy white quartz. Gold is largely restricted to the sugary quartz.

Quartz veins in the Kenricia Mine area occur within zones where there has been a strong component of shear. These zones are parallel, or slightly oblique to regional foliation and contain concentrations of quartz, carbonate, tourmaline and biotite. Gold is restricted to quartz veins and to immediately adjacent altered wall rocks.

Locality A. Here the lapilli-tuff is moderately deformed, with clasts elongate parallel to foliation.

Locality B. The No. 1 vein has been traced for 475 m and has a maximum width of 1.8 m. It is best exposed at the east end of the trenches where it is in contact with highly sheared wall rock. Fragments of wall rock are found throughout the vein but are concentrated near the margins. Quartz varies from sugary and white to coarse and clear; minor sulphides, tourmaline and biotite are present in both types. The host schists contain up to 3% fine disseminated pyrite, are strongly silicified and are cut by quartz-carbonate veinlets. Coarse biotite is common at margins of the veinlets and also as flakes which cut across foliation and clast boundaries in the host rocks. Analyses of sugary quartz, wall-rock schist and clear quartz showed the gold content to be 310 ppb, 260 ppb and 505 ppb respectively. A second sample of clear quartz contained 855 ppb gold.

Locality C. Some 20 m south of No. 1 vein is an outcrop of strongly foliated dacitic lapilli-tuff. The rock is essentially a quartz-sericite-biotite-carbonate schist and contains rare broken remnants of sodic oligoclase. An analyzed sample contained 3 ppb gold.

Locality D. At this locality several veins, collectively known as No. 1a vein, comprise less than 30% of a zone of sheared dacitic lapilli-tuff. Bluish-white sugary quartz lies within the east-trending foliation plane as well as northeast-trending fractures, and encloses numerous fragments of wall rock. Veinlets of tourmaline cross the quartz. The host schists contain carbonate and 3% euhedral pyrite. A sample of vein material contained 3590 ppb gold, and the pyritiferous schist contained 540 ppb gold.

Return to Kenora. En route we will pass exposures of foliated volcanics and volcanogenic sediments.

DAY 2

Proceed 5.3 km east along Highway 17 from the Tourist Information Centre, to the junction of the Jones Road. Felsic pyroclastic

rocks are interlayered with basalts near the town, but much of the route passes through massive, pillowed basalts. At the Highway 17 - Jones Road intersection, quartz monzonite with large euhedral microcline megacrysts lies at the southern edge of an oval stock. Turn left (northeast) on the Jones Road. This gravel road crosses the eastern edge of the quartz monzonite stock up to the railroad crossing and then passes through massive to foliated basalt. Proceed 1.6 km beyond the railroad crossing, turn left (north) on dirt tracks through open clearings and then west through light bush to the Scramble Gold Mine (total distance of 1.1 km from the Jones Road). The dirt track crosses the axial trace of the northeast-trending Airport Anticline.

STOP 5 SCRAMBLE GOLD MINE (Figure 11)

Gold was discovered on this property in the 1890's and several interests "scrambled" to obtain a working option on it. The Scramble Gold Mining Company was incorporated in 1897. Two shafts were sunk and some underground development took place, but there was no gold production. The property lay idle until 1984 when Boise Cascade Corporation, current owners of the property, began investigating its gold potential.

Bedrock exposed in a recently stripped area consists of northwest-facing pillowed basalt and medium-grained basalts that are either subvolcanic sills or coarser-grained flows. Felsite dikes were intruded into the basalts, especially along northeast-striking zones of shearing.

Within the stripped area, sampling by Boise personnel has indicated that gold is more abundant in a zone subparallel to, and immediately southeast of the felsite dikes. Our sampling suggests that gold values are highest (0.37 oz. Au per ton over 1.5 m) at locality G (Figure 11) where the zone transects a felsite dike.

The dikes and the gold-bearing zone are in a zone of strong shearing which is about 20m wide, sharply bounded to the northwest (Locality D) and southeast by relatively undeformed pillowed basalts (Localities C and K). Remnants of massive basalt in the sheared zone (Locality J) are not obviously pillowed. Major deformation within the zone is believed to post-date emplacement of the dikes because: the dikes themselves are sheared, with foliation in the basalt cutting the dikes, at a low angle; where dikes pass from undeformed rock into the shear zone they have suffered drag (Locality D); fold styles in basaltic rocks of the shear zone are mimicked in the narrow, attenuated portions of felsic dikes.

Quartz and quartz-tourmaline veins occur in both the dikes and the basalt host. They commonly show evidence of folding and boudinage (Localities I and J), but must have been emplaced late because they fill shears in the felsite.

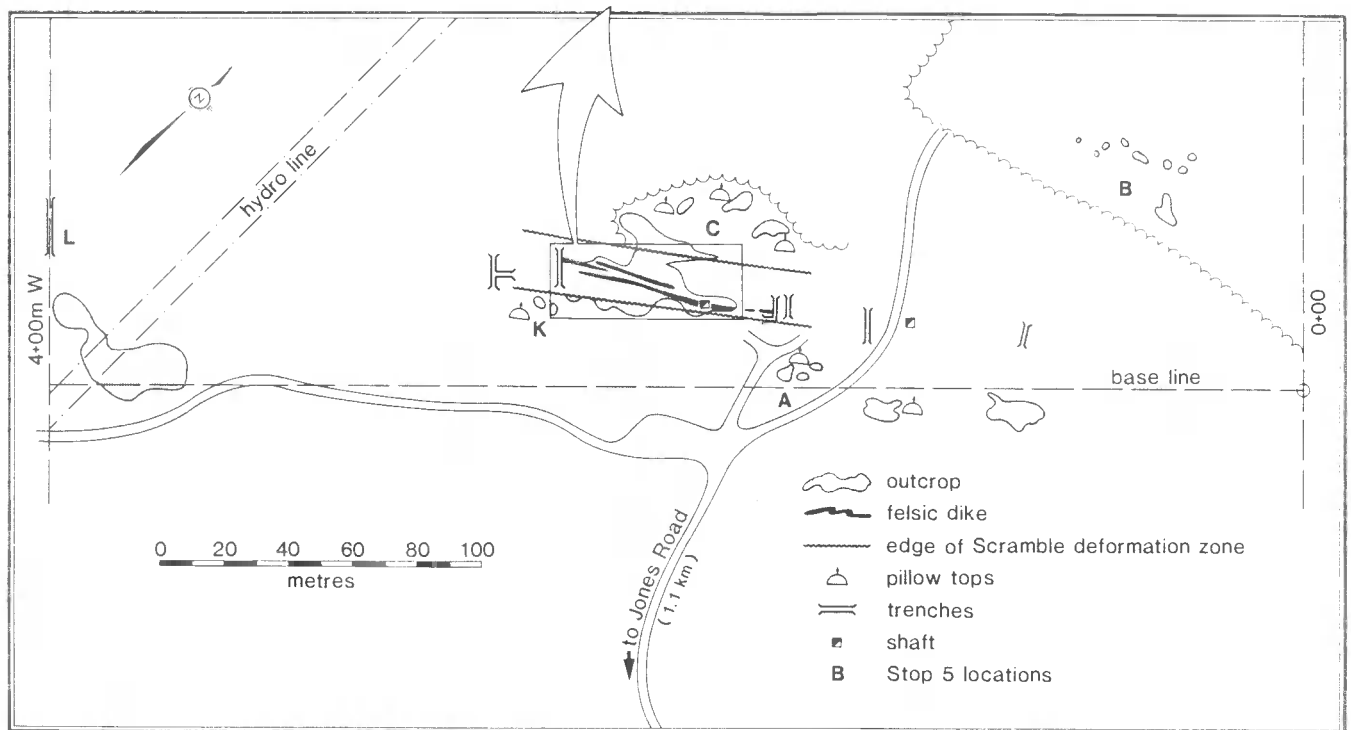
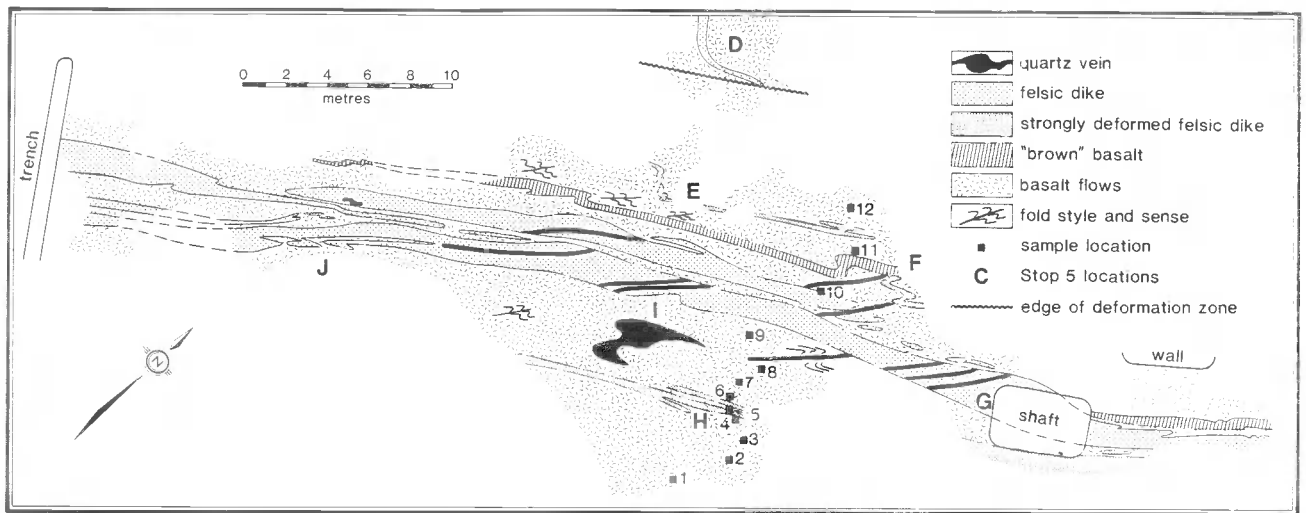


Figure 11: Geology of the Scramble Gold area. The area enclosed within the rectangle in the lower map is shown enlarged above.

Abundant minor folds are present in the shear zone. These are predominantly Z folds (Localities E, F and G) though S folds also occur (Localities F and J). Shear surfaces related to the development of Z folds strike at a variety of angles (approximately N45°E) but are invariably oblique to shear-zone boundaries (about N52°E).

Samples collected across the shear zone (1 to 13, Figure 11) show distinct petrographic and geochemical patterns. Relatively unaltered basalts (1 and 13) contain abundant hornblende (70-80%), lesser feldspar (10-25%), minor retrograde biotite and minor ilmenite and carbonate. Toward the centre of the deformation zone biotite increases at the expense of hornblende, and epidote and sphene are present in minor amounts. Samples from the central part are magnetite-bearing.

The geochemical pattern is shown in figure 12. While the pattern is complex, there is, coincident with the highest gold contents, an increase in Fe₂O₃, MgO and K₂O, and a decrease in CaO, Na₂O and Al₂O₃. The central part of the deformation zone is, in general, marked by an increase in K₂O and in the oxidation state of iron.

Locality A Pillowed lavas southeast of the Scramble deformation zone are observed on either side of the trail leading toward the airport runway. Although not too distinct, tops are interpreted to be toward the northwest.

Locality B Pillowed basalts are in contact with coarser-grained massive mafic rocks with 2-3 mm hornblende porphyroblasts, and rare garnets. The coarse variety may be either coarse flows or subvolcanic sills. Pillows are elongate, but appear to face north. Several fine-grained felsic dikes, similar to those to be seen at the Scramble shaft, cut both rock types. Margins of the dikes are feathered, with small shears displaying a dominantly dextral horizontal component.

Locality C Pillowed lavas adjacent to the stripped area around the Scramble shaft are exposed in a number of small outcrops. Features to observe are: distinct shapes of larger pillows indicating tops to the northwest; smaller, amoeboid pillows, with no clear top indicators; mafic dikes feeding the pillowed sequence; minor offset of pillow rims by shearing and faulting.

Locality D A felsic dike is at high angle to foliation. A sharp flexure in the dike marks the northwest boundary of the Scramble deformation zone. Note also quartz-tourmaline veins cutting the dike at a high angle (approx. N40°E) and occupying shear surfaces that extend into the host basalt.

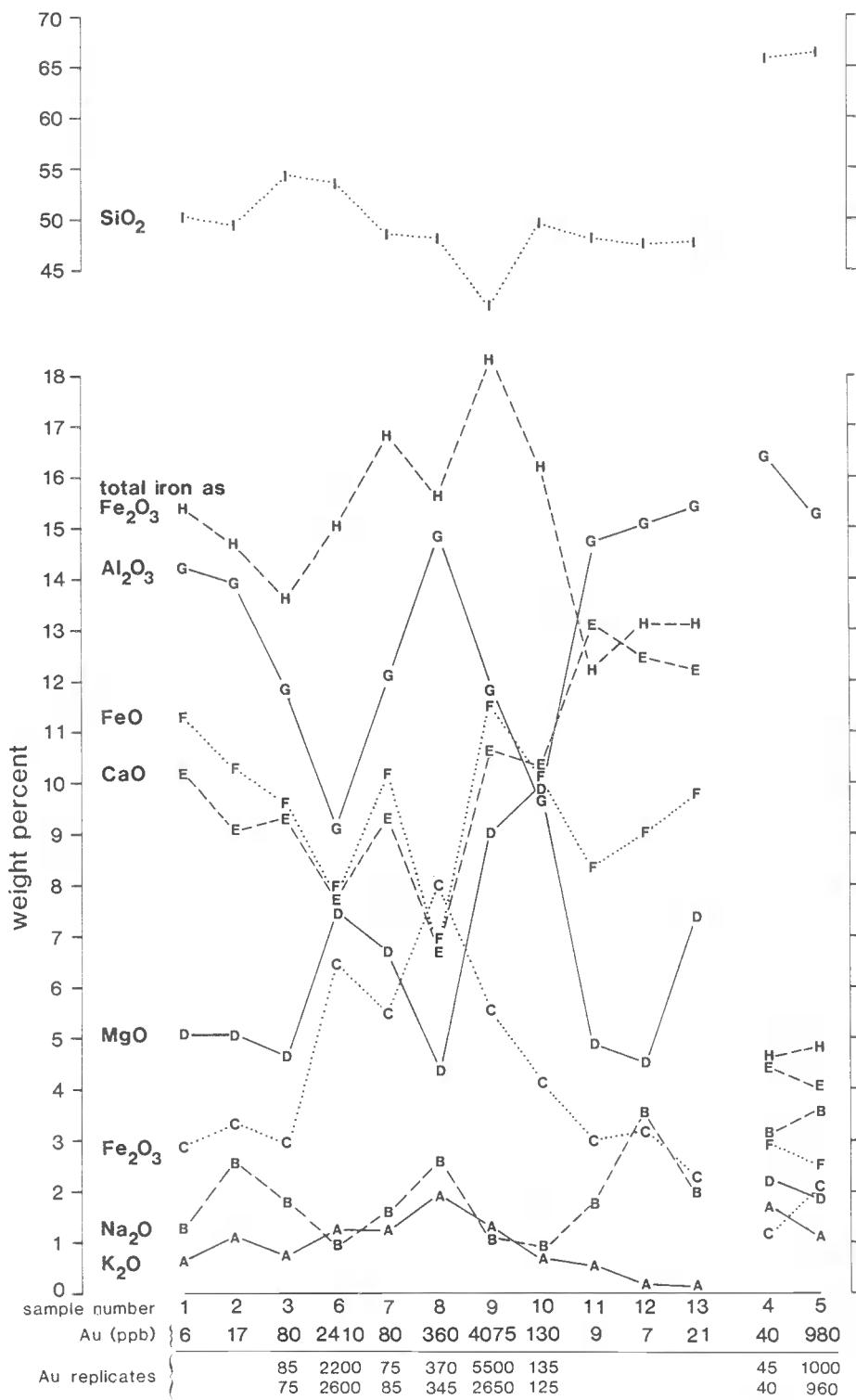


Figure 12: Geochemical traverse across the Scramble deformation zone. The location of samples is illustrated in Figure 11. Samples 1 to 3 and 6 to 13 are mafic metavolcanic rocks; 4 and 5 are strongly deformed felsic dikes.

Locality E Pronounced Z-folding of a tectonic fabric in basaltic lavas within the deformation zone. A number of felsic units are present and are interpreted to be early felsic dikes, rather than tuffs or flows. Fold axes plunge steeply. A diffuse zone of brown-coloured basalt can be traced toward Locality F.

Locality F S-folding is indicated in the "brown zone", which terminates northeastward against a north-northeast trending shear surface. Intense Z-folding in felsic and mafic rocks is evident. Mafic rocks are interpreted to be pillowed basalt flows. Felsic units are dikes.

Locality G A wide felsic dike can be seen to dip steeply northwestward in the shaft. Preliminary grab samples from the property indicated highest gold values to be within rocks at the shaft, both in the dike and the mafic volcanics. Immediately southwest of the shaft, a number of narrow quartz-tourmaline veins occupy shear surfaces in the dike, and strike at about N40-45°E. This trend is also seen in a quartz vein along strike, within the basalts, that occupies a shear zone exhibiting right-lateral movement (Z-folding).

Locality H Two felsic units interpreted to be dikes mark the south-east margin of the Scramble deformation zone. One of these dikes (Sample 5, Figure 11) contains 1000 ppb gold; well above background values in the basalt outside the deformation zone.

Locality I A large isolated quartz lens occurs here. The structural position of this lens is unclear. It could be the dislocated equivalent of a wide quartz-tourmaline vein 20 metres northeast of the Scramble shaft. In the main felsic dike, numerous narrow quartz-tourmaline veins, striking about N40°E, cross the dike at a low angle. Toward Locality J, numerous examples of these can be seen.

Locality J Examples of left-lateral shear, with associated S-folds, are seen within the main felsic dike. Right-lateral shear is evident within splays off the dike into basaltic country rock but the relationship to the sinistral shears is unknown. Deformed amoeboid pillows, similar to those seen at Locality C, occur within the Scramble deformation zone, on the northwest side of the main felsic dike. From this point, looking toward the Scramble shaft, a pronounced open-Z flexure in the felsic dikes is evident.

Locality K Pillow lavas southeast of the Scramble deformation zone suggesting tops toward the northwest.

Locality L On line 4+00W of Boise's grid, a trench

marks the location of an IP anomaly that the company has traced from the Scramble shaft area. The geophysical anomaly has a strike similar to that of the Scramble deformation zone.

Return, via Jones Road, to Highway 17, and turn left(east). At the crest of the first hill which the highway crosses is an exposure of pillowed basalt with white carbonate filling inter-pillow gaps. Reaction between the carbonate and the basalt has generated andradite garnet and epidote. The basalt is cut by granitic dikes. Beyond the crest the outcrop is diorite and quartz diorite of the Island Lake intrusion which noses out a short distance to the south. At the Garbage Dump turnoff, 1.6 km beyond the south end of Hilly Lake, turn left (northeast) and bear left for 200 m. To the left (northwest), stripping of overburden has been carried out in the vicinity of the Silverman occurrence.

STOP 6A SILVERMAN GOLD PROSPECT

The contact between massive quartz diorite of the Island Lake intrusion and massive basalt is very near the parking area. The contact is irregular in detail, but has a regional trend of N65⁰E over a distance of at least 1.5 km. The southeast contact of the intrusion was probably fault controlled and post-emplacement movement parallel to the fault has resulted in narrow zones of mylonite in the quartz diorite. Inclusions of basalt, some of which have been partly digested, are locally abundant.

At the Silverman prospect a number of shears trending about N35⁰E and having near vertical dips contain quartz, tourmaline and sulphides. The main workings consist of a 13.4 m shaft and two open cuts from which 184 tons of ore were shipped for milling in 1941. The average gold content of the ore was 0.575 oz per ton. The main vein was up to 50 cm wide but was discontinuous; it was emplaced within a strong shear zone, and where the main vein pinches there are several parallel thin quartz veins within the shear.

The quartz is sugary and black tourmaline is abundant locally. Pyrrhotite, pyrite and traces of chalcopyrite are present in amounts up to 2 percent.

Samples of fresh quartz diorite and highly sheared quartz diorite adjacent to the mined vein were analyzed. Relative to the fresh rock the sheared rock showed a net loss of SiO₂, Al₂O₃, Na₂O, As, Sb and Pb, and a gain of K₂O, CO₂ and H₂O. The gold content of the fresh rock was less than 2 ppb and of the sheared rock 8 ppb.

. Clearing and blasting has been carried out in the area between the Silverman vein and the road. Here, quartz-tourmaline veins occur in a swarm, occupying shear zones with a general N50⁰E trend. Small drags and

Z-folds indicate dextral shear. Their style and orientation is similar to those at the Scramble. Deformation continued after emplacement of the veins, since they are themselves drag-folded. In places the veins occupy strong shear zones a few tens of centimetres wide, or the shears themselves are apparently unaltered, while the immediate host diorite is enriched in tourmaline and silica. Granitoid veins, which occur in both the diorite and basalt inclusions are cut by felsite dikes that have been strongly sheared parallel to their length.

Return to the parking area and continue northeast for 0.8 km to the Treasure prospect. The trail is almost on the quartz diorite-basalt contact for about 100 m northeast of Stop 6, but there is little bedrock visible until the Treasure prospect is reached.

STOP 6B TREASURE GOLD PROSPECT (Figure 13)

The mine was developed between 1891 and 1900, when two shafts were sunk on the main zone and a third on a separate zone 200 m to the southeast. Additional work was done about 1955 when a small mill was set up. Recorded production is about 35 ounces of gold.

Quartz diorite underlies the area, the main vein being about 400 m northwest of the contact with basalt. The vein lies at the hanging wall of a shear zone which strikes from N35°E to N50°E and dips about 75° northwest. The vein's average width is about 15 cm, but it widens to 60 cm and in places pinches out completely. Tourmaline is common and pyrite is a minor constituent. Gold occurs as fine grains and is also smeared on slip surfaces.

It has been reported (Joop Langelaar, personal communication, 1983) that, in the main shaft, a series of quartz veins is present at the footwall of the shear zone. These veins are at an angle to the hanging-wall vein but are 'folded' in a manner indicating normal movement on the shear plane, and so have little lateral continuity. The footwall veins were either emplaced earlier than the hanging-wall vein, or both were emplaced simultaneously within a fault zone and its associated pinnate fractures.

The main vein appears to curve westward at its south end, and possibly to terminate against a mylonite which strikes N65°E and dips 75° to the southeast. The mylonite has a width of from 10 cm to 3 m. Chemical analyses of the mylonite indicate that, relative to the enclosing quartz diorite, the mylonite contains more silica (69.5% vs 62.2%) and soda (4.05% vs 3.35%) and carbonate (1.28% vs 0.13%), but contains less of other metal oxides (notably iron and magnesium) and trace metals. The gold content of both rocks is less than 2

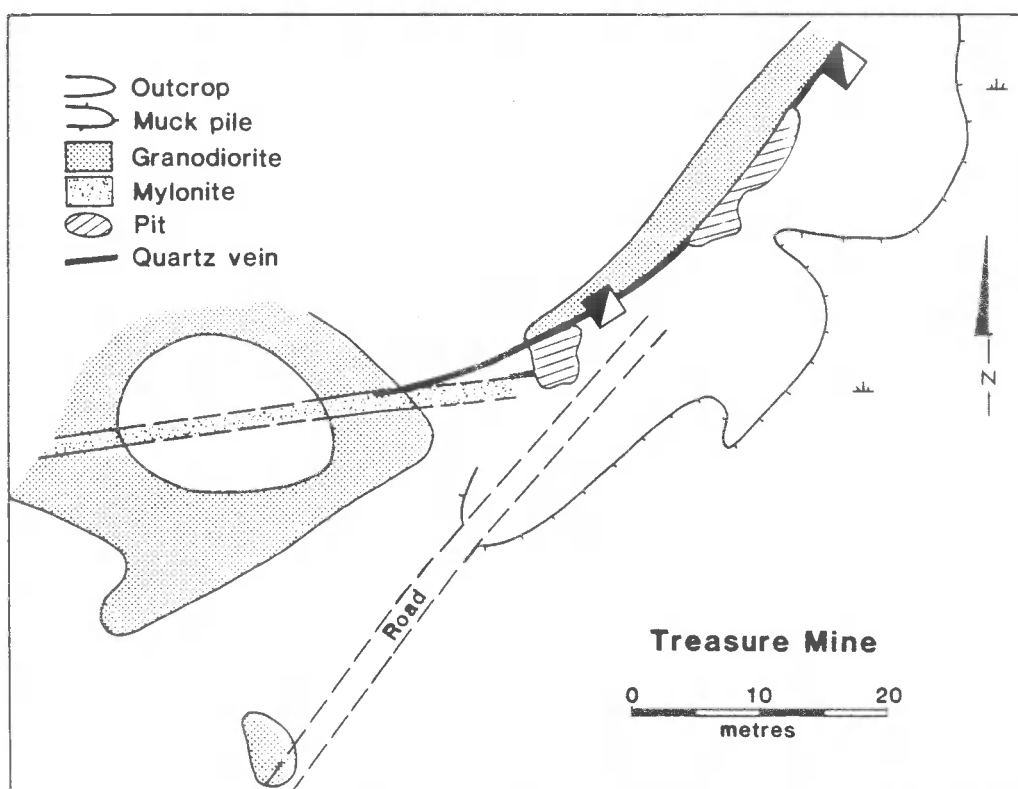


Figure 13: Geology of the Treasure gold occurrence.

ppb..

Replicate analyses of a chip sample across 23 cm of the main vein showed the gold content to be erratic (from 20 ppb to 17.7 ppm). A grab sample of the vein averaged 50 ppm Au and of the footwall schist about 90 ppb Au. Another sample of the mylonite near the shaft contained 12 ppb Au. All samples contained less than 2 ppm silver.

Return to the highway and turn left (southeast) for a distance of 1.6 km. All of the highway cuts are in basalt. Turn right (southwest) down a cottage-access road for about 200 m, where a trail leads south to the Pine Portage Mine.

STOP 7 PINE PORTAGE GOLD MINE (Figure 14)

Originally discovered in 1882, the Pine Portage mine gained early recognition for its abundant visible gold. Within 3 years, a stamp mill had been built on an intermittent stream and some underground development was carried out. Since that time the property has been explored several times.

The prospect is hosted by west-facing, north-north-east trending, pillowed and massive basaltic flows, and lies less than 50 m east of the steeply dipping contact of the Dryberry granodiorite batholith. The granodiorite is massive to weakly foliated and contains a few basaltic inclusions. Granitic dikes in the basalt are rare.

Vein material infills a strongly sheared and faulted zone which is at least 3.5 m wide. Near the inclined shaft the vein strikes approximately north and dips from N65° to 75°E; it has been traced north about 200 m where it curves slightly to the east. The main vein follows the well-defined hanging wall of the zone and has an irregular footwall contact and is commonly wider where the footwall amphibolite is brecciated. The vein consists of discontinuous lenses of quartz, with carbonate and inclusions of schist. An old assay plan shows that the mineralized section had a maximum width of up to 5 m.

Along the hanging wall there is a zone of fault gouge several centimetres wide, characterized by abundant clay minerals. No silicification of this zone was detected and only low gold contents were found. Relative movement is defined by the rotation of a well-developed "S" foliation into the plane of the "C" foliation indicating reverse movement (Simpson and Schmid, 1983). Later "S"-shaped tension gashes indicate normal movement; this may have resulted from relaxation of applied stress. Quartz post-dates the major period of movement, occupying structures parallel to the "C" fabric as well as the tension gashes.

Pyrite is the principal sulphide. Chalcopyrite,

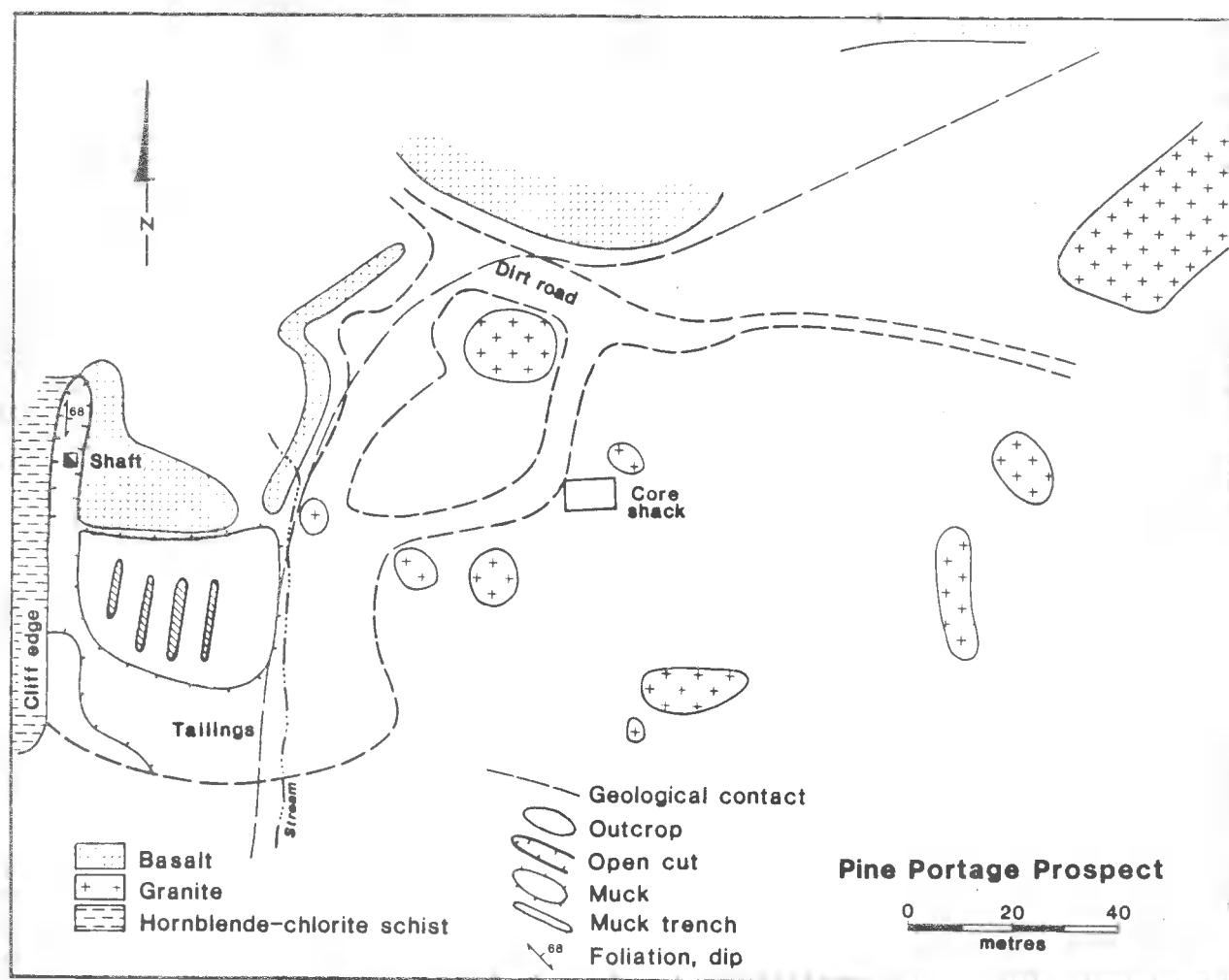


Figure 14: Geology of the Pine Portage gold occurrence.

galena and sphalerite are present in lesser amounts and there is rare pyrrhotite, twinned ilmenite, arsenopyrite, covellite and gold. The main gangue minerals are quartz, ankerite, calcite, and mica and clay minerals. Fine gold (less than 20 microns) is found in the gangue and as inclusions in pyrite; coarser gold is erratically distributed. The larger pyrite grains are zoned and in places enclose silicates, pyrrhotite, sphalerite or chalcopyrite. Sphalerite is generally found in ribbons of mafic schist. Tellurium has been detected by microprobe analysis (Forsgren 1980); calaverite (AuTe_2) may be present. The most characteristic feature of the mineralization is the high silver content, being about twice the gold content; most gold occurrences in the Kenora area have a silver content which is near or below detection limits.

The fault with which the Pine Portage prospect is associated extends into the Dryberry batholith. Mineralization may be largely restricted to irregularities in the fault plane. Forsgren (1980) concluded that "this epigenetic, sheared fissure vein is thought to be the result of wall rock alteration and lateral secretion. The vein is proposed to be a redistribution of host rock chemistry with induced volatiles CO_2 , H_2O and S. Meteoric water may be the source of these volatiles."

Return along the trail which, in part, coincides with the northward extension of the Pine Portage fault. At the highway, across from the road leading to stop 7, irregular lenses and veins of tourmaline and pyrite-bearing quartz occur in the basalts. The contact of the basalts with the Dryberry granodiorite is well exposed in a highway cut, 350m southeast of this point. Continue east for 7.2 km to the junction of Highway 71 and then turn right (south) 5.5 km to Rushing River Provincial Park. From here it is a further 6.4 km to the turnoff onto the Witch Bay Road. All of the rocks along the route thus far, and for the first 5 km along the Witch Bay Road, are part of the Dryberry batholith. The second 5 km of the Witch Bay Road is within basalts which are the approximate stratigraphic equivalent of those at the Pine Portage Mine. At a junction about 10 km from Highway 71, turn to the right and continue 300 m to the site of the former shaft and mill of the Wendigo Gold Mine.

STOP 8 THE WENDIGO GOLD-COPPER MINE (Figure 15)

Gold was first discovered on the property in 1899, when an 18m shaft was sunk on the main vein. Work continued during the next year and about 1000 tons of gold-copper ore were milled. The mine lay idle until 1933 when it was re-examined by Wendigo Gold Mines Ltd. Over the next ten years considerable underground and surface work was done before the mine closed in early 1943. Total production from the property was 67,000 oz

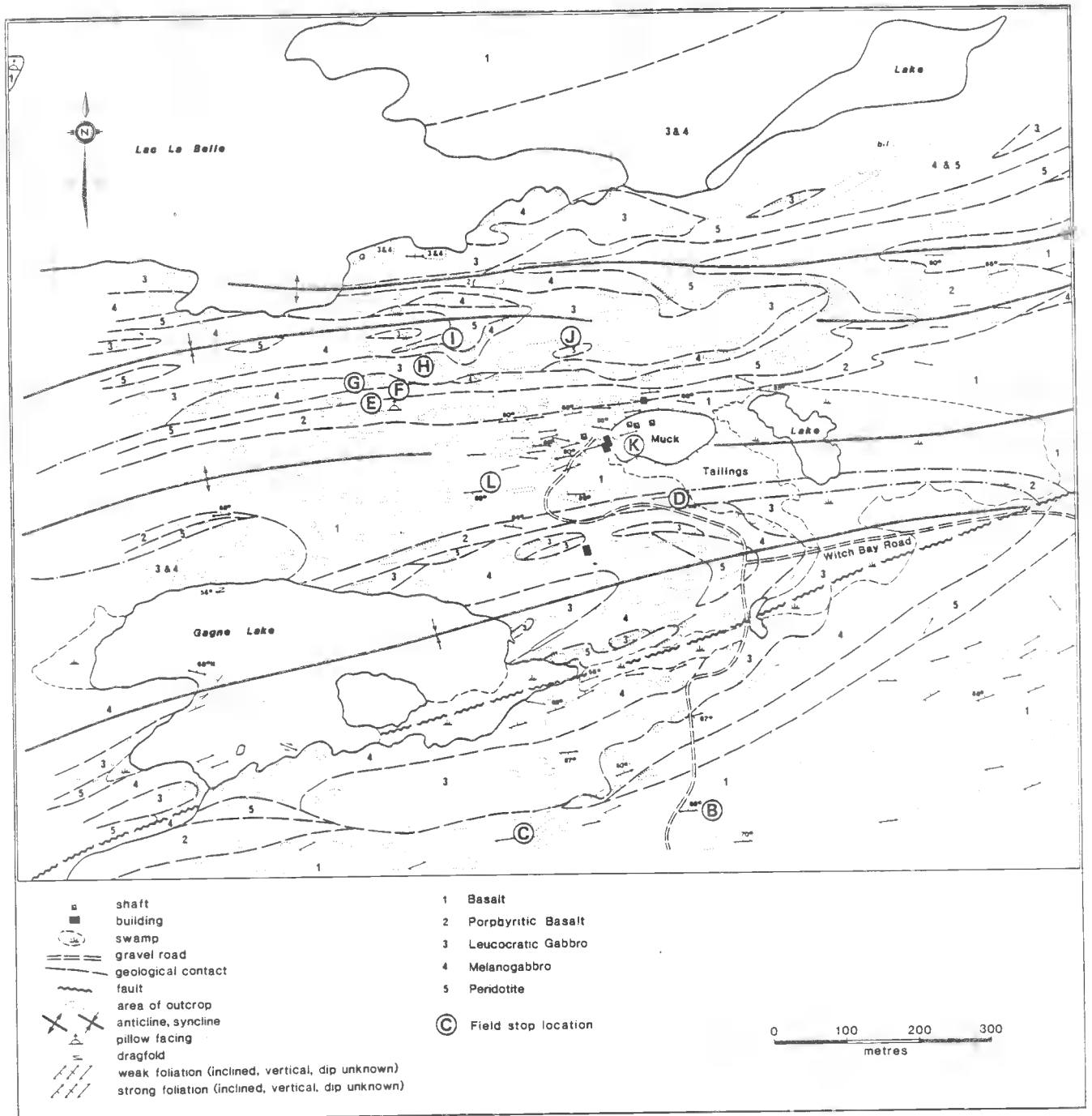


Figure 15: Surface geology of the Wendigo Gold Mine.

of gold, 14,762 oz of silver and 1,886,246 pounds of copper from 206,054 tons milled, establishing the Wendigo as the largest past producer in the western Wabigoon Subprovince. Average grade of ore milled was 0.33 oz of Au/ton.

Wendigo Mine lies on the south limb of the Hay Island antiform (Figures 5 and 6) within mafic volcanics and mafic to ultramafic sills. Metamorphic grade is lower greenschist facies. Analogies between the stratigraphy here and at the Duport Mine on Shoal Lake can be drawn; the two occur close to the top of a lower mafic volcanic cycle.

Felsic pyroclastic rocks are exposed south of the mine along the north shore of Witch Bay. The felsic rocks are intensely deformed, commonly to sericite schist. Only rarely are good relic primary textures preserved. The felsic sequence is the locus of an extensive fault system, the Andrew Bay-Witch Bay fault zone (Figure 5), which Ayer (1984) suggests is the extension of the Crowduck Lake-Rush Bay dextral fault zone (Davies and Smith, 1984). The fault, therefore, may extend from Indian Bay of Shoal Lake to Witch Bay of Lake of the Woods, a distance of over 70 km.

The mine site is primarily underlain by porphyritic and equigranular basalt, which has been intruded by gabbro and peridotite sills. Pillows are observed locally in the basalts, but top determinations are often questionable. Composition varies from tholeiite to high-iron tholeiite. The porphyritic basalt is an excellent marker horizon, and overlies the massive basalt. Pillows observed near the shaft area indicate tops to the north. This basalt is a Mg-tholeiite and characterized by up to 20% white feldspar phenocrysts less than 2.5 cm in diameter.

Much of the property is underlain by gabbro and peridotite comprising thick, differentiated sills. Fine-grained peridotite lies at the base of some sills directly overlying the porphyritic basalt. The peridotite, where less altered, is composed principally of serpentine, talc and magnetite. Peridotite is commonly overlain by melagabbro transitional to leucogabbro. The melagabbro is locally magnetic and displays rhythmic layering, but more commonly is non-magnetic and massive. Amphibole has replaced original pyroxene. The original feldspar:pyroxene ratio was close to 50:50. In places the leucogabbro is porphyritic, containing feldspar phenocrysts up to 3 cm in diameter, similar to those observed within the porphyritic basalt; Coarser peridotite occurs as dikes and as narrow layers in the cores of the sills, suggesting that the sills may consist of more than one cycle. This peridotite locally displays excellent rhythmic layering. Coarse peridotite dikes and/or sills intrude the host volcanics in a few places.

Several tight, east trending, west plunging folds have been identified by means of the reversal of differentiation trends in the sills, and by limited top determinations in the basalts. Basalt located within the anticlinal cores of these folds is strongly foliated to sheared. This is most noticeable in the vicinity of the main shaft and close to Gagne Lake where wide sections of basalt are strongly sheared. Notably, strong deformation is apparently absent from the gabbro bodies. The texture of the leucocratic gabbro shows virtually no strain, and while some strain is evident within both the peridotite and the melagabbro, it does not appear sufficient to be consistent with the tight folding.

Gold mineralization is restricted to zones of silicification, notably to four east-trending steeply north-dipping veins. All the veins were examined while the mine was in operation, but production was restricted to the No. 1 vein. Veins 2, 3 and 4 are located about 460 m east of the main shaft. The attitudes of the veins are slightly discordant to the enclosing lava flows (Figure 16). The veins occupy a zone about 50 m wide which extends to the peridotite contact to the north. They are found within both porphyritic and massive basalt, much of which is largely altered to chlorite schist. The south (ore proximal) side of the peridotite sill has been intensely altered to a soft "talcy" rock. Chlorite schist proximal to the main ore zone is enriched in carbonate and quartz, and contains abundant epidote and zoisite, minor sphene, and lesser amounts of clinozoisite, opaques and sericite.

The No. 1 vein, which strikes N80°E and dips about 79° to the north, maintains an average width of 30 cm, pinching and swelling to a maximum width of 76 cm. Quartz contains pyrite, pyrrhotite and chalcopyrite. Locally, sulphide exceeds quartz. Little of the vein material is presently exposed. Thomson (1936) estimated that, throughout the mine, half the vein material is sulphide, and that the schist immediately adjoining the quartz is commonly well mineralized with sulphides. In general, sulphides are in streaks and lenses which parallel the shearing. Milky-white, unmineralized quartz veins, containing traces of ankerite, may be seen on the surface and underground, locally crosscutting the mineralized quartz, indicating two generations of silicification. Faulting has been observed in places, but offsets are restricted to a few metres.

Petrographic work carried out in 1934 and 1935 (Canada Department of Mines and Resources, 1936) showed native gold to be present as relatively coarse grains within grey translucent quartz: pyrite and chalcopyrite are the dominant sulphides, with minor amounts of pyrrhotite, sphalerite, and arsenopyrite. Brownell (1943) noted distinct mineralogical changes in the vein

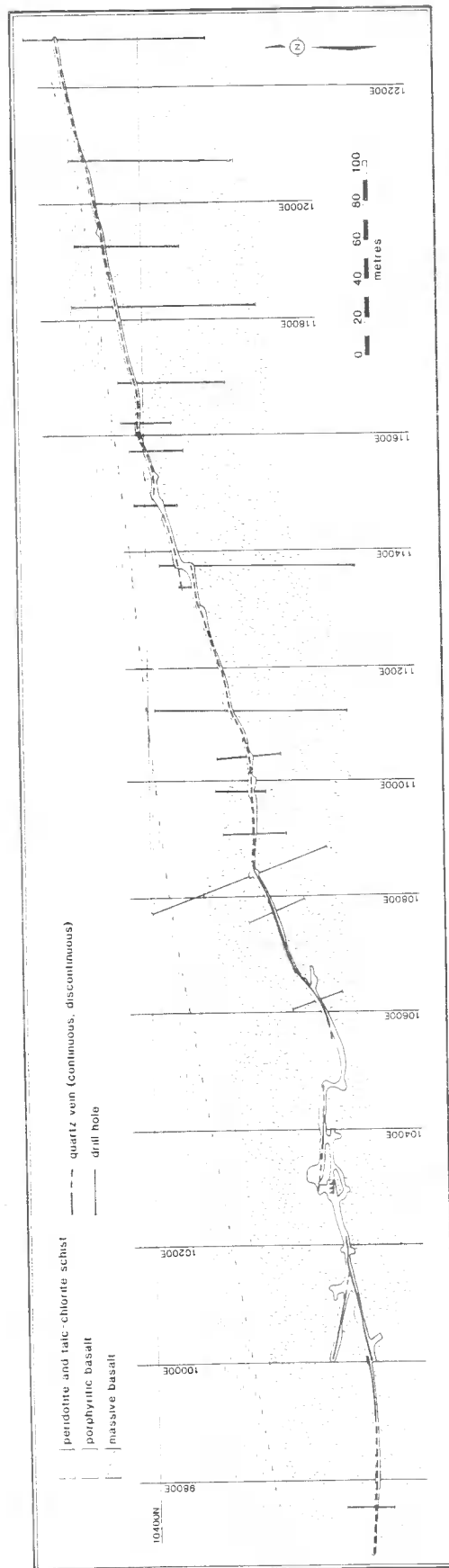


Figure 16: Geology of the 500 level of the Wendigo Gold Mine (after Brownell, 1943).

which were directly related to a sudden decrease in gold content; in the upper levels of the mine, gold was accompanied by pyrite and chalcopyrite, whereas below the 335m level pyrrhotite is the dominant sulphide and gold contents are much lower. Brownell (1943) suggested that this relationship is essentially temperature dependent, ie. pyrrhotite is normally deposited at higher temperatures than pyrite, chalcopyrite and gold. He concluded that, below the 335m level, no additional ore might be expected.

The area is characterized by intense deformation and probably lies within the influence of the Andrew Bay-Witch Bay fault zone. Since the rock assemblage is heterogeneous, most of the strain was taken up by the least competent rocks (i.e. massive and porphyritic basalt). The contrasting competencies of the two major rock types (intrusive vs extrusive) resulted in zones of dilatancy being created along the contact between the two units. These zones of dilatancy acted as permeable channelways along which hydrothermal fluids moved upwards. The vertical zonation noted by Brownell (1943) may represent two distinct stability fields with the 335 m level corresponding to the stability field boundary.

Locality A This locality is slightly east of the map boundary and shows relatively undeformed porphyritic basalt.

Locality B Relatively undeformed massive basalt.

Locality C Slightly elongate pillows within fine-grained basalt. Compare this outcrop with more-deformed rocks which will be seen later.

Locality D At this outcrop of highly deformed porphyritic basalt, remnant feldspar phenocrysts are barely recognizable. This rock is rich in carbonate and fine pyrite.

Locality E From the old mine site, follow the trail to the west passing over varieties of chlorite schist which have undergone various degrees of deformation. At this locality, virtually undeformed pillowed porphyritic basalt faces north. From this point we will make a traverse north across the differentiated sill.

Locality F On the north side of a valley, which is interpreted to be underlain by talc schist, peridotite displays little sign of deformation and is representative of the fine-grained, basal portion of the differentiated sill.

Locality G Melagabbro in these outcrops is generally magnesium rich, and locally contains segregations,

inclusions or veinlets of leucogabbro. In places a weak foliation is evident.

Locality H These rocks are representative of the transitional zone between melagabbro and leucogabbro. Feldspar-actinolite ratio is close to 50:50, and the rock is generally massive, showing little sign of strain.

Locality I Between localities H and I, coarse feldspar phenocrysts occur locally within the leucogabbro, indicating the top of the first ultramafic to mafic cycle. Primary layering within outcrops of coarse melagabbro and peridotite can be observed at this locality.

Locality J At this location a coarse peridotite dike intrudes leucogabbro. Contacts are sharp and serpentine fills fractures in the adjacent gabbro.

Locality K Return to the mine site, passing over deformed, pillowed, porphyritic basalt and chlorite schist. The only present exposure of the main Wendigo vein is, unfortunately, quite poor. Here, the chlorite schist can be examined further, as there are several good vertical exposures in the general area. The adjacent muck-pile contains the last material taken from the mine, and probably came from below the 335 m level.

Return to Highway 71, making one additional stop along the way. At a distance of 3.2 km from the road junction south of the Wendigo mill site, turn sharp left (north) onto a forest-access road. Approximately 500 m along this road a widening occurs at a bend: this is the final stop on the field trip. A trail leads north about 60 m to Kite Lake, and half way along this trail are the workings of Witch Bay Gold Mines Ltd.

STOP 9 WITCH BAY GOLD PROSPECT

The geology in the vicinity of Kite Lake has not been mapped in detail. Based on aeromagnetic data it would appear that the principal lithologic units of the Wendigo mine area (ie. basalt, gabbro and peridotite), extend east-northeast as far as the central part of Kite Lake, where they terminate against granodiorite of the Dryberry batholith.

Fine-grained basalt underlies an area south of Kite Lake. Near the southernmost tip of Kite Lake an east-striking zone of schist and breccia, from one to two metres wide and dipping south at from 55° to 70°, contains some carbonate and one or more quartz veins and lenses with a maximum width of 45 cm. The quartz is fractured, encloses minor silicified basalt, and contains abundant chalcopyrite and pyrite in and near

some of the fractures. The mineralization and its stratigraphic position is similar to that at the Wendigo.

Two shafts about 30 m apart have been sunk on the zone. A chip sample across 50 cm of the mineralized quartz vein at the east shaft assayed 1.93 ounces of gold per ton. A third shaft was put down a further 75 m to the east. A fourth shaft was sunk on a separate zone approximately 100 m to the southwest. The zone is not exposed, but material on the dump indicates that it is carbonatized, sheared basalt with minor quartz and with chalcopyrite, pyrite and pyrrhotite.

References

- Ayer, J.A., 1984, Geology of the Bigstone Bay Area, District of Kenora; p.19-24 in Summary of Fieldwork, 1984, Ontario Geological Survey, Edited by John Wood, Owen L. White, R.B. Barlow, and A.C. Colvine, Ontario Geological Survey, Miscellaneous Paper 119, 309p.
- Ayer, J.A., Gil, K.M., McCurdy, M., and assistants, 1984, Precambrian Geology of the Bigstone Bay Area, Northern Part, Kenora District (52 E/9). Scale 1:15 840 or 1 inch to 1/4 mile.
- Ayer, J.A., Gil, K.M. McCurdy, M., and assistants, 1984, Precambrian Geology of the Bigstone Bay Area, Southern Part, Kenora District (52 E/9). Scale 1:15 840 or 1 inch to 1/4 mile.
- Blackburn, C.E., 1981, Kenora-Fort Frances sheet. Ontario Geological Survey, Geological Compilation Series, Map 2443, scale 1:253,440.
- Blackburn, C.E., Bond, W.D., Breaks, F.W., Davis, D.W., Edwards, G.R., Poulsen, K.H., Trowell, N.F. and Wood, J., 1985, Evolution of Archean Volcanic-Sedimentary Sequences of the Western Wabigoon Subprovince and its Margins: A Review; in Evolution of Archean Supracrustal Sequences, Edited by Ayers, L.D., Thurston, P.C., Card, K.D., and Weber, W., G.A.C. special Paper 28.
- Brownell, G.M., 1943, Wendigo Gold Mines Ltd., Assessment Files, Kenora.
- Blackburn, C.E., and Janes, D.A., 1983, Gold Deposits in Northwestern Ontario; p.194-210 in The Geology of Gold in Ontario, edited by A.C. Colvine, Ontario Geological Survey, Miscellaneous Paper 110, 278p.
- Canada Department of Mines and Resources, 1936, Ore Dressing and Metallurgical Investigation No.621; Investigations in Ore Dressing and Metallurgy, Jan.-Jun., 1935, p.103-116.
- Davies, J.C., 1965, Geology of the High Lake-Rush Bay Area, District of Kenora; Ontario Department of Mines, Geological Report 41, 57p; 2 maps, 1 inch to 1/2 mile.
- Davies, J.C., 1983, Gold Deposits of the Lake of the Woods Area, Kenora District; p.241-245 in Summary of Field Work, 1983, by the Ontario Geological Survey, edited by John Wood, Owen L. White, R.B. Barlow, and A.C. Colvine, Ontario Geological Survey, Miscellaneous Paper 116, 313p.
- Davies, J.C. and Smith, P.M., 1984, The Structural and Stratigraphic Control in the Lake of the Woods Area; p.185-193 in Summary of Fieldwork, 1984, Ontario Geological Survey, Edited by John Wood, Owen L. White, R.B. Barlow, and A.C. Colvine, Ontario Geological Survey, Miscellaneous Paper 119, 309p.
- Forsgren, F.M., 1980, A study of the Pine Portage Prospect, Kenora District, Ontario, Unpublished B.Sc. Thesis, University of North Dakota.
- King, H.L., 1983, Precambrian Geology, Kenora-Keewatin Area, Eastern Part (NTS 52 E/16SW) Kenora District, Ontario; Ontario Geological Survey, Preliminary Map P.2618, scale 1:15,840 or 1 inch to 1/4 mile.

- Pedora, J.M., 1976, Mineralization of the High Lake Pluton and Adjacent Country Rocks; Unpublished M.Sc. Thesis, University of Manitoba, Winnipeg, Manitoba.
- Simpson, C. and Schmid, S.M., 1983, An Evaluation of Criteria to Deduce the Sense of Movement in Sheared Rocks. GSA Bull., V.94, p.1281-1288, 11 figures, November 1983.
- Smith, P.M., 1984, The Geological Setting of the Duport Mine, Cameron Island, Shoal Lake; p.194-200 in Summary of Fieldwork, 1984, Ontario Geological Survey, Edited by John Wood, Owen L. White, R.B. Barlow, and A.C. Colvine, Ontario Geological Survey, Miscellaneous Paper 119, 309p.
- Sutherland, I.G., and Colvine, A.C., 1979, The Geology and Mineralization of the Pickerel Arm, Canoe Lake, and High Lake Bodies; p.233-243 in Summary of Field Work, 1979, by the Ontario Geological Survey, edited by V.G. Milne, O.L. White, R.B. Barlow, and C.R. Kustra, Ontario Geological Survey, Miscellaneous Paper 90, 245p.
- Thomson, J.E., 1936, Gold deposits on the Lake of the Woods; Ont. Dept. Mines, Annual Report for 1935, Vol.44, Pt.4, p.29-47.
- Thomson, J.E., 1937, Geology of the North Central Part of the Lake of the Woods, Kenora District, Ontario: Ontario Department of Mines, Annual Report for 1936, Volume 45, Part 3, p.1-43, Accompanied by Map 45b, scale 1:63,360 or 1 inch to 1 mile.
- Trowell, N.F., Blackburn, C.E. and Edwards, G.R., 1980, Preliminary Synthesis of the Savant Lake - Crow Lake Meta-volcanic, Metasedimentary Belt, Northwestern Ontario, and its bearing upon Mineral Exploration; Ontario Geological Survey, Miscellaneous Paper 89, 30p.

Geological Relationships in
the Vicinity of
the Wabigoon - Winnipeg River
Subprovincial Interface in the Kenora area¹

Introductory Discussion and Field Guide
31st Annual Institute on Lake Superior Geology
Kenora, Ontario

by

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THE RELATIONSHIP OF
SUPRACRUSTAL SEQUENCES
TO A BASEMENT COMPLEX IN THE WESTERN ENGLISH RIVER
SUBPROVINCE*

* This introductory statement is a modified version of a paper of the same title published elsewhere (Beakhouse, 1985). It is intended as a introduction to the regional context of the field trip area.

ABSTRACT

Three distinctly different types of terrane are recognized within and immediately adjacent to the English River Subprovince. Greenstone-tonalite terrane forms the Uchi and Wabigoon Subprovinces whereas the Ear Falls -- Manigotagan belt (EFMB) and Winnipeg River belt (WRB) of the English River Subprovince are paragneiss and batholithic terranes, respectively. The EFMB is a major, linear sedimentary basin with detritus derived in adjacent terrains from laterally equivalent volcanic rocks and granitic rocks. The greenstone-tonalite - paragneiss interface represents a facies change between fundamentally different depositional environments. In the WRB, rare remnants of a predominantly volcanic supracrustal sequence occur as inclusions within highly metamorphosed and deformed plutonic rocks (orthogneiss). Field relationships and geochronological investigations indicate that orthogneiss, together with certain sodic plutons, predate widespread 'Kenoran' volcanism in the Wabigoon Subprovince to the south. The relatively potassic composition and slightly elevated Sr isotope initial ratios of late plutons from the central axis of the WRB suggest that they are derived from this earlier crust.

Volcano-plutonic activity, perhaps analogous to that observed in the greenstone-tonalite terranes, was initiated prior to 3.0 Ga and led to the development of a 'pre-Kenoran' sialic microcontinent now represented by the WRB. Synchronous development of volcano-sedimentary sequences in greenstone-tonalite terrane and greywacke-turbidite sequences in paragneiss terrane occurred adjacent to, and in part on, this earlier crust. This earlier crust was greatly thinned, or absent, beneath much of the Wabigoon subprovince. The interfaces between different terranes represent either facies transitions (EFMB-Uchi) or unconformities (WRB-EFMB and WRB-Wabigoon) that are loci for subsequent deformational and igneous processes that obscured the primary relationships between different terranes.

INTRODUCTION

The English River belt was first defined as a subprovince of the Superior Province by Wilson and Brisbin (1963). More recent geologic mapping, particularly Project Pioneer (Manitoba Mines Branch and University of Manitoba; McRitchie and Weber, 1971) and Operation Kenora - Ear Falls (Ontario Geological Survey; Breaks et al. 1978), has shown that the English River Subprovince is actually compound, with the Ear Falls - Manigotagan belt in the north and the Winnipeg River belt in the south (Beakhouse 1977; Breaks and Bond, 1977). Breaks et al. (1978) refer to these belts as the Northern Supracrustal domain and Southern Plutonic domain, respectively. Regional geophysical investigations (see references in Beakhouse 1977) have added important information on the third dimension and, on the basis of a re-evaluation of geophysical data, Hall and Brisbin (1981) suggested that these belts should be elevated to subprovince status. The Ear Falls - Manigotagan and Winnipeg River belts represent two fundamentally different types of "high-grade" terrane, referred to here as paragneiss and batholithic terrane, respectively. Both are distinctly different from greenstone-tonalite or low-grade terranes which occur in the adjacent Wabigoon and Uchi sub-provinces (Figure 1).

This paper will discuss the nature of these two types of 'high-grade' terrane and their associated supracrustal rocks and their relationship to each other and the adjacent greenstone-tonalite terranes. Emphasis is placed on the broad relationships between the various supracrustal assemblages and a large basement complex that pre-dates some of these supracrustal rocks. The English River Subprovince, together with adjacent parts of the Wabigoon and Uchi greenstone-tonalite Subprovinces (Figure 1), is unique in that all 3 types of terrane and all possible combinations of interfaces between these terranes are represented.

LITHOLOGIES

The rocks in the English River Subprovince can be broadly grouped into 7 suites. In approximate order of decreasing age these are (1) high-grade supracrustal (predominantly meta-volcanic) remnants; (2) orthogneiss; (3) paragneiss; (4) internal greenstone belts; (5) sodic plutons; (6) potassic plutons; (7) mafic plutons (Figure 2). Lithologies will be described only briefly, more complete lithologic descriptions are presented by Breaks et al. (1978) and Beakhouse (1983).

Supracrustal Remnant Suite

The orthogneiss suite, and to a lesser extent the plutonic suites, contain minor, widely distributed, highly metamorphosed and deformed volcanic and rare sedimentary inclusions. These inclusions are usually small (<20m) and discontinuous or disrupted, but can be up to 1 km thick and 5 km long. Where the inclusions occur in orthogneiss, intrusive contact relationships indicate that most supracrustal remnants are older than the tonalitic orthogneiss (Gower, 1978; Gower and Clifford, 1981; Beakhouse, 1983; Beakhouse et al. 1983). However, some

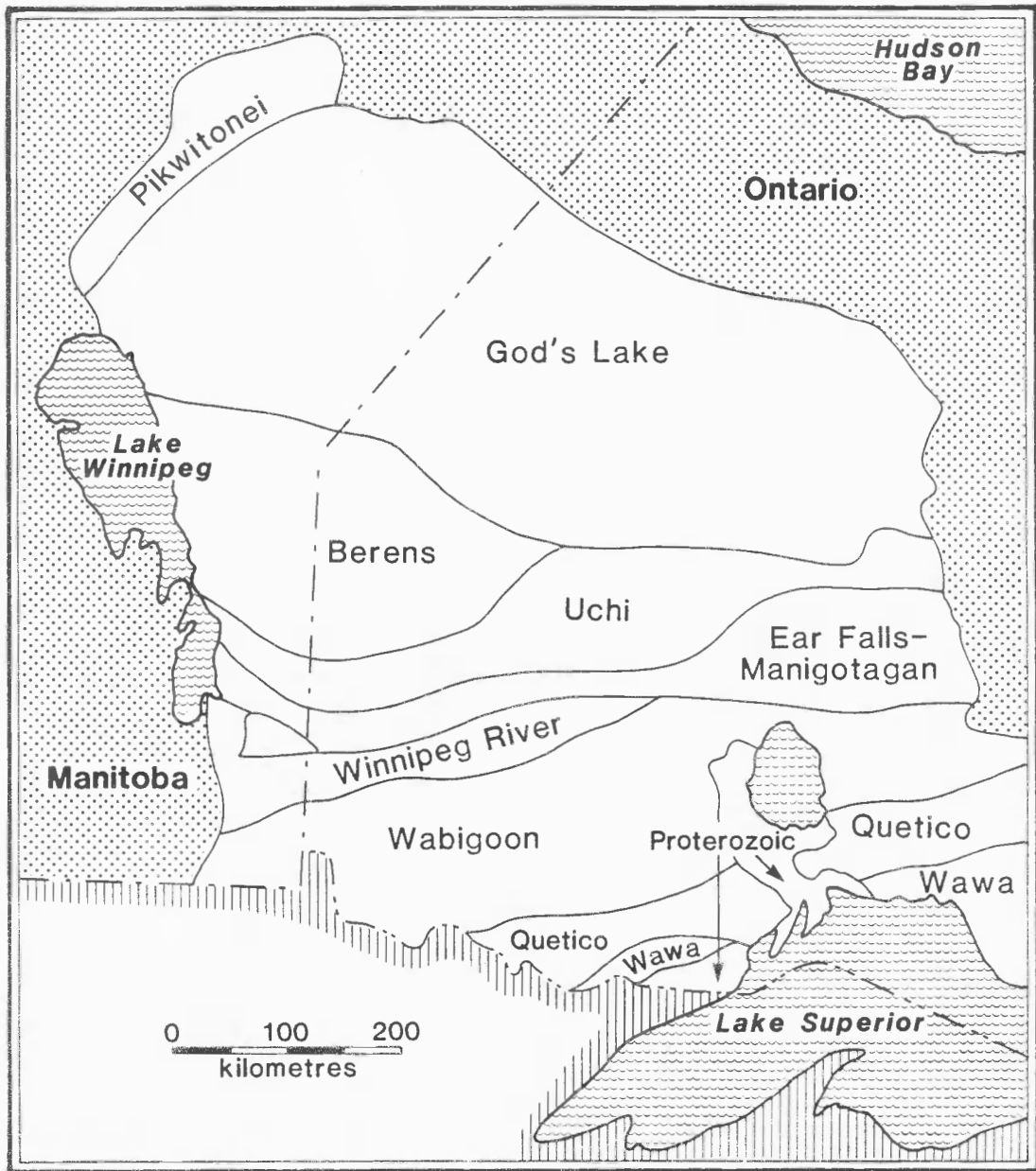


Figure 1: Map illustrating the distribution of various types of terranes in the western Superior province. Greenstone-tonalite terranes include the Sachigo, Uchi, Wabigoon and Wawa subprovinces. The Ear Falls-Manigotagan belt and Quetico subprovince are paragneiss terranes. The Berens River subprovince and Winnipeg River belt are batholithic terranes.

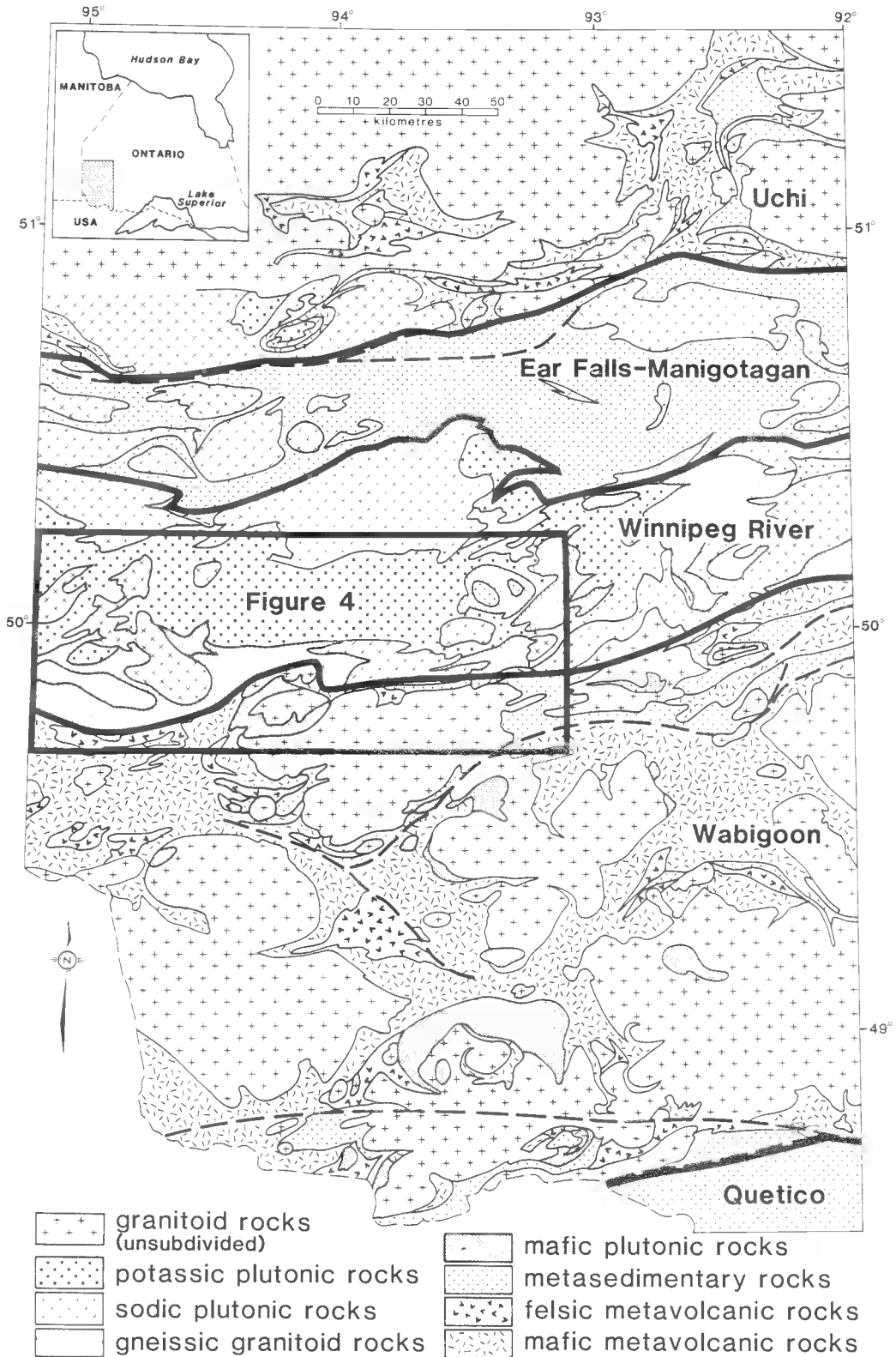


Figure 2: Geologic map of part of the western Superior province including parts of the Uchi, Ear Falls - Manigotagan, Winnipeg River, Wabigoon and Quetico subprovinces. The geology is compiled primarily from Ferguson et al. (1970), Blackburn (1981) and Breaks et al. (1978).

inclusions may be tectonic inclusions of younger units. Amphibolite inclusions predominate. Some inclusions are volcanic as indicated by the local presence of highly deformed pillow structures. In places the pillowed units are interlayered with compositionally and texturally similar, massive amphibolites and with coarse-grained amphibolites, probably representing a sequence of massive and pillowed basaltic flows and related gabbro sills. On the other hand, many of the inclusions are massive amphibolite of uncertain origin. Some of these inclusions are mineralogically and texturally similar to the volcanic amphibolite and may also be volcanic. Others, however, represent disrupted, younger gabbroic sills and dikes.

Felsic volcanic rocks have not been conclusively identified. The paucity of felsic volcanic rocks may be a primary feature of this early volcanism or may reflect their less refractory nature during subsequent metamorphism and partial melting (Beakhouse et al. 1983). Sedimentary units are rare and comprise medium- to thick-bedded arkose and greywacke, with interlayered pelitic and calc-silicate layers. Thin-bedded, oxide-facies iron formation is recognized on the south shore of Umfreville Lake and in the Cedar Lake area (Westerman, 1977; Beakhouse, 1983).

The widespread distribution and occurrence as inclusions in old tonalites (discussed below) suggests that the volcanic amphibolite, the metasedimentary, and possibly many of the massive amphibolite inclusions are remnants of an old, largely volcanic, supracrustal sequence. This sequence may have been similar to the greenstone terranes of the better preserved Uchi and Wabigoon Subprovinces, and appears to have been coextensive with the Winnipeg River belt. It is the earliest stage in the development of the belt. The less abundant metavolcanic and metasedimentary inclusions in the younger plutonic suites may have been derived either from this older supracrustal sequence or from a younger sequence.

Orthogneiss Suite

The orthogneiss suite (Figure 2) is a compositionally and texturally heterogeneous assemblage of layered gneisses. Felsic plutonic components predominate and undeformed intrusive rocks and inclusions of supracrustal rocks are a minor component.

The gneissic phases range in composition from quartz diorite to granite. Fine- to medium-grained, biotite or hornblende-biotite tonalitic gneiss is the oldest plutonic component and contains inclusions of the older supracrustal sequence. This was intruded by a wide variety of fine-grained to pegmatitic, quartz dioritic to granitic phases, in which the gneissosity ranges from weakly to strongly developed, and by mafic dikes. The mafic dikes are deformed and metamorphosed to amphibolite facies; they intruded many of the post-tonalite granitoid phases but are intruded by others.

The gneissosity is a result of post-emplacement deformation and metamorphism which affected the supracrustal remnants, the various granitoid phases, and the mafic dikes. Another effect of the metamorphism was the development, in many of the rock units,

of concordant, discontinuous, leucocratic anatectic lenses that are commonly less than 1 cm thick. The composition of these lenses is a function of source rock composition and degree of melting. Common associations are quartz dioritic to tonalitic lenses in amphibolite, tonalite to granodioritic lenses in tonalitic gneiss, and granodioritic to granitic lenses in granitoid gneiss.

The variable degree of gneissosity development in the post-tonalite granitoid phases, and the differing age relationships between these phases and the mafic dikes indicate a considerable age span of granitoid emplacement prior to and during the deformation. Pegmatitic and aplitic granitic dikes and plugs that postdate major deformation are also common. They are probably temporally and genetically related to the potassic plutonic suite.

The complex nature of the orthogneiss suite results from the intimate admixture of these variably deformed rock types. Heterogeneity is manifest on two scales. On the outcrop and hand specimen scale the distinctive layered appearance is a consequence of: 1) primary sedimentary and volcanic layering (minor), 2) transposition of originally discordant structures into essential parallelism by intense deformation, similar to that described by Myers (1978), 3) pre-emplacement anisotropy such as layering and foliation controlling the geometry of intrusive phases, and 4) recrystallization and partial melting synchronous with layer-normal, principal stress resulting in foliation and lenses of anatectic melt being oriented parallel to layering.

A larger scale layering, metres to hundreds of metres thick, is defined by units characterized by particular assemblages such as tonalite + amphibolite or granitoid gneiss + anatectic layers. These units are not continuous along strike.

Paragneiss

Paragneiss that is composed principally of plagioclase, quartz and biotite with common almandine garnet and minor K-feldspar, cordierite, sillimanite and muscovite is the dominant rock type of the Ear Falls-Manigotagan belt. In many places original sedimentary bedding, 5 to 70 cm thick, can be recognized by variations in mineralogy, texture, and susceptibility to anatexis. The bedding is particularly well defined by inter-layered fine- to medium-grained, granoblastic gneiss representing psammite of probable greywacke composition, and medium to coarse-grained, porphyroblastic gneiss representing semipelite or siltstone. The reversal of grain size with the psammite being finer grained than semi-pelite is a consequence of metamorphic recrystallization. Normal graded bedding is recognized rarely by progressive changes in the abundance of biotite and garnet which reflect original variations in clay content (Van de Kamp and Beakhouse, 1979). Other sedimentary structures such as cross-bedding and slump structures are rarely preserved. Pegmatitic leucosome originating from the partial melting of these gneisses is abundant and widely distributed. Metamorphic mineral assemblages and partial melting indicate widespread, upper

amphibolite facies metamorphism (McRitchie and Weber, 1971b; Freund and Turnock, 1971; Harris, 1976) and local low pressure granulite facies conditions (Thurston and Breaks, 1978).

Because of the metamorphic grade, the sedimentology of the paragneiss progenitor is difficult to decipher, but van de Kamp and Beakhouse (1979) considered that the paragneiss represents turbidites deposited below wave base. The provenance is also difficult to resolve because the primary depositional mineralogy must be inferred from metamorphic mineral assemblages and chemical compositions. The chemical and mineralogical composition of the least migmatitic paragneiss is suggestive of immature sediments (van de Kamp and Beakhouse, 1979). These authors suggested that relatively direct derivation from felsic volcanism explains certain of the chemical attributes of the paragneiss. Breaks et al. (1978), on the other hand, argued that trace element abundances require a variety of source materials; high Cr and Ni reflect mafic or ultramafic sources and elevated Ba and Rb felsic sources.

The interfaces with the adjacent Winnipeg River belt on the south and the Uchi Subprovince on the north are modified by faults and plutonic activity, but primary stratigraphic relationships have been preserved in some areas. In the Rice Lake area of Manitoba, the paragneisses are considered to be, in part, a lateral facies equivalent of the Edmunds Lake Formation of the Uchi Subprovince (Campbell, 1971; McRitchie and Weber, 1971b). In Ontario, Thurston and Breaks (1978) noted that the paragneiss is stratigraphically equivalent to Cycle I volcanic rocks in the Confederation Lake area of the Uchi Subprovince. Cerny et al. (1981) recognized a similar relationship on the south side of the Ear Falls-Manigotagan belt where the paragneisses are postulated to be stratigraphically equivalent to the Flanders Lake Formation in the Bird River greenstone belt of Manitoba. Metaconglomerates occur in both the Edmunds Lake and Flanders Lake Formations and locally at other places near the margins of the Ear Falls-Manigotagan belt (e.g., Breaks et al., 1978), and contain a variety of clasts derived from mafic to felsic volcanic units, mafic to felsic plutons, and sedimentary units including ferruginous chemical metasediments (Bateman, 1939; Breaks et al. 1978; Cerny et al. 1981; Campbell, 1971). These stratigraphic relationships suggest that more proximal facies of the paragneiss may occur in the greenstone-belt, volcano-sedimentary sequences flanking the Ear Falls-Manigotagan belt.

The provenance of the paragneiss was apparently the adjacent greenstone-granitoid terranes and was variable in composition, although volcanoclastic metasediments may be regionally or locally dominant (cf. Ojakangas, 1985).

Internal Greenstone Belts

Two relatively well preserved greenstone belts occur entirely within the English River Subprovince, along the boundary between the Winnipeg River and Ear Falls - Manigotagan belts. The Bird River greenstone belt, the larger and better preserved of the two belts comprises mafic to felsic metavolcanic rocks, clastic metasedimentary rocks and mafic to ultramafic intrusions

(Davies et al. 1962; Karup-Moller and Brummer 1971; Trueman 1975; Cerny et al. 1981). Regional metamorphic grade increases outward from greenschist to almandine-amphibolite as the belt grades into paragneiss of the Ear Falls - Manigotagan belt (Butrenchuk 1970).

Felsic metavolcanic and clastic metasedimentary rocks are more abundant than in most greenstone belts of the Wabigoon and Uchi Subprovinces, but this may reflect the relatively small size of the greenstone belt and partial engulfment by younger plutons rather than any primary differences. Other greenstone belts, however, do lack intrusions comparable to the layered, Cr-enriched Bird River sill.

The Separation Lake greenstone belt is mainly massive and pillowed metabasalt flows with subordinate gabbro and peridotite intrusions and oxide-facies iron formation; metamorphic grade is amphibolite facies (Beakhouse, 1975). The belt is in fault contact with paragneiss to the north and the southern margin is a 1 to 2 km wide migmatitic zone with younger plutons.

Numerous gabbro, and iron formation rafts are widely distributed in the younger plutonic suites, but are especially abundant along the north margin of the Winnipeg River belt east of Separation Lake where they may be an extension of the Separation lake greenstone belt, and in the Sand Lake area south of the Separation Lake belt. These rafts lack the intense deformation and partial anatexis that characterizes the early supracrustal remnants.

As noted previously the greenstone-belt sequences appear to be stratigraphically equivalent to the paragneisses of the Ear Falls-Manigotagan belt (Cerny et al., 1981) and to greenstone-belt sequences of the Uchi Subprovince. Contacts between the greenstone belt-paragneiss assemblage and the older orthogneiss-supracrustal remnant assemblage have not been observed. However, as will be discussed later, the relative ages of the two assemblages can be deduced from indirect field evidence and geochronologic data.

Plutonic Rocks

Two felsic plutonic suites, sodic and potassic, underlie 80% of the Winnipeg River belt and 50% of the western English River Subprovince (Figure 2). Potassic plutons commonly intruded adjacent sodic plutons. A third (mafic) suite is not abundant.

The sodic suite occurs in two settings: 1) intruded into orthogneiss near the margins of the Winnipeg River belt, and 2) as diapiric intrusions into the paragneiss of the Ear Falls - Manigotagan belt. The sodic suite includes quartz diorites, tonalites and granodiorites that are commonly recrystallized and, in some cases, deformed. Primary igneous textures are recognizable but are commonly modified by the development of a foliation or lineation. Weakly gneissic fabrics are present locally, near contacts, and probably originated in response to laminar flow during emplacement although certain units appear to pre-date deformation.

The potassic suite is most abundant in the west-central part of the Winnipeg River belt where it forms large batholiths

(Figure 2). It also occurs as stocks, dikes and sills throughout the English River Subprovince (Breaks et al., 1978). This suite includes granodiorites and granites that, except near late faults, lack the secondary recrystallization foliation characteristic of the sodic suite. Primary planar and linear flow structures are rare but well preserved.

Inclusions of virtually every older suite are found in the potassic suite but individual units are characterized by specific assemblages of inclusions. Potassic plutons in the Winnipeg River belt are characterized by amphibolite, orthogneiss and variously recrystallized quartz dioritic to granodioritic inclusions whereas the relatively small plutons in the Ear Falls - Manigotagan belt contain variously digested paragneiss inclusions.

EVIDENCE FOR 'PRE-KENORAN' CRUST IN THE WINNIPEG RIVER BELT

The Kenoran Orogeny is the culminating Archean orogenic event in the Superior Province. It was originally dated as the period between 2390 Ma and 2590 Ma (Stockwell, 1964) based on K-Ar mineral ages. These ages are now widely regarded as representing minimum cooling ages and the age and definition of the orogeny has been revised repeatedly (e.g. Gower and Clifford, 1981, Harland, 1983). Goodwin (1977) concluded, largely on the basis of U-Pb zircon ages, that most metavolcanic and metasedimentary rocks of the western Superior Province were deposited between 2710 and 2760 Ma, were subsequently deformed and metamorphosed by the Kenoran orogeny and had been intruded by late- or post-tectonic plutons by 2660 Ma. Recent U-Pb zircon data from the western Wabigoon Subprovince (Hart and Davis, 1969; Davis et al. 1982; Davis and Trowell, 1982; Davis and Edwards, 1982; Blackburn et al., 1984) agree with this conclusion, and indicate that volcanic and associated plutonic activity here apparently occurred over the restricted time interval of 2703 to 2755 Ma, and that post-tectonic plutons were emplaced by 2695 Ma. This is a useful standard against which to evaluate isotopic ages from the Winnipeg River belt because of the high precision and consistency of the ages and the proximity to the Winnipeg River belt.

Goodwin (1977) also recognized an older (>3000 Ma) Lac Seul event, and subsequent work has recognized metavolcanic and meta-plutonic rocks intermediate in age between 2760 and 3000 Ma (Nunes and Thurston, 1980; Beakhouse, 1983; Corfu et al., 1985). These intermediate ages complicate Goodwin's two-fold distinction and suggest that, in some parts of the western Superior Province, crustal evolution may have proceeded semi-continuously for 400 Ma. To facilitate discussion the terms 'Kenoran' and 'pre-Kenoran' are used informally here to refer to rocks younger and older, respectively, than 2760 Ma. It is emphasized that these terms are adopted for convenience only and without any intent to formally re-define the Kenoran Orogeny.

Field Evidence

The orthogneiss suite intruded the older supracrustal remnant suite and most phases of the orthogneiss suite predate

the plutonic suites. The orthogneiss is interpreted to be older than the paragneiss, internal greenstone belts and Wabigoon subprovince volcanism based on several indirect field relationships.

- 1) The orthogneiss is cut by mafic dikes (Bald, 1981). Some of these dikes, such as those in the Tannis Lake area (6 on Fig. 2), may be feeders to Kenoran volcanism (Clark et al. 1981), although Gower (1978) has noted that some dikes in the Kenora area are calc-alkaline and are probably not related to the basal tholeiitic volcanism in the adjacent Wabigoon Subprovince.
- 2) Polymictic conglomerate at Perrault Lake in the Winnipeg River belt contains massive and foliated plutonic clasts (Breaks et al. 1978). It is unlikely that the foliated plutonic clasts were derived from synvolcanic plutons, and, although it cannot be established that these clasts were derived from the orthogneiss, they do demonstrate the existence of an elevated granitoid terrane prior to the late plutonism.
- 3) In the western end of the Ear Falls-Manigotagan belt, orthogneiss occupies the cores of antiforms and structurally underlies the paragneiss representing a possible basement-cover relationship (Ermanovics et al., 1979). However, tectonic juxtaposition of these units is also possible. In the Cedar Lake area, Westerman (1979) considered that the Twilight paragneiss in the cores of domal structures is a result of subhorizontal tectonic interleaving of basement and cover sequences.
- 4) The earliest deformation that affected the orthogneiss in the Cedar Lake area may not be present in the Twilight paragneiss (Westerman, 1977) implying that the tonalites were emplaced and deformed prior to the deposition of the paragneiss progenitor. Similarly, deformation in the paragneisses of the Ear Falls-Manigotagan belt appears to be less complex than in adjacent orthogneiss. However, it must be stressed that the apparent differences in structural complexity of paragneiss and orthogneiss could also be the result of differing ductility contrasts and rheologic states during the same deformation event.

The field relations outlined above give inconclusive results about the age relationships between the greenstone belt-paragneiss assemblage and the orthogneiss. At the present state of knowledge alternative explanations are possible for all of the field relations, a common situation in such complexly deformed terranes. In spite of these problems, the field relations can be interpreted to indicate that the orthogneiss is older than the greenstone belt-paragneiss assemblage and is pre-Kenoran. This interpretation is supported by geochronologic data.

Geochronological Evidence

A direct approach to establish the existence and distribution of pre-Kenoran crust is the isotopic dating of those rocks interpreted, on the basis of field relationships, to be the oldest in a particular area. Supracrustal remnants are difficult

to date isotopically because most are mafic volcanic rocks. Rubidium - strontium investigations of such rocks are hampered by a small range in Rb/Sr which limits precision, susceptibility to open system behaviour, and by low Rb/Sr which results in a very slight increase in initial Sr isotopic composition for metamorphically rotated isochrons. Zircons are not a common primary mineral in mafic volcanic rocks and, if present, they may be of metamorphic origin with Zr derived from the alteration of pyroxene to amphibole.

Granitoid gneiss components of the orthogneiss are more appropriate for Rb-Sr and U-Pb zircon investigation -- they may be relatively younger than the supracrustal remnants and the tonalite gneiss but they could, if derived from the melting of such rocks, have Sr isotope initial ratios that reflect the crustal residence history of their precursor. The rock type that has received the most attention is the fine-grained tonalite gneiss component of the orthogneiss. Radiometric dating of such rocks is not without problems but they are amenable to a variety of types of Rb-Sr and U-Pb zircon investigation (Beakhouse 1983).

The available geochronological data for the orthogneiss and sodic plutons from the Winnipeg River belt suggest that these rocks are slightly to significantly older than volcanism and plutonism in the Wabigoon Subprovince (Figure 3). 'Pre-Kenoran' tonalites have been recognized at Sen Bay on eastern Lac Seul (Krogh et al. 1976a), Tannis Lake (Clark et al., 1981; D.W. Davis, personal communication, 1983), in the Kenora area (Beakhouse, 1983) and have been interpreted (from combined U-Pb zircon and Rb-Sr isotopic studies) to be present in the Cedar Lake area (Beakhouse, 1983). Rb-Sr whole rock ages tend to be younger than U-Pb zircon ages for the same units and the poorer precision of the Rb/Sr ages limits their usefulness in inter-subprovince correlation. Preliminary data from 3 other localities (Kenora, Daniels Lake, Cedar Lake) have minimum U-Pb zircon ages ranging from 2837 to 3168 Ma (Corfu et al., 1985). In addition, D.W. Davis (personal communication, 1985) has determined a minimum age for a granitoid boulder in a conglomerate near the Winnipeg River-Wabigoon interface in the Sioux Lookout area of 2897 Ma.

The U-Pb zircon geochronological data (Figure 3) suggest that although the Winnipeg River belt had a more protracted history than the Wabigoon Subprovince, there are some similarities. Most late- and post-tectonic granitic plutons in the Winnipeg River belt were emplaced between 2660 Ma and 2705 Ma with the more highly differentiated granites emplaced last. This compares favourably with the age of 2695 Ma for a post-tectonic pluton (the Taylor Lake stock) from the Wabigoon Subprovince (Davis et al. 1982). Tonalitic phases of the granitoid gneiss component of the orthogneiss from the Cedar Lake area apparently crystallized at approximately the same time as volcanic and tonalitic plutons in the Wabigoon Subprovince but have model Sr isotope initial ratios (approximately 0.704) that suggest that they originate from the melting of older crust (Beakhouse 1983). This period of melting, crystallization and associated metamorphism is broadly equivalent to a period of volcano-plutonic

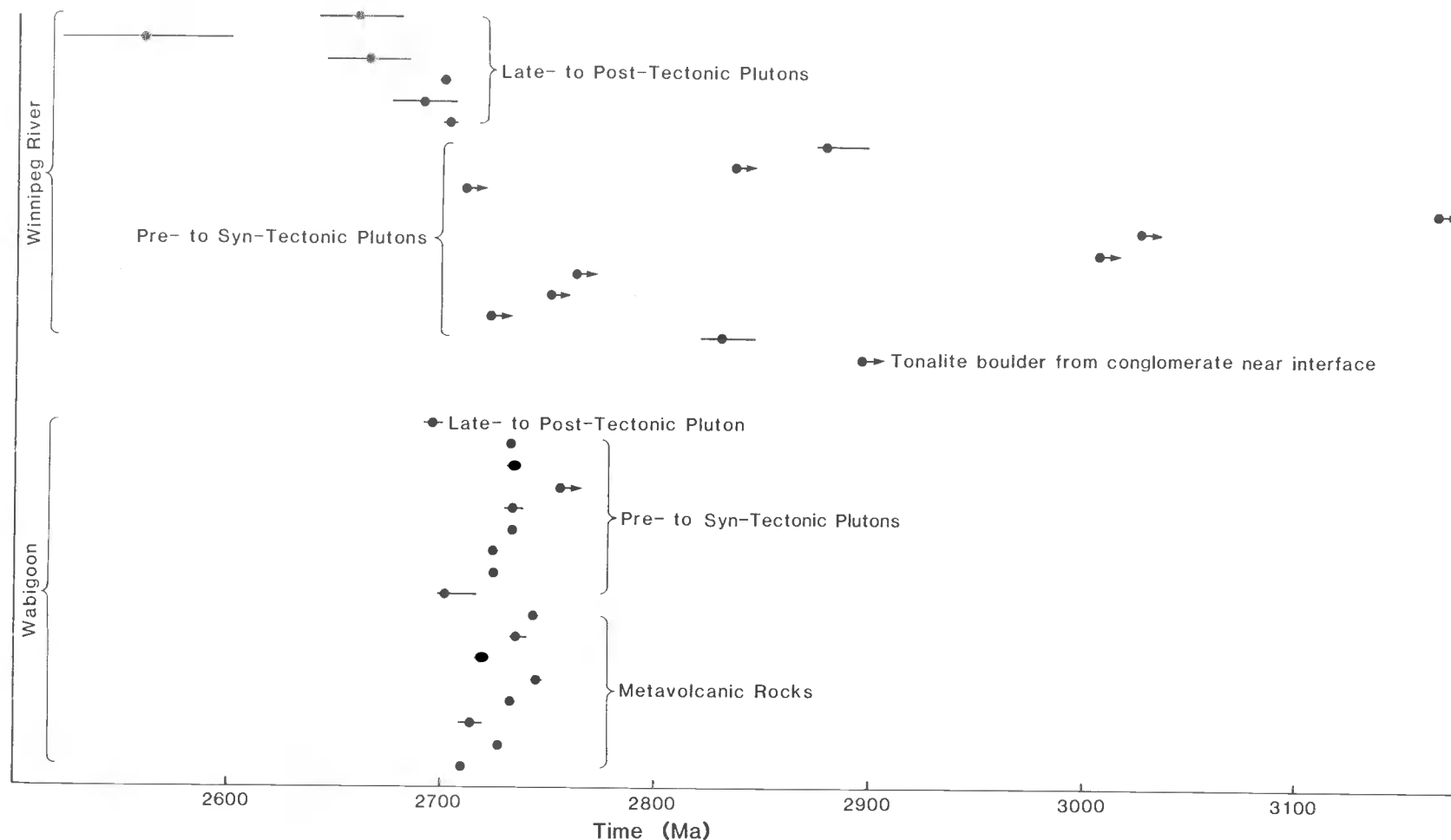


Figure 3: Summary of U-Pb zircon geochronology available for the Winnipeg River - Wabigoon subprovince area. Where error bars are not visible they are contained within the size of the symbol. Symbols with arrows indicate minimum ($^{207}\text{Pb}/^{206}\text{Pb}$) ages. Data are from Krogh et al. (1976a,b), Davis and Edwards (1982), Davis and Trowell (1982), Davis et al. (1982), Beakhouse (1983) and Corfu et al. (1985) with additional unpublished data from F. Corfu (personal communication, 1985) and D.W. Davis (personal communication, 1985).

activity in the Wabigoon Subprovince. Metamorphic zircons from a Kenora area tonalite gneiss formed somewhat earlier (approximately 2790 Ma, Beakhouse 1983) but it is not clear if this represents a discrete metamorphic event or a single, protracted metamorphic event. The crystallization age (U-Pb zircon) of this same tonalite gneiss (approximately 2830 Ma) is significantly older than the volcano-plutonic activity in the Wabigoon Subprovince but not so old as the "Lac Seul event" (Goodwin 1977).

Sodic suite plutons in the Winnipeg River belt are concentrated along the northern and southern margins of the belt and spatial association of sodic plutons and orthogneiss is especially evident along the southern margin. Breaks et al. (1978; 1981) point out that the sodic suite grades imperceptibly into orthogneiss and that the former suite "has evolved over a relatively long period of intrusion". These authors consider that one of the oldest dated units in the Winnipeg River belt, the tonalite gneiss of the Sen Bay Complex, which is more than 3008 Ma (Krogh et al. 1976a) is a sodic pluton. The Dalles granodiorite has a minimum ($^{207}\text{Pb}/^{206}\text{Pb}$) zircon age of 2762 Ma (Beakhouse 1983) that is slightly older than Wabigoon Subprovince volcanism and plutonism (Figure 3). Although there is a gradation between some sodic plutons and orthogneiss, many sodic plutons intruded orthogneiss at their present crustal levels.

In the sodic suite there is an apparent discrepancy between isotopic ages and field relationships. Identical Rb-Sr isochron ages of 2630 Ma are reported for the Dalles and Melick sodic plutons from the Kenora area (Wooden 1978). The emplacement of the Dalles granodiorite is responsible for the last major deformational event in the Kenora area (Gower 1978; Gower and Clifford 1981) and this, combined with the Rb-Sr geochronologic data, suggests that this batholith was emplaced to its present crustal level during the Kenoran orogeny. The minimum U/Pb zircon age of 2762 Ma (Beakhouse, 1983), however, indicates pre-Kenoran crystallization. Field relationships for the Melick pluton are also at variance with the apparently young Rb-Sr age and suggest that this unit was emplaced relatively early in the deformational history of the Kenora area (Gower and Clifford 1981).

Taken together, the evidence suggests that some sodic plutons are slightly to considerably older than 'Kenoran' volcano-plutonic activity in the Wabigoon Subprovince and are similar in age to the enclosing tonalite gneisses. Upward mobilization of these plutons took place during the Kenoran orogeny and is responsible for the intrusive contacts and the last major deformational event. The Dalles batholith, in particular, has many of the attributes of a classical mantled gneiss dome (Eskola 1948). Kroner et al. (1981) describe a similar relationship for tonalitic gneisses from Finland. Deformational fabrics and structures similar to those in the orthogneiss are preserved in some of the smaller plutons, as for example the Melick tonalite, but are rare in larger plutons. The absence of early deformational fabrics may reflect their obliteration during upward mobilization or lack of gneissic

development in the interior of large, relatively homogeneous batholiths that acted as buttresses against which deformation took place.

Geochemical Evidence

The geochemical characteristics of the potassic plutonic rocks of the Winnipeg River belt support the interpretation that pre-Kenoran crust was extensive. The predominant plutonic rocks of shield areas are tonalite and related rocks (quartz diorite, trondhjemite and low-K granodiorite; Glikson 1979). Burke and Kidd (1978) suggest that granites are rare in the Superior Province; a generalization that reflects, in part, the emphasis that has been placed on greenstone-tonalite terranes. Late- and post-tectonic granodioritic to granitic plutonic rocks are abundant in the Winnipeg River belt.

Typical geochemical characteristics of Archean plutonic rocks of tonalitic affinity include low K_2O , Rb, K_2O/Na_2O , Rb/Sr and Sr isotope initial ratios with moderately enriched LREE and depletion in HREE and Y (Glikson 1979). Such characteristics are explained by partial melting of a metamorphically transformed tholeiitic basalt at mantle or lower crustal depths (Hanson and Goldich 1972; Arth and Hanson 1972, 1975; Glikson 1979). In contrast, the potassic Lount Lake batholith, from the central axis of the Winnipeg River belt has relatively higher K_2O , Rb, K_2O/Na_2O , Rb/Sr and Sr isotope initial ratios, variable but generally high LREE abundances and moderate to high Y abundances (Beakhouse 1983). Other potassic plutons are even more geochemically evolved (e.g., Lac du Bonnet pluton; Cerny et al. 1981). Certain younger plutonic rocks (Dalmein, Mpagenie and Sicunusa - types) from the Kaapvaal craton of South Africa have geochemical characteristics (Condie and Hunter 1976) similar to the potassic suite plutons. These characteristics are not compatible with the origin postulated for tonalites and are more compatible with partial melting of pre-existing sialic crust.

The initial Sr isotopic composition of the Lount Lake batholith (.7022 to .7044; Beakhouse 1983) is higher than that considered 'normal' for 2.7 - 3.0 Ga tonalites (Peterman 1979). Such initial ratios are compatible with these 2.63 - 2.70 Ga plutons being derived from the partial melting of a source that is 200 to 400 Ma older having Rb/Sr ratios comparable to those of tonalitic gneisses from the orthogneiss suite (Beakhouse 1983). The production of large volumes of such magmas implies extensive pre-existing sialic crust in the Winnipeg River belt.

BOUNDARY PROBLEMS

The primary nature of the interfaces between different terranes are of fundamental importance for development of models for Archean crustal evolution. These interfaces are loci for faulting and intrusive activity because the interfaces represent regional scale discontinuities in density, ductility and rheology; the younger events obscure the primary stratigraphic relationships. The Sydney Lake Fault (Breaks et al. 1978; Stone 1977) forming the north boundary of the Ear Falls-Manigotagan belt is one of the largest of the interface fault systems. This

fault zone is up to 1 km wide and has numerous splays. Smaller fault zones are recognized along other interfaces. Stratigraphic relations at the interfaces can be deciphered only where fault and intrusive activity were relatively minor. This can be done directly only at the interfaces between the Ear Falls-Manigotagan paragneiss and the Uchi Subprovince on the north and the Bird River internal greenstone belt on the south.

The contact between the Uchi greenstone-tonalite subprovince and Ear Falls-Manigotagan paragneiss originated as a lateral facies change between fundamentally different depositional environments. The distribution of these environments was established by 2959 Ma (the age of the oldest cycle in the Uchi belt; Nunes and Thurston, 1980) and persisted for at least 200 Ma. This implies that the subprovinces of the western Superior Province are primary features representing different volcanic and sedimentary environments and are not a consequence of differential uplift.

The nature of the interface between the Winnipeg River batholithic terrane and the adjacent Ear Falls-Manigotagan paragneiss and Wabigoon greenstone-tonalite terranes cannot be determined directly. However, based on both inconclusive field data and more conclusive geochronologic data these interfaces can be interpreted to be unconformities. The Winnipeg River - Wabigoon interface, a portion of which is the subject of this field trip is discussed more fully below.

A first step in understanding the nature of interfaces is to define the properties that distinguish the subprovinces and locate the transition on a geologic map. As discussed above, the Winnipeg River belt differs from the Wabigoon subprovince in having a 1) lower proportion of supracrustal rocks, 2) generally higher metamorphic grade, 3) higher proportion of granite in relation to tonalite, 4) older age and 5) contrasting structural style. Defining the position of this transition on the ground is not a trivial problem. Between Kenora and the Manitoba-Ontario provincial boundary, where we will be examining the interface, there is a very sharp contact between older gneisses of the Winnipeg River belt and lavas and sedimentary rocks of the Wabigoon subprovince. Although a meaningful lithologic contact can be drawn on a map, the metamorphic transition (over 100-1000 m) is not so abrupt and deformation associated with the interface persists an unknown, but significant (<2 km) distance to the north and south of the contact.

Between Kenora and Vermilion Bay the location of the interface is more uncertain. Most workers have placed the sub-provincial boundary along the north edge of the Vermilion Bay greenstone belt. This position of the interface is based largely on the observations that 1) no volcanic remnants within the Winnipeg River belt are comparable to the Vermilion Bay greenstone belt in terms of size and degree of preservation and 2) the lobate form of the Dryberry batholith to the south of the Vermilion belt is typical of that observed for batholiths within the Wabigoon subprovince. Two observations suggest that the Vermilion Bay belt and Dryberry batholith could be placed within the Winnipeg River belt. Firstly, the Dryberry batholith is

similar to the potassic plutons of the Winnipeg River belt, and unlike other Wabigoon subprovince batholiths, in terms of aeromagnetic signature, magnetic susceptibility and perhaps lithology (Hall, 1968; Beakhouse, 1977). Secondly, an amphibolitic "tail" extending off the west end of the Vermilion Bay greenstone belt is folded along with Winnipeg River belt gneisses into a large recumbent structure in the Silver Lake area (Beakhouse et al. 1983).

The position of the subprovince interface between Vermilion Bay and Sioux Lookout is controversial. Breaks et al. (1978) locate the subprovincial boundary at the sedimentary-volcanic contact resulting in a highly irregular contact. Although some of these metasediments are highly metamorphosed and partially melted they can be correlated with the Warclub Group which can be traced well into the Lake of the Woods greenstone belt (Trowell et al., 1980). The proposed boundary must therefore, at some point (not illustrated on the map of Breaks et al. 1978) cross stratigraphy. An alternate interpretation is that the subprovincial boundary lies between these supracrustal rocks and the granitoid rocks to the north. Unfortunately, this contact is poorly exposed.

The significance of the discussion above is not simply to point out that further investigations are required to elucidate the nature and position of the subprovincial boundary. Rather, it serves to emphasize that subprovince interfaces must be regarded as imprecisely bounded zones rather than a line on a map. Within these interface zones processes characteristic of either subprovince may operate in addition to processes uniquely related to the interface (eg. faulting and associated, tectonically controlled, epiclastic sedimentation).

Assuming the arguments presented above are correct and the Winnipeg River belt represents an extensive tract of pre-Kenoran sialic crust and the deposition of supracrustal sequences in the Wabigoon subprovince post-date the formation of this sialic crust, there are two alternate interpretations for the nature of the interface. One alternative is that the interface is a tectonic contact and the subprovinces were juxtaposed after their formation. Alternatively, the Wabigoon supracrustal sequences could have been deposited upon, or more likely against, this sialic crust in which case a major unconformity occurs within the interface zone. The latter interpretation is supported by the possible presence of feeder dikes in the Tannis Lake area and, more convincingly, by the presence of an ancient boulder in a conglomerate near Sioux Lookout (D.W. Davis, personal communication, 1985). Clearly, intense deformation has been superimposed on this unconformity and minor or major dislocations have occurred along it. It is also possible that the Winnipeg River terrane could have been juxtaposed (by the closing of an ocean??) against the Wabigoon subprovince during volcanism and sedimentation and that the unconformable relationship applies only to the youngest sequences in the Wabigoon.

SUMMARY OF GEOLOGICAL RELATIONSHIPS

Field, geochronological and geochemical evidence suggest a

complex geologic history spanning at least 500 Ma. The earliest event is deposition of a supracrustal sequence preserved as inclusions in younger plutonic rocks. The rock types in this association are broadly similar to those in the better preserved, younger greenstone belts, although felsic metavolcanic units have not been identified conclusively.

Tonalite gneiss of the orthogneiss suite and certain of the sodic plutons are older than 'Kenoran' volcanism in the Wabigoon Subprovince. More data are required to establish whether these tonalites were all formed 1) during a 3.0 - 3.1 Ga event (Goodwin 1977) or 2) semi-continuously from 3.1 Ga until the Kenoran orogeny. The antiquity of some tonalites in the Winnipeg River belt suggests deposition of the oldest supracrustal rocks took place prior to 3.0 Ga. The tonalitic gneiss and sodic plutons are geochemically similar to rocks of tonalitic affinity that occur in bi-modal association with basalts of the younger greenstone-tonalite terranes and originate from the partial melting metamorphosed tholeiitic basalt.

The precursor to the paragneiss in the Ear Falls-Manigotagan belt was deposited synchronously with volcano-sedimentary sequences in the adjacent greenstone-tonalite terranes. The time interval over which this sedimentation occurred is not resolved. Volcanism in the Uchi-Confederation Lakes area commenced by 2959 Ma and continued for 220 Ma (Nunes and Thurston 1980). Sedimentation occurred synchronously with the oldest volcanic cycle (Thurston and Breakey 1978) and probably continued during deposition of the younger volcanic cycles (P.C. Thurston 1982, personal communication). The detritus for the paragneiss precursor was apparently derived externally from flanking greenstone terranes and possibly from a sialic microcontinent, now represented by the Winnipeg River belt; they were deposited from turbidity currents in relatively deep water. The provenance included felsic volcanics which are locally or regionally predominant, felsic plutonic, mafic volcanic and pre-existing sedimentary units. Crustal structure perturbations (see references in Beakhouse 1977) are a manifestation of the tectonic regime associated with basin development.

The major thermotectonic event between 2660 and 2760 Ma, that culminated with the Kenoran orogeny, resulted in the generation of large volumes of felsic plutonic rocks in all sub-provinces of the western Superior Province. Plutonic rocks of the greenstone subprovinces are dominantly tonalitic in composition and probably originated from the partial melting of mantle derived basalts that were only slightly older than the plutons (Hanson and Goldich, 1972; Arth and Hanson, 1972, 1975; Glikson, 1979). In contrast, geochemical evidence indicates that the extensive potassic plutonic rocks in the Winnipeg River belt could not have originated from the partial melting of rocks derived from the mantle during the Kenoran orogeny. Instead they apparently originated from the partial melting or upward mobilization of a "pre-Kenoran" volcano-plutonic, greenstone-type, association.

IMPLICATIONS FOR CRUSTAL EVOLUTION

The recognition of a geological history spanning 500 Ma poses certain constraints, and many questions, for crustal evolution in the western Superior Province. The limited, reliable geochronological data suggests semi-continuous activity during this 500 Ma interval rather than two "accretion super-events" (Goodwin, 1977).

Sialic crust of unknown extent was present prior to the deposition of 'Kenoran' supracrustal sequences but the role of this crust in subsequent crustal evolution is unclear. The pre-Kenoran sialic crust has been reworked and intruded and dismembered by extensive Kenoran felsic plutons derived from the partial melting of greenstone-tonalite assemblages of the earlier crust. These older volcanic rocks may be analogous to the better preserved, pre-Kenoran volcanic cycles now recognized in greenstone belts elsewhere in the Superior Province (Nunes and Wood 1980; Nunes and Thurston 1980). The paucity of mantle derived Kenoran volcanic and plutonic rocks in the central part of the WRB suggests that these rocks were not emplaced through the older crust in this region. The presence of mafic dikes near the southern margin of the WRB, possibly representing feeders to Kenoran volcanism, and the occurrence of an ancient granitoid boulder in a conglomerate near Sioux Lookout implies that the younger supracrustal sequences may be, in part, ensialic but the present juxtaposition of relatively older and younger sequences may also be tectonic. The paucity of granitoid rocks originating from the anatexis of sialic crust in the Wabigoon subprovince suggests that a Winnipeg River belt-type basement complex is absent or greatly thinned beneath this area.

The recognition of lateral facies relationships between greenstone and paragneiss sequences establishes the contemporaneity of these terranes and implies that the belt-like, sub-province structure of the western Superior Province is a primary characteristic and not a consequence of differential uplift. The pattern of pre-existing sialic nuclei and younger, coeval, interdigitating volcanic arcs and sedimentary basins is geometrically, though not necessarily genetically, analogous to modern plate tectonic regimes.

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ROAD LOG AND DESCRIPTION OF STOPS

- 0.0 km We will proceed west on highway 17 from Husky the Muskie in the town of Kenora. En route to the first stop, highway 17 is parallel to, and lies 1 to 2 kilometres south of, the Wabigoon - Winnipeg River interface (Figure 4). Moderately to intensely flattened and lineated metavolcanic rocks and granitoid rocks (Winnetka Lake stock) are exposed in road cuts along this route.
- 2.5 km Junction with highway 596
- 10.0 km Junction with highway 641
- 11.2 km Junction with Kenora bypass
- 23.7 km Junction with Pye's Road. For the next 9.1 km we will traverse the elliptical Winnetka Lake stock that lies along the Wabigoon - Winnipeg River interface. This stock is compositionally zoned with a granite to granodiorite core and a hornblende diorite rim (Davies, 1965) and its elliptical geometry is, in part, a consequence of post-crystallization deformation (Brisbin, 1981).
- 32.8 km The contact of the Winnetka Lake stock with enveloping Wabigoon subprovince metavolcanic rocks is exposed in low outcrops on the south side of the highway.
- 35.1 km Junction with the Rush Bay Road
- 41.3 km Junction with the Royal Lake Road. The first stop is on the north side of the highway just beyond this junction.
- The lithologic diversity of the Keewatin series metavolcanic and metasedimentary rocks cannot be adequately represented on this field trip. Stops 1A and 1B are included to illustrate some of the factors which complicate interpretation of these rocks.
- 41.35 km STOP 1A Keewatin series Pillow Lava
Outcrop on the north side of the highway below the sign for the Royal Lake Resort.
In this outcrop we will examine a fine grained, buff to light green weathering, feldspar-phyric pillowed flow. In thin section the rock consists of plagioclase phenocrysts up to 2 mm in longest dimension in a finer grained matrix consisting of feldspar, quartz, actinolite, biotite and epidote. It is difficult to estimate proportions of quartz and feldspar but veins and diffuse patches of quartz suggest the rock has been silicified.

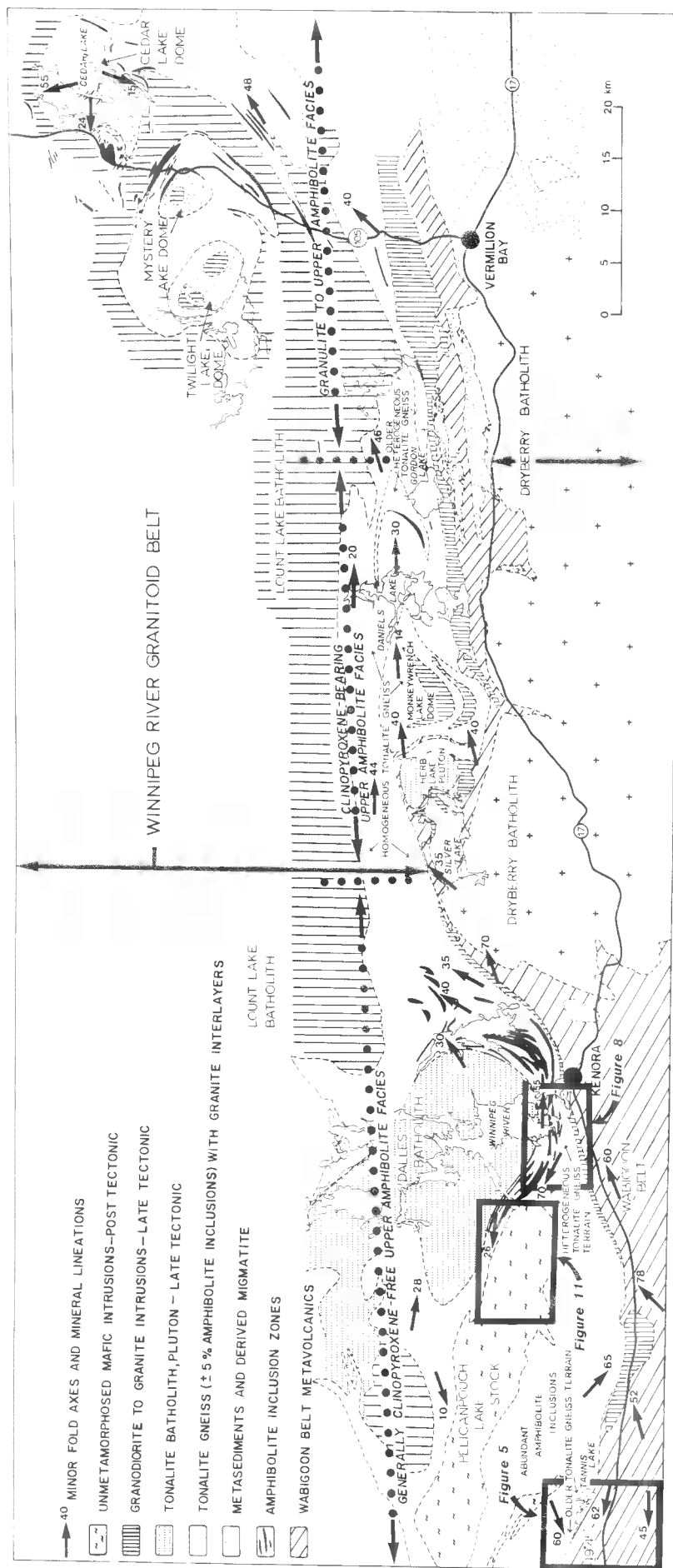
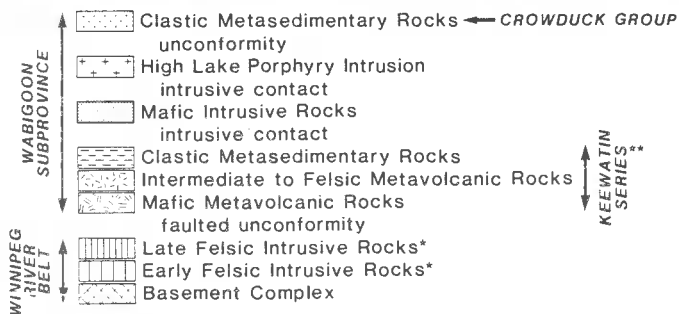
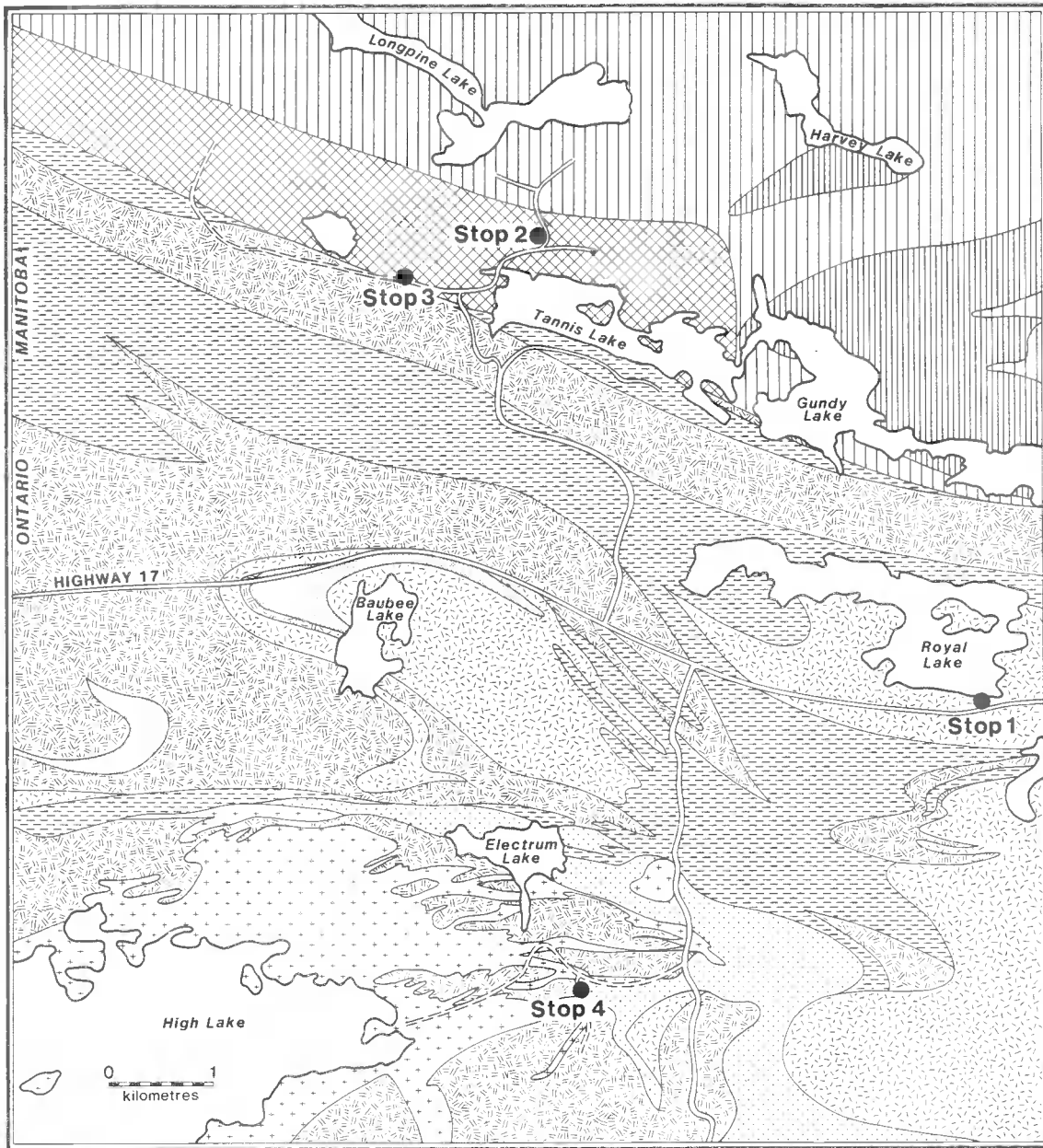


Figure 4: Geologic map of the Wabigoon-Winnipeg River interface between the Manitoba-Ontario border and Vermilion Bay (from Beakhouse et al., 1983). The location of figures 5, 8 and 11 are outlined.



*The relative age of felsic intrusive rocks that intrude the basement complex with respect to the older unconformity and Wabigoon subprovince lithologies is not known.

**Rocks in the Keewatin Series are subdivided lithologically and the order does not imply age relationships.

Figure 5: Geologic map of the Tannis Lake-High Lake area generalized from Davies (1965) with additional information on the distribution of the basement complex from Bald (1981). The location of this map is illustrated in figure 4.

A chemical analysis of this rock (sample 84GPB8010, appendix) is difficult to interpret due to this alteration. Elements whose abundance may not be profoundly affected by this alteration suggest that the rock was geochemically evolved and is not simply a silicified basalt. Relative to tholeiitic basalts from the area (samples 83GPB323, 84GPB8007 and 84GPB8009) this rock has lower Co, Cr, Ni and Ti and higher Zr and LREE. The fractionated REE pattern ($La_n/Yb_n = 7$) is also distinctive. These characteristics are similar to those of certain basaltic andesites from the Confederation Lake belt (eq. samples 439 and 448; Thurston and Fryer, 1983) having SiO_2 contents of approximately 55 percent. The relative significance of primary abundances and geochemical mobility accompanying secondary alteration in determining the observed compositions and apparent calc-alkaline affinity of this rock is poorly constrained. This poses a problem in comparing the chemistry of mafic and intermediate dikes that cut an interpreted basement complex to the north with that of the flows which they may be feeding.

- 41.8 km STOP 1B Keewatin Series Metasedimentary Rocks
 Outcrop on the north side of the highway
 Thin to medium bedded metasedimentary rocks. Metasandstones predominate but metasiltstone and meta-argillite (the latter often disrupted) are common in parts of the outcrop. Graded bedding is recognizable locally. Although these rocks were deposited in a sedimentary environment they are probably closely related to felsic volcanic activity.
 Throughout much of the outcrop, foliation (mineral foliation and flattening of fragments) is at a high angle to bedding, mimicking the regional scale structure over much of the area (Davies, 1965). In intensely deformed zones, bedding is transposed so as to be essentially parallel to the foliation. The foliation is parallel to the axial plane of a megascopic fold at the crest of the roadcut.
- 43.35 km Junction with Shoal Lake Road
- 44.1 km Junction with Gundy Lake Road - turn north on Gundy Lake Road - reset odometer to 0.0.
- 3.1 km South Shore Road junction - turn left
- 3.5 km Powerline
- 4.05 km Piute Lake Road junction - turn right
- 4.6 km Causeway at end of Tannis Lake. The Wabigoon - Winnipeg River interface lies along the south shore of

Tannis Lake (Figure 5). The outcrop on the north side of the causeway (west side of road) consists of foliated tonalite cut by mafic dikes similar to those at the next stop

5.05 km Junction - turn left

5.15 km STOP 2 Tonalite with Mafic Dikes.

Outcrop on the west side of the road.

The principal rock type in the outcrop is a fine to medium grained, grey, strongly lineated biotite tonalite. Clark et al (1981) determined a Rb-Sr whole rock isochron age, based on regional sample selection, of 2950 ± 150 Ma ($R_i = .7028 \pm .0014$). Recent U-Pb zircon geochronology on a sample selected from this outcrop confirms the antiquity of this unit and indicates a minimum ($^{207}\text{Pb}/^{206}\text{Pb}$) age of 3028 Ma (D.W. Davis, 1984 personal communication).

The tonalite here is relatively "clean" and contains only a few inclusions of massive amphibolite of indeterminate origin. These amphibolite inclusions are more discontinuous than the mafic dikes cutting the outcrop and have subtle differences in composition and texture but the possibility that they represent disrupted dikes cannot be totally discounted on this outcrop. Elsewhere, amphibolite inclusions with calc-silicate pods and highly stretched pillows demonstrate a volcanic origin for some of the inclusions and provide evidence for pre-3.0 Ga volcanism in the Winnipeg River belt.

The tonalite is cut by deformed, equigranular and porphyritic (plagioclase phenocrysts), amphibolitic mafic dikes. The age of the mafic dikes is not well constrained except that they are younger than the tonalite (3.028 Ga) and older than the last deformation and certain of the variously deformed granitic pegmatite dikes. The age of these intrusive phases and deformation is not constrained here but, by analogy with other areas in the Winnipeg River belt, the dikes are thought to be older than 2.7 Ga. The age constraints and proximity to the Wabigoon subprovince metavolcanic rocks here led to the suggestion that the dikes cut an older basement complex and fed Keewatin series volcanism (Bald, 1981; Clark et al, 1981). Modal and chemical analyses of the tonalite (sample 321A) and equigranular mafic dike (sample 321B) are presented in the appendix.

Proceed back towards the highway for 1.05 km to the junction with the Piute Lake Road - turn right (west) - proceed 0.55 km.

STOP 3 Wabigoon - Winnipeg River Contact

The Wabigoon-Winnipeg River contact in this area corresponds to a linear topographic depression and

exposures of the contact are not abundant. The outcrops on the road immediately before we stop and most of those to the north of the road are fine to medium grained, foliated to gneissic, dominantly tonalitic rocks cut by mafic dikes (Figure 6). The outcrops to the south of the road consist of fine to medium grained, amphibolitic mafic volcanic rocks with interbedded chert-magnetite iron formation. Some highly deformed pillow structures are present on the largest outcrop. The medium grained mafic rocks may represent sills or transposed dikes. A thin tuffaceous horizon occurs near the north side of the thicker iron formation unit.

The mafic dikes cutting the larger tonalite outcrops north of the road are atypical in terms of their composition and heterogeneity. The banding and calcic pods within them resembles those associated with metavolcanic amphibolite inclusions in the Winnipeg River belt. These mafic units are subparallel to the foliation in the tonalite but at one contact (Figure 6) a low angle discordance is noted. The relationships developed in the last stop description and context in which these unusual dikes occur affords a highly interpretive though plausible explanation. It is conceivable that these dikes are feeding flows within a few tens of metres (in the deformed state) of where we are observing them. It is reasonable to expect that the processes (autobrecciation, interaction with sea water) that produce "volcanic" structures and alteration in flows would produce similar features in the upper portions of venting dikes.

The actual contact between the Winnipeg River belt and Wabigoon subprovince is not exposed here. Immediately north of the road is a small outcrop of highly sheared mafic rock cut by at least one deformed mafic dike. It is not known if the sheared mafic rock is a flow or dike. The highly sheared nature of this rock and the negative topographic expression of the contact suggests the presence of a fault. Does this fault represent a major dislocation that juxtaposes two unrelated assemblages, at least one of which is allochthonous or does it represent a minor (or major) adjustment along an original unconformity?

Return to Highway 17 - turn left (east) - proceed 0.75 km to the Shoal Lake Road - proceed south on the Shoal Lake Road for 2.9 km - turn right (west) onto narrow road - proceed 0.8 km west on the narrow road - the exposures at stop 4 are located approximately 30 metres south of the road.

STOP 4 High Lake Unconformity

In this outcrop, metabasalts of the Keewatin series are cut by porphyry dikes related to the High Lake intrusion and both are unconformably overlain by the

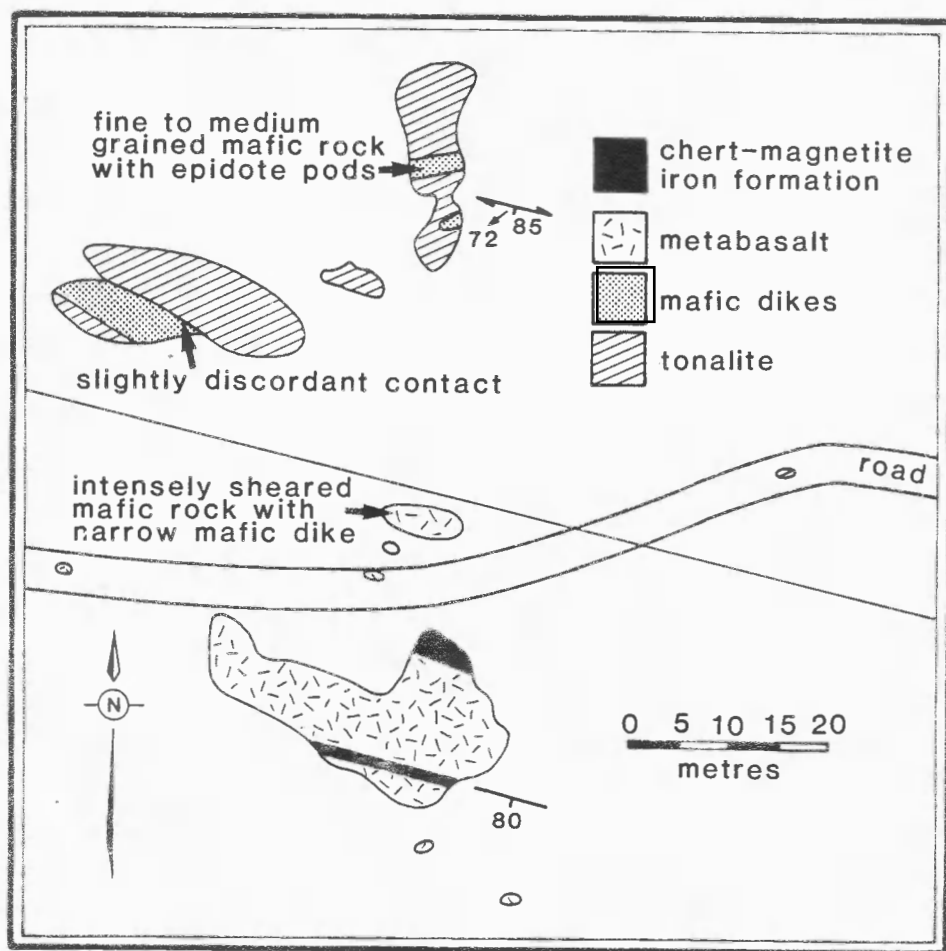


Figure 6: Generalized outcrop map of stop 3.

metasedimentary rocks of the Crowduck group (Figure 7). The metabasalts (most of which are pillowed) and High Lake porphyry dike are exposed on the western edge of the outcrop.

In this outcrop, the Crowduck group is divisible into four units that reflect a general coarsening upwards trend (Figure 7) Stratigraphic relationships are complex, however, because many units were eroded prior to or concomittant with deposition of overlying units. Such scouring relationships are well displayed immediately east of the northernmost exposure of the porphyry dike. Near this same locality, the effect of local provenance in the conglomeratic facies is illustrated by a lense of conglomerate richer in mafic volcanic clasts than the remainder of the outcrop.

Throughout most of the conglomeratic facies in this outcrop, clasts of the porphyry predominate and are conspicuously larger than the other clast types which include relatively abundant mafic to felsic volcanic rocks and minor chert, iron formation and massive sulphides. The thin to medium bedded sandstone-siltstone sequence exposed near the unconformity in the eastern part of the large outcrop display spectacular soft sediment deformation structures including slump folds, load casts and flame structures.

A regolith is developed beneath the Crowduck group and is most apparent in the porphyry unit. It is not clear how much of the irregular nature of the unconformity surface can be attributed to pre-Crowduck group topography. The presence of clasts with exposed cross-sectional areas in excess of one square metre clearly requires at least local steep gradients to have been present. The geometry of the unconformity mimics the pattern displayed by the regional folding with east-west axial planes that deforms the Crowduck group. The orientation of foliation in this outcrop is approximately parallel to these axial planes and is markedly discordant to bedding in the Crowduck group. These observations suggest that at least part of the irregularity of the unconformity is a consequence of post-Crowduck deformation. In the southwestern part of the outcrop, thinly bedded sandstones and siltstones of the Crowduck group overly the regolith above the porphyry dike and butt up against the adjacent Keewatin series metabasalt. Similar units immediately overlying the metabasalt are several metres stratigraphically above the base of the sequence over the porphyry unit. Does this represent pre-Crowduck topography or progressive syn-depositional faulting?

Quartz and quartz-tourmaline veins cut the conglomerate and in the vicinity of the High Lake stock are associated with gold mineralization (Davies, 1965). Locally, the veins appear to be confined to porphyry clasts suggesting pre-erosional veining but when

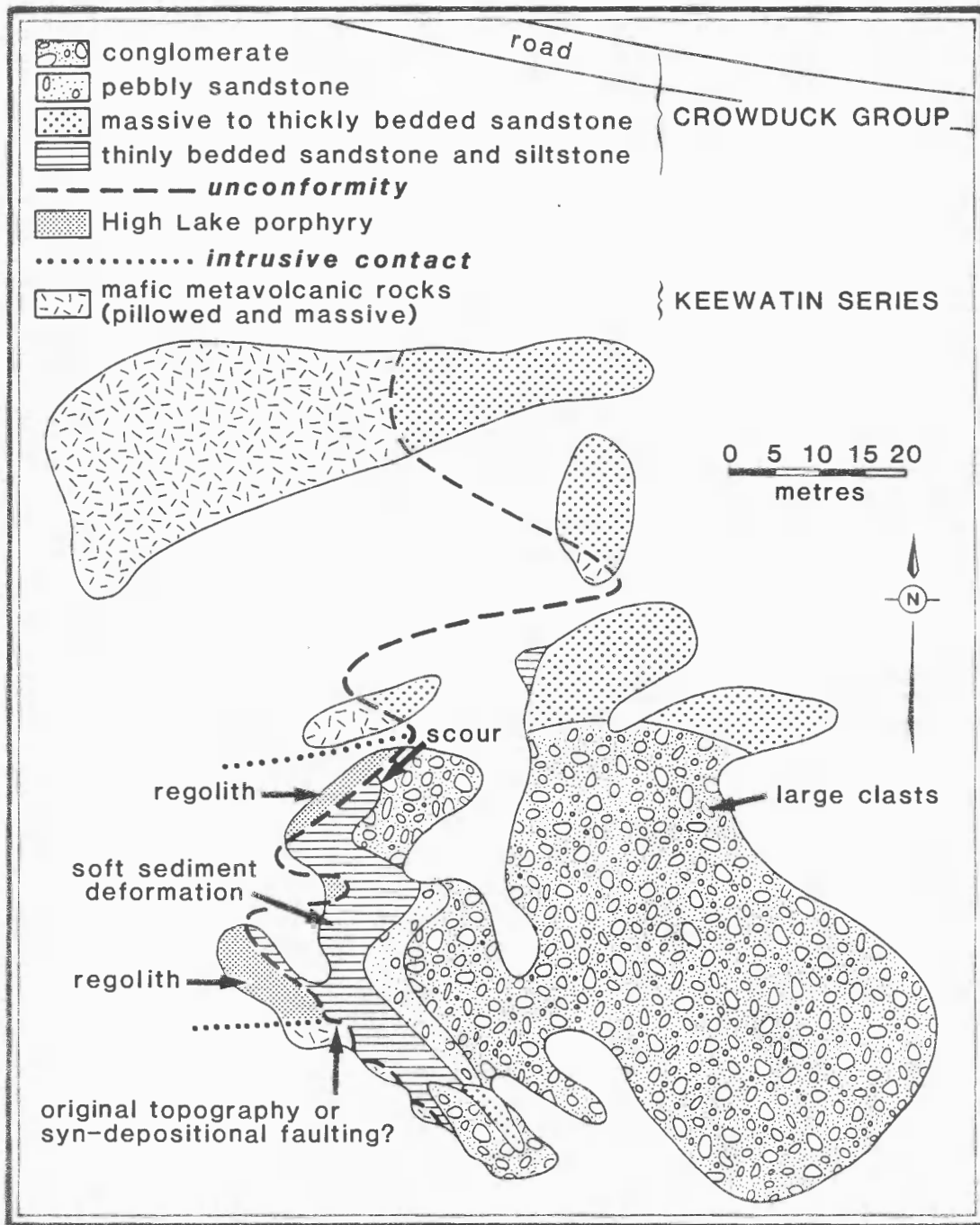


Figure 7: Generalized outcrop map of stop 4.

examined in detail most of these have 'tails' that project into the matrix. Some veins are continuous and clearly post-depositional. The concentration of this veining in porphyry clasts is interpreted to be due to heterogeneous deformation arising from the differing mechanical properties of the various types of clasts and matrix.

The Crowduck group has no obvious, direct relationship to the Wabigoon-Winnipeg River interface. However, when viewed from a regional perspective, many Temiskaming-type supracrustal sequences do appear to be spatially associated with subprovince interfaces. Other examples of similar sequences occurring near an interface include those at Rainy Lake, Beardmore-Geraldton and Kirkland Lake. This observation may reflect the concentration of later tectonic activity and associated tectonically controlled sedimentation at subprovince interfaces.

Return to highway and proceed east to the junction of highway 17 and 596 in the town of Norman (between Keewatin and Kenora)- set odometer to zero - turn north on 596.

1.5 km Junction - turn right

1.6 km STOP 5 Marginal Granodiorite Unit

We will stop near the crest of the hill and walk north on Highway 596 to the shores of Darlington Bay. We will examine a series of outcrops which provide a cross-section of the Marginal Granodiorite. This unit is continuous along the contact between metavolcanic rocks of the Wabigoon subprovince and tonalitic gneisses of the Winnipeg River belt for at least 30 km (Figure 8). Along this length its width ranges from less than 100 metres to approximately 700 metres but 200 metres is typical. The width of the unit here is approximately 350 metres and displays relationships typical of the unit along its length.

In the first outcrop at the top of the hill (5A; Figure 9) we can observe the sharp contact between the Keewatin metavolcanics and Marginal Granodiorite. The most striking textural characteristic of the granodiorite is the presence of subhedral to lenticular, poikilitic microcline megacrysts. They have some characteristics of both phenocrysts and porphyroblasts but are interpreted to be late magmatic megacrysts. Locally, evidence for some late blastic growth or overgrowth of microcline is evidenced by the coexistence of augen and subhedral megacrysts. The homogeneity of the granodiorite in this outcrop is typical of that displayed along strike in the southern one third of the unit adjacent to the Keewatin metavolcanic rocks.

As we proceed down the hill we will briefly examine a

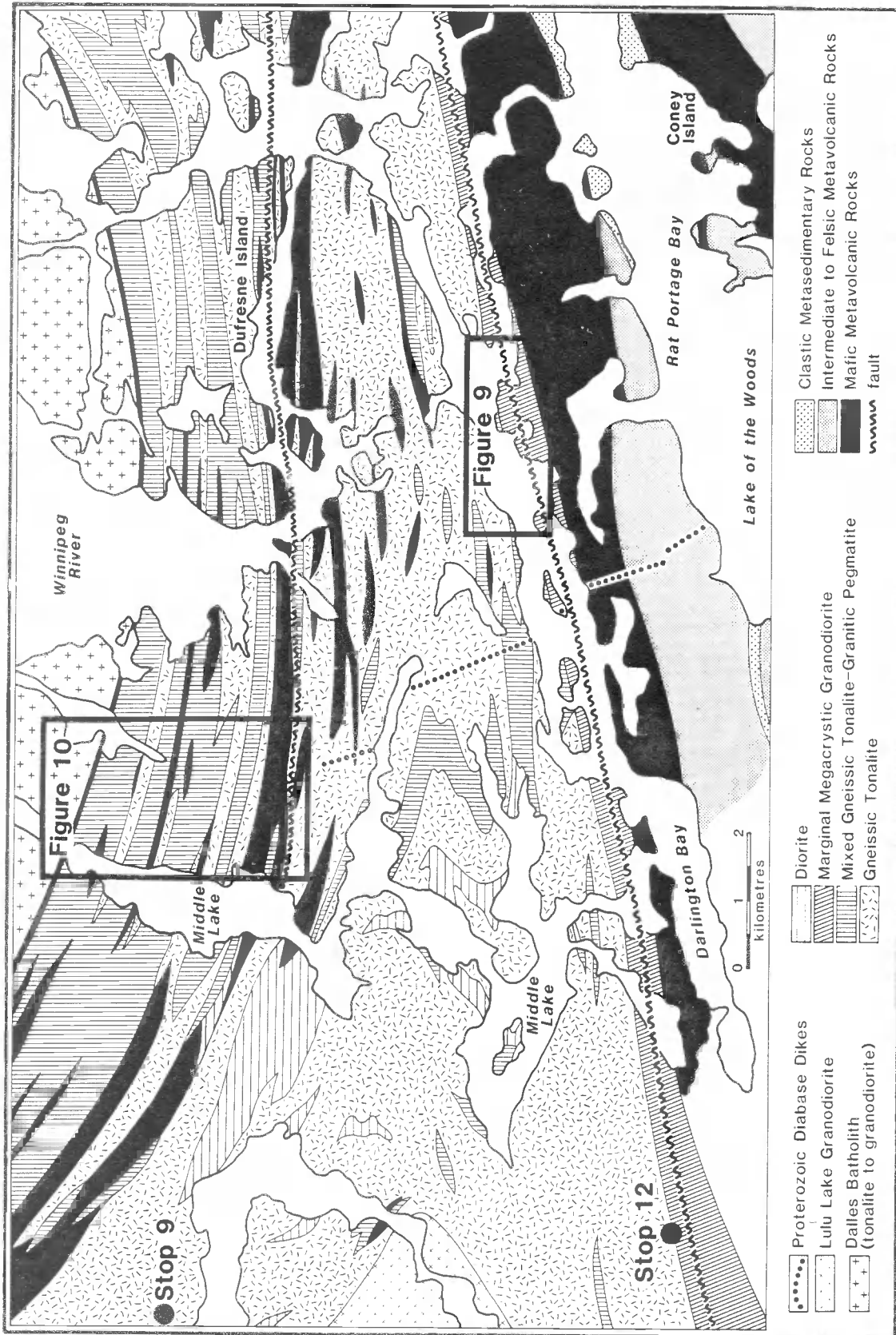


Figure 8: Geologic map of the Keewatin area (simplified from Gower (1978) with minor revisions). The location of this map is illustrated in figure 4.

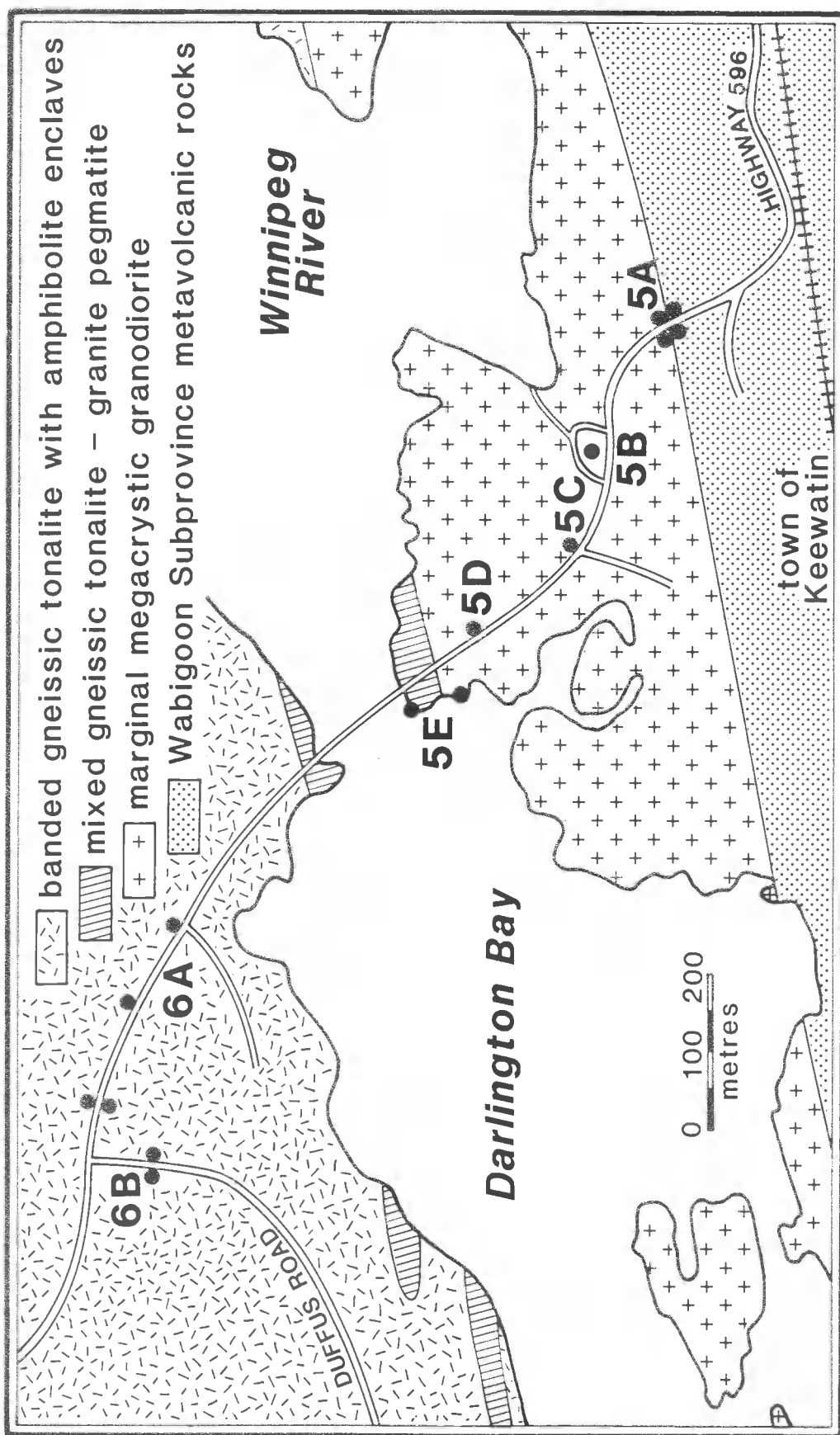


Figure 9: Geologic map of the Darlington Bay bridge area illustrating the locations of stops 5 and 6. Refer to figure 8 for the location of this map.

series of outcrops (5B, C and D, Figure 9) which illustrate the changes across the unit. Note the variable intensity of fabric development, local development of a banded character (best seen in outcrops 5B and 5D) and increasing abundance of amphibolite inclusions and secondary microcline seams and megacrysts. Depending on water level, one can see the transition from Marginal Granodiorite to granitic pegmatite-tonalite gneiss complex in the outcrops on the shore of Darlington Bay (outcrop 5E, Figure 9). We will examine a better exposure of this contact later in the day.

A modal and chemical analysis of a sample from outcrop 5D (sample 313) is presented in the appendix. Zircons separated from this sample yield a preliminary U-Pb zircon age of $2705 \pm 9/7$ Ma (Corfu et al., 1985). The unusual geometry of this unit together with its regional context, suggest it was emplaced into a zone of weakness that separated two contrasting terranes and that subsequent to the emplacement of the Marginal granodiorite, this contact was tectonically active. The nature of the pre-emplacement zone of weakness is not clear and the granodiorite precludes direct observation of the contact but it could have been either an unconformity or fault. The principal evidence for an unconformable relationship is geochronological: the tonalites north of the contact are 2.83-2.88 Ga (Beakhouse, 1983; Corfu et al., 1985) and are significantly older than volcanism in the Wabigoon subprovince, although metavolcanic rocks in the vicinity of the interface have not themselves been dated. Tholeiitic mafic dikes such as at Tannis Lake are not recognized in the Keewatin area. Some deformed narrow mafic dikes cut both the gneisses and granodiorite but these are not voluminous and, as noted by Gower (1978), are of calc-alkaline affinity. Modal and chemical analyses of two such dikes (samples 270 and 272) from the Darlington Bay area are presented in the appendix.

Proceed across Darlington Bay bridge

Reset odometer to 0.0 km at the north end of bridge

0.1 km

STOP 6 Banded Tonalite Gneiss

From here we will walk along the roadcuts on the east side of highway 596 up to Duffus Road and examine some exposures on this road immediately west of the highway (Figure 9). The distinctive banding within these rocks is interpreted to be a consequence of the intense deformation of a heterolithic plutonic complex consisting primarily of:

- fine to medium grained, biotite $\pm$ hornblende tonalite to leuco-tonalite
- fine grained amphibolite with associated diorite-

quartz diorite

- variously deformed granitic pegmatite dikes
- K- feldspar seams and megacrysts

Note the tight folding and shallow angle discordancies in gneissosity. Cross-cutting relationships between the various tonalite phases are present but are not abundant, possibly because pre-existing anisotropy in the rock has controlled the emplacement of subsequent intrusive phases. Discordant contacts are difficult to recognize due to intense transposition into the plane of the gneissosity.

3.4 km

STOP 7 Metavolcanic Amphibolite

The outcrop is on the west side of the road approximately 100 metres south of the junction between Highway 596 and the Kenora bypass (Figures 8 and 10). The principal rock type in this outcrop is a fine grained, equigranular, massive to banded amphibolite. In lakeshore exposures 1 to 2 km east of here on the Winnipeg River system, such banding can be clearly shown to represent highly flattened pillow structures. Also diagnostic of a volcanic origin are the presence of calc-silicate pods and bands that are interpreted to be metamorphosed seawater alteration phenomena. With the exception of the example discussed at stop 3, such calcium-rich mineral assemblages have not been noted within amphibolites for which a mafic dike origin can be demonstrated.

Concordant sheets of medium grained biotite tonalite constitute 25 percent of the outcrop. A narrow, fine grained diabase dike, interpreted to be of Middle to Late Precambrian age, crosscuts all other lithologies in the outcrop.

3.5 km

Kenora bypass

4.35 km

Amphibolitic metabasalt with pegmatite dikes (alternate stop)

4.9 km

STOP 8A. Pegmatite - Tonalite Gneiss Complex

We will stop at a road into a gravel pit on the north side of the highway.

Outcrops on the south side of the road consist of tonalite gneiss with minor amphibolite enclaves and abundant pegmatoid phases. The deformed pegmatoid rocks occur as concordant to sub-concordant sheets that range in composition from granite to alkali granite to microcline $\pm$ minor quartz. The latter composition is not that of a magmatic rock and it is likely that some or all of these pegmatoid phases are generated by vapour phase transfer and metasomatism rather than crystallized from silicate melts. Some phases, however, do have granitic minimum melt composition and it is not unreasonable to expect that the presence of a

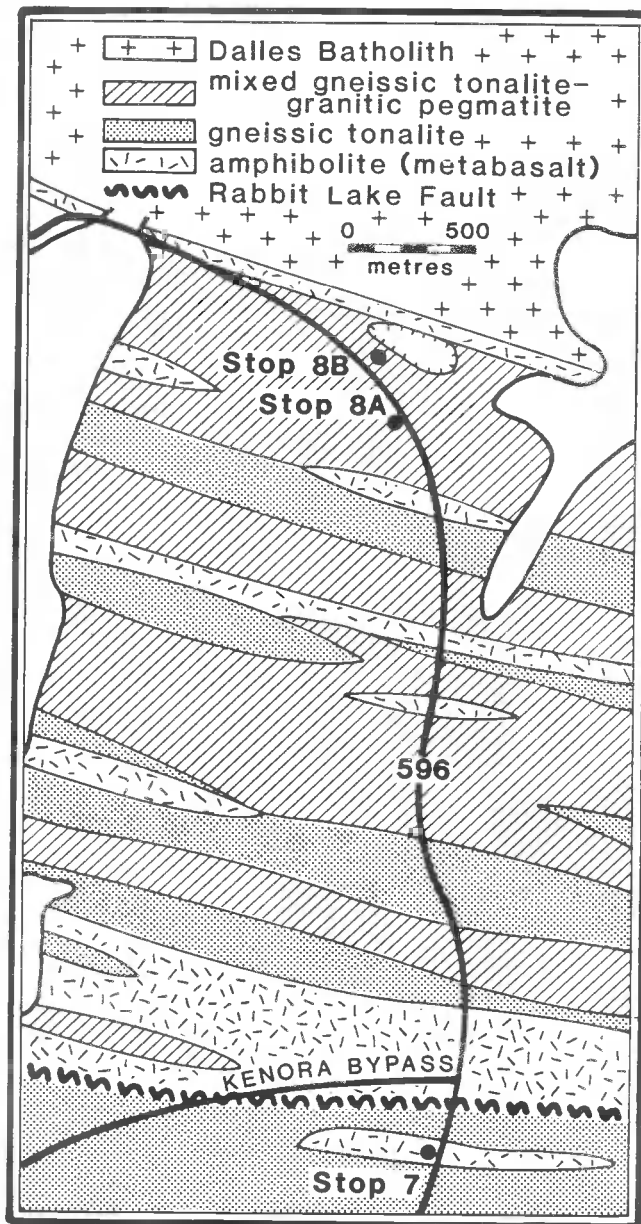


Figure 10: Geologic map of an area along Highway 596 between the Rabbit Lake fault and the Dalles batholith with the locations of stops 7 and 8. Refer to figure 8 for the location of this map.

vapor phase could initiate partial melting of the tonalite. The geometry of the microcline rich phase ranges from continuous planar sheets to pinch and swell sheets to isolated microcline megacrysts. A number of undeformed granitic pegmatite dikes which exhibit compositional and textural zonation are present in the outcrop.

STOP 8B. Standing on the south lip (but not too close to the edge - there are dangerous overhangs) of the gravel pit approximately 100 metres west of the turn-off, one can see on the opposite side of the pit exposures of the Dalles batholith (furthest away), in contact with an amphibolitic metabasalt unit (Figure 10). Near the bottom of the pit the amphibolite unit is in contact with pegmatite-tonalite gneiss such as seen in the outcrops discussed above. This thin (approximately 100 metres) amphibolite unit is laterally persistent for approximately 15 km along the Dalles - gneiss contact.

A similar amphibolite unit occurs around the Herb Lake pluton 38 km to the east-northeast of here. The Herb Lake pluton has many of the characteristics of the Dalles batholith and the gneisses and amphibolites around this unit are deformed into a large, recumbent, nappe-like structure domed upwards by the Herb Lake pluton (Beakhouse et al., 1983). One sample from the root zone of this complex in the Daniels Lake area has a similar age (2837 Ma, Corfu et al., 1985) to that of the banded tonalite gneisses in the Kenora area. This raises the possibility that the 2.83-2.85 Ga gneisses in the Kenora area may be in large part allochthonous.

- 9.8 km STOP 9 (optional) Diorite Intrusion Breccia - Psuedo-conglomerate
 Outcrop on the north side of the highway
 This intrusion breccia contains abundant fragments representing a wide range of rock types and resembles a conglomerate. Mafic to ultramafic fragments predominate. Enclaves of similar lithologies are common in dioritic phases of the Trout Lake pluton to the west and may be cognate inclusions. Intermediate to felsic fragments are also present and probably represent accidental inclusions. The matrix of the breccia is a medium grained, equigranular, hornblende diorite.
- 10.8 km Sandy Lake
- 12.0 km Junction - continue straight ahead on Highway 596
- 15.95 km STOP 10 (optional) Megacrystic Granodiorite
 We will examine low roadcuts occurring on both sides of the road.
 The stops made to this point are not truly

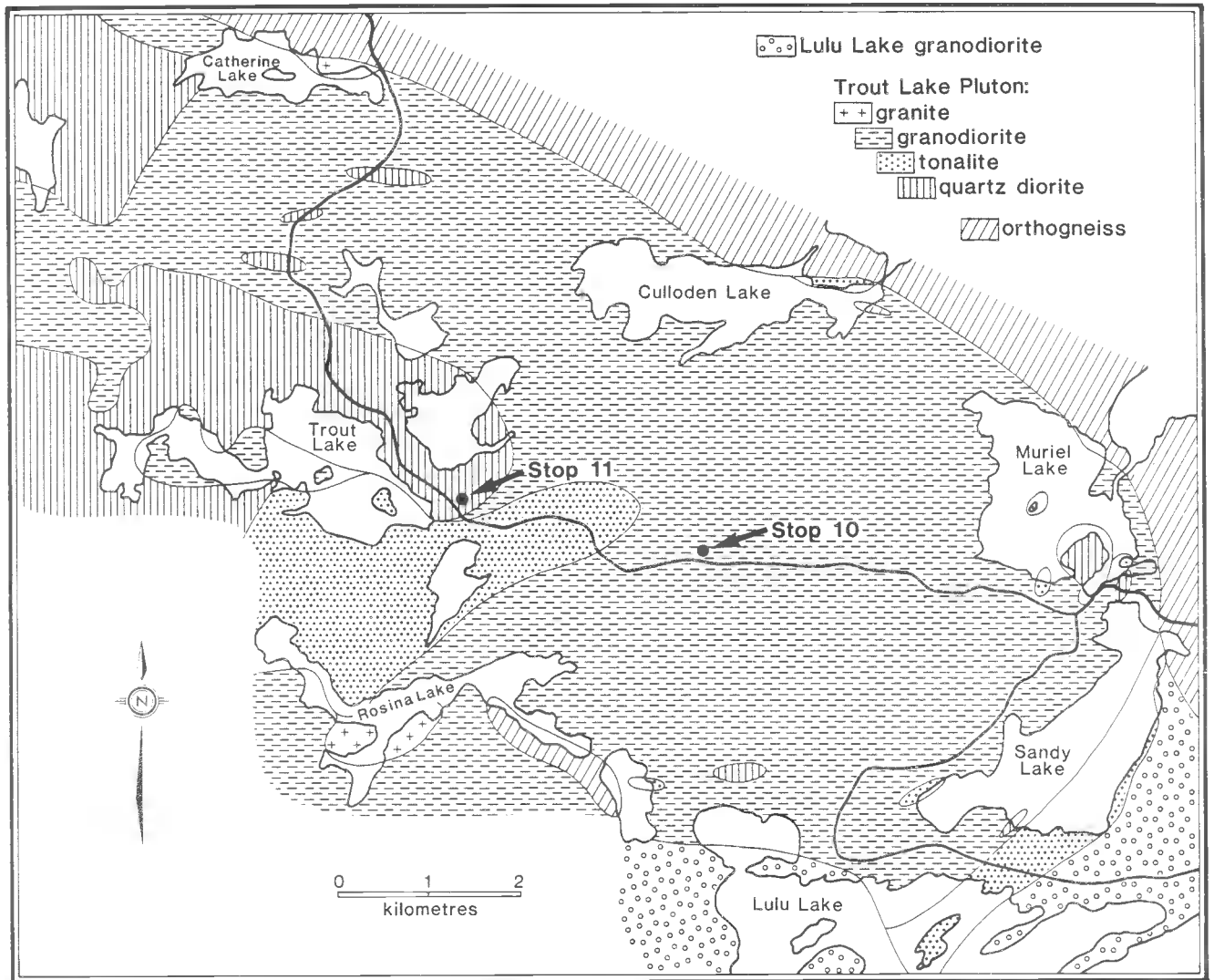


Figure 11: Geologic map of the Trout Lake area (from Beakhouse, 1983) illustrating the locations of stops 10 and 11. The location of this figure is illustrated in figure 4.

representative of the nature of the Winnipeg River belt. This reflects the emphasis of the field trip as well as logistical considerations and the desire to keep field trip participants awake. When considering the Wabigoon-Winnipeg River interface from a regional perspective it is important to keep in mind that relatively highly geochemically evolved, massive, late to post-tectonic granitic rocks predominate in the Winnipeg River belt, in marked contrast to their paucity in the Wabigoon subprovince. Many of the granodiorites within the late plutonic complexes are characterized by the presence of microcline megacrysts such as are abundant in this outcrop. This particular unit is, however, atypical in that amphibolite inclusions are abundant in parts of the unit and both hornblende and biotite (in contrast to biotite alone in most granodiorites) are present. These differences, together with its association with a mafic pluton (Figure 11), suggest it may have a different origin.

- 19.35 STOP 11 (optional) Trout Lake quartz diorite
The outcrop on the east side of the road consists of inhomogeneous hornblende-biotite quartz diorite to diorite containing a wide variety of inclusions which are all more mafic than the host. The nature of the enclaves form a continuum from sharply bounded inclusions to diffusely bounded inclusions to diffuse mafic rich patches. It is not clear how much of the amphibole could be xenocrystic and represent disrupted ultramafic enclaves. A sample from this outcrop yields a U-Pb zircon age of 2700 ± 2 Ma (Beakhouse 1983).

Return to the junction of Highway 596 and the Kenora bypass - turn west (right) on the bypass and proceed 5.7 km and stop just before the bridge over the old Minaki Road and the railroad tracks.

STOP 12 Marginal Granodiorite - Tonalite gneiss contact
Outcrop on both sides of the road contain the contact between the Marginal granodiorite and well banded tonalitic gneiss (Figure 8). The south end of the outcrop consists of intensely sheared megacrystic granodiorite containing a few flattened amphibolite inclusions and highly deformed, discordant leucotonalite dikes. The north end of the outcrop consists of banded amphibolite - tonalite - quartz diorite gneiss. Sills of highly deformed, megacrystic granodiorite are difficult to recognize but occur throughout much of the outcrop but decrease in abundance to the north.

Continue on Kenora bypass to Highway 17, turn east and return to Kenora.

References

- Arth, J.G. and Hanson, G.N., 1972. Quartz diorites derived by partial melting of eclogite or amphibolite at mantle depths: *Contributions to Mineralogy and Petrology*, 37, p.161-174.
- Arth, J.G. and Hanson, G.N., 1975. Geochemistry and origin of the early Precambrian crust of northeastern Minnesota: *Geochimica et Cosmochimica Acta*, 39, p.325-362.
- Bald, R.C., 1981. Petrogenesis of Early Archean, gneissic tonalite - granodiorite from the English River subprovince, Gundy Lake area, northwestern Ontario: Unpublished M.Sc. thesis, University of Manitoba, 120pp.
- Bateman, J.D., 1939. Geology and gold deposits of the Uchi-Slate Lakes area, Ontario Department of Mines, Volume 48, part 8, p.1-43.
- Beakhouse, G.P., 1975. The English River subprovince: Centre for Precambrian Studies, 1975 Annual Report, 51-66.
- Beakhouse, G.P., 1977. A subdivision of the western English River subprovince: *Canadian Journal of Earth Sciences*, 14, 1481-1489.
- Beakhouse, G.P., 1983. Geological, geochemical and Rb-Sr and U-Pb zircon geochronological investigations of granitoid rocks from the Winnipeg River belt, northwestern Ontario and southeastern Manitoba: unpublished Ph.D. thesis, McMaster University, Hamilton, Ontario.
- Beakhouse, G.P., 1985. The relationship of supracrustal sequences to a basement complex in the western English River subprovince. In, Ayres, L.D. Thurston, P.C., Card, K.D. and Weber, W., eds.: *Evolution of Archean supracrustal sequences*, Geological Association of Canada, Special Paper 28.
- Beakhouse, G.P., Stott, G.M., and Sutcliffe, R.H., 1983. Geological Studies in the Kenora-Cedar Lake Area; p.5-10 in *Summary of Field Work, 1983*, by the Ontario Geological Survey, edited by John Wood, Owen L. White, R.B. Barlow, and A.C. Colvine, Ontario Geological Survey, Miscellaneous Paper 116; 313p.
- Blackburn, C.E., 1981. Kenora-Fort Frances sheet. Ontario Geological Survey, Geological Compilation Series, Map 2443, scale 1:253,440.
- Blackburn, C.E., Bond, W.D., Breaks, F.W., Davis, D.W., Edwards, G.R., Poulsen, K.H., Trowell, N.F., and Wood, J., 1985, *Evolution of Archean Volcanic-Sedimentary Sequences of the Western Wbigoon Subprovince and its Margins*: In Ayres, L.D. Thurston, P.C. Card, K.D., and Weber, W., eds., *Evolution of Archean Supracrustal Sequences*: Geological Association of Canada, Special Paper 28.
- Breaks, F.W. and Bond, W.D., 1977. Manifestations of recent reconnaissance investigations in the English River subprovince, northern Ontario: *Geotraverse Conference, 1977*, University of Toronto, p.170-211.
- Breaks, F.W. and Bond, W.D., 1981. Precambrian geology of the English River - Marchington Lake area. Ontario Geological Survey, Preliminary Map P-2293, scale 1:63,360.
- Breaks, F.W., Bond, W.D. and Stone, D., 1978. Preliminary geological synthesis of the English River Subprovince,

- northwestern Ontario and its bearing upon mineral exploration: Ontario Geological Survey, Miscellaneous Paper 72, 55p.
- Brisbin, W.C., 1981. The Winnetka Lake stock, a deformed Archean pluton in northwestern Ontario, abstract. Geological Association of Canada, Abstracts 6, p.A-6.
- Bruce, E.L., 1924. Geology of the upper part of the English River valley, District of Kenora: Ontario Department of Mines, Volume 33, part 4, p.1-11.
- Burke, K. and Kidd, W.S.F., 1978. Were Archean continental geothermal gradients much steeper than those of today? *Nature*, 272, 240-241.
- Burwash, E.M., 1923. Geology of the Ontario - Manitoba boundary (from Winnipeg River to Bloodvein River, 1921): Ontario Department of Mines, volume 32, part 2, p.1-64.
- Butrenchuk, S., 1970. Metamorphic petrology of the Bird River area, southeastern Manitoba: unpublished. M.Sc. thesis, University of Manitoba.
- Campbell, F.H.A., 1971. Stratigraphy and sedimentation of part of the Rice Lake Group, Manitoba: Manitoba Department of Mines and Natural Resources, Mines Branch Publication 71-1, p.135-188.
- Cerny, P., Trueman, D.L., Ziehlke, D.V., Goad, B.E. and Paul, B.J., 1981. The Cat Lake - Winnipeg River and Weskusko Lake pegmatite fields, Manitoba: Manitoba Department of Energy and Mines, Mineral Resources Division, Economic Geology Report ER 80-1, 216p.
- Clark, G.S., Bald, R. and Ayres, L.D., 1981. Geochronology of orthogneiss adjacent to the Archean Lake of the Woods greenstone belt, northwestern Ontario: A possible basement complex: *Canadian Journal of Earth Sciences*, 18, 94-102.
- Condie, K.C. and Hunter, D.R., 1976. Trace element geochemistry of Archean granitic rocks from the Barberton region, South Africa: *Earth and Planetary Science Letters*, 29, 389-400.
- Corfu, F., Beakhouse, G.P., Stott, G.M. and Sutcliffe, R.H., 1985. U-Pb geochronology and tectonic history of the Winnipeg River granitoid terrain, northwestern Ontario, abstract. 31st Annual Institute on Lake Superior Geology.
- Davies, J.C., 1965. Geology of the High Lake-Rush Bay area, District of Kenora. Ontario Department of Mines, Geological Report 41.
- Davies, J.F., Bannatyne, B.B., Barry, G.S. and McCabe, H.R., 1962. Geology and mineral resources of Manitoba: Manitoba Mines Branch publication.
- Davis, D.W. and Trowell, N.F., 1982. U-Pb zircon ages from the eastern Savant Lake - Crow Lake metavolcanic-metasedimentary belt, northwest Ontario: *Canadian Journal of Earth Sciences*, 19, p.868-877.
- Davis, D.W. and Edwards, G.R., 1982. Zircon U-Pb ages from the Kakagi Lake area, Wabigoon subprovince, northwestern Ontario: *Canadian Journal of Earth Sciences*, 19, p.1235-1245.
- Davis, D.W., Blackburn, C.E. and Krogh, T.E., 1982. Zircon U-Pb ages from the Wabigoon-Manitou Lakes region, Wabigoon subprovince, northwest Ontario: *Canadian Journal of Earth*

- Sciences, 19, 254-266.
- Derry, D.R., 1930. Geology of the area from Minaki to Sydney Lake, District of Kenora; Ontario Dept. of Mines, Volume 39, part 3, p.24-41.
- Ermanovics, I., McRitchie, W.D. and Houston, W.W., 1979. Petrochemistry and tectonic setting of plutonic rocks of the Superior province in Manitoba: in Barker, F., ed., Trondhjemites, Dacites and Related Rocks, Amsterdam, Elsevier, p.323-362.
- Eskola, P.E., 1948. The problem of mantled gneiss domes. Quaterly Journal of the Geological Society of London, 104, p.461-476.
- Ferguson, S.A., Brown, D.D., Davis, J.C. and Pryslak, A.P., 1970. Red Lake-Birch Lake Sheet. Ontario Department of Mines, Geological Compilation Series, Map 2175, Scale 1:253,440.
- Freund, H.H. and Turnock, A.C., 1971. Petrology of the paragneiss at Quesnel Lake, Manitoba: Manitoba Department of Mines and Natural Resources. Mines Branch Publ. 71-1, 227-234.
- Glikson, A.Y., 1976. Stratigraphy and evolution of primary and secondary greenstones: significance of data from shields of the southern hemisphere: in Windley, B.F., ed., The Early History of the Earth; John Wiley, New York, p.257-277.
- Glikson, A.Y., 1979. Early Precambrian tonalite-trondhjemite sialic nuclei: Earth Science Reviews, 15, 1-73.
- Goodwin, A.M., 1968. Evolution of the Canadian Shield: Proceedings, Geological Association of Canada, 19, 1-14.
- Goodwin, A.M., 1977. Archean basin-craton complexes and the growth of Precambrian shields: Canadian Journal of Earth Sciences, 14, 2737-2759.
- Goodwin, A.M., Pettijohn, F.J. and Norris, A.W., 1972. Stratigraphic analysis of supracrustal rocks: in Goodwin, A.M. et al., The Superior province, p.527-623: Price, R.A. and Douglas, R.J.W., eds., Variations in Tectonic Styles in Canada, Geological Association of Canada, Special Paper 11, p.529-540.
- Gower, C.F., 1978. The tectonic and petrogenetic history of Archean rocks from the Kenora area, English River subprovince, northwest Ontario: unpublished Ph.D. thesis, McMaster University, Hamilton, Ontario, 622p.
- Gower, C.F. and Clifford, P.M., 1981. The structural geometry and geological history of Archean rocks at Kenora, northwestern Ontario - a proposed type area for the Kenoran orogeny: Canadian Journal of Earth Sciences, 18, p.1075-1091.
- Grunsky, E.C., 1981. An algorithm for the classification of subalkalic volcanic rocks using the Jensen cation plot. p.61-65 in Summary of Field Work, 1981, edited by John Wood, O.L. White, R.B. Barlow and A.C. Colvine. Ontario Geological Survey, Miscellaneous Paper 100.
- Hall, D.H. and Brisbin, W.C., 1981. Geophysics of the Archean crust in Manitoba and northwestern Ontario - An overview (abstract): 1981 Annual Meeting, Geological Association of Canada, Program with abstracts, 6, p.24.

- Hall, D.H., and Brisbin, W.C., 1982. Overview of Regional Geophysical Studies in Manitoba and Northwestern Ontario: Canadian Journal of Earth Sciences, V.19, p.2049-2059.
- Hanson, G.N. and Goldich, S.S., 1972. Early Precambrian rocks of the Saganaga Lake - Northern Light Lake area, Minnesota-Ontario: Part II: Petrogenesis: Geological Society of America Memoir, 135, 179-192.
- Harland, W.B., 1983. Precambrian geochronology in Canada. Geological Magazine, 120, 195-203.
- Harris, N.B.W., 1976. The significance of garnet and cordierite from the Sioux Lookout region of the English River gneiss belt, northern Ontario: Contributions to Mineralogy and Petrology, 55, p.91-104.
- Hart, S.R. and Davis, G.L., 1969. Zircon U-Pb and whole rock Rb-Sr ages and early crustal development near Rainy Lake, Ontario: Geol. Society of America Bulletin, 80, p.595-616.
- Karup-Moller, S. and Brummer, J.J., 1971. Geology and sulphide deposits of the Bird River claim group, southeastern Manitoba: in Turnock, A.C., ed., Geoscience Studies in Manitoba, Geological Association of Canada, Special Paper 9, p.143-154.
- Krogh, T.E., Harris, N.B.W. and Davis, G.L., 1976a. Archean rocks from the eastern Lac Seul region of the English River gneiss belt, northwestern Ontario, part 2. Geochronology: Canadian Journal of Earth Sciences, 13, p.1212-1215.
- Krogh, T.E., Davis, G.L., Ermanovics, I., and Harris, N.B.W., 1976b. U-Pb isotopic ages of zircons from the Berens block and English River gneiss belt: Proceedings of the 1976 Geotraverse Conference, Precambrian Research Group, University of Toronto, p.46.
- Kroner, A., Puustinen, K. and Hickman, M., 1981. Geochronology of an Archean tonalitic gneiss dome in northern Finland and its relation with an unusual overlying volcanic conglomerate and komatiitic greenstone: Contributions to Mineralogy and Petrology, 76, p.33-41.
- McRitchie, W.D. and Weber, W., 1971. Metamorphism and deformation in the Manigotagan gneissic belt, southeastern Manitoba: Manitoba Department of Mines and Natural Resources, Mines Branch, Publication 71-1, 235-284.
- McRitchie, W.D. and Weber, W. (eds.), 1971. Geology and geophysics of the Rice Lake region, southeastern Manitoba (Project Pioneer): Manitoba Department of Mines and Natural Resources, Mines Branch Publication 71-1, 430p.
- Myers, J.S., 1978. Formation of banded gneisses by deformation of igneous rocks: Precambrian Research, 6, p.43-64.
- Nunes, P.D. and Thurston, P.C., 1980. Two hundred and twenty million years of Archean evolution: a zircon U-Pb age stratigraphic study of the Uchi-Confederation Lakes greenstone belt, northwestern Ontario: Canadian Journal of Earth Sciences, 17, p.710-721.
- Nunes, P.D. and Wood, J., 1980. Geochronology of the North Spirit Lake area, District of Kenora - Progress Report: p.7-14, in Summary of Geochronology Studies, 1977-1979, edited by E.G. Pye, Ontario Geological Survey, Miscellaneous Paper 92,

45p.

- Ojakangas, R.W., 1985. Review of Archean Clastic Sedimentation, Canadian Shield: Major Felsic Volcanic Contributions to Turbidite and Alluvial Fan-Fluvial Facies Associations: in Ayres, L.D. Thurston, P.C., and Card, K.D., and Weber, W., eds., Evolution of Archean Supracrustal Sequences: Geological Association of Canada, Special Paper 28.
- Peterman, Z.E., 1979. Strontium isotope geochemistry of late Archean to late Cretaceous tonalites and trondhjemites: in Barker, F., ed., Trondhjemites, Dacites and Related Rocks, Amsterdam, Elsevier, p.133-147.
- Stockwell, C.H., 1964. Age determinations and geological studies. Part II. Geological studies: Geological Survey of Canada, Paper 64-17(II).
- Stone, D., 1977. The Sydney Lake fault zone: Proceedings of the 1977 Geotraverse Conference, Precambrian Research Group, University of Toronto, p.94-108.
- Thurston, P.C. and Breaks, F.W., 1978. Metamorphic and tectonic evolution of the Uchi-English River subprovince: in Metamorphism in the Canadian Shield, Geological Survey of Canada, Paper 78-10, 49-62.
- Thurston, P.C. and Fryer, B.J., 1983. The geochemistry of repetitive cyclical volcanism from basalt through rhyolite in the Uchi-Confederation greenstone belt, Canada. Contributions to Mineralogy and Petrology, 83, 204-226.
- Trowell, N.F., Blackburn, C.E. and Edwards, G.R., 1980. Preliminary synthesis of the Savant Lake - Crow Lake metavolcanic - metasedimentary belt, northwestern Ontario, and its bearing upon mineral exploration. Ontario Geological Survey, Miscellaneous Paper 89, 30p.
- Trueman, D.L., 1975. Bird River - Winnipeg River area: in Summary of Field Work, 1975, Manitoba Department of Mines, Resources and Environmental Management, Geological Services Branch, Geological Paper 2/75.
- van de Kamp, P.C. and Beakhouse, G.P., 1979. Paragneisses in the Pakwash Lake area, English River gneiss belt, northwestern Ontario: Canadian Journal of Earth Science, 16, p.1753-1763.
- Westerman, C., 1977. Tectonic evolution of a part of the English River subprovince, northwestern Ontario: unpublished Ph.D. thesis, McMaster University, Hamilton, Ontario, p.292.
- Wilson, H.D.B., 1971. The Superior province in the Precambrian of Manitoba: in Turnock, A.C., ed., Geoscience Studies in Manitoba, Geological Association of Canada, Special Paper 9, p.41-49.
- Wilson, H.D.B. and Brisbin, W.C., 1963. Nature of the granitic crust in the Superior province: Abstract, Geological Society of America, 76th Annual Meeting, p.180.
- Wilson, H.D.B. and Morrice, M.G., 1977. The volcanic sequence in Archean shields: in Baragar, W.R.A., Coleman, L.C. and Hall, J.M., eds., Volcanic Regimes in Canada, Geological Association of Canada, Spec. Paper 16, p.355-374.
- Wooden, J.L., 1978. Rb-Sr isotopic studies of Archean rocks of the eastern Lac Seul and Kenora areas, English River subprovince, Ontario: in Smith, I.E.M. and Williams, J.G., eds., Proceedings, 1978 Archean Geochemistry Field Conference, University of Toronto Press, Toronto, Ontario, p.131-149.

APPENDIX

Modal and chemical analyses of selected samples from the field trip area

Table 1: Modal analyses for samples from the area of the field trip

Metavolcanic Rocks						Dikes			
Winnipeg River Belt					Wabigoon	Intermediate		Mafic	
275	279	290A	309	312	323	270	272	321B	
plagio- clase	39.3	23.0	26.3	29.9	31.3	33.0	45.9	46.2	25.0
K- feldspar-		-	-	-	-	-	-	-	-
quartz tr		3.0	2.0	1.6	0.3	1.4	-	-	1.8
horn- blende	60.0	73.7	59.0	67.7	67.5	61.1	50.2	48.6	71.0
biotite	-	-	11.2	0.8	0.4	-	1.2	4.5	-
muscovite-		-	-	-	-	-	-	-	-
epidote	0.3	-	-	tr	tr	-	0.9	0.2	-
opaque tr		0.3	0.3	-	-	3.3	1.8	0.1	1.6
sphene	0.4	-	1.2	tr	0.5	-	tr	0.4	0.6
zircon	-	-	tr	-	-	-	tr	tr	-
apatite tr		-	-	-	-	-	tr	tr	-
allanite	-	-	-	-	-	-	-	-	-
carbonate-		-	-	-	-	1.2	-	-	-

Tonalites							Granodiorites	
	262	290B	310	320A	320B	321A	300	313
plagioclase	59.4	61.9	58.9	66.4	71.5	69.7	48.0	40.8
K-feldspar	1.1	0.5	0.3	6.8	2.3	2.3	17.4	13.2
quartz	30.6	26.4	30.5	19.2	22.1	21.0	26.5	34.8
hornblende	-	-	0.9	-	-	-	-	-
biotite	6.2	10.4	6.9	5.4	3.0	5.7	3.8	4.2
muscovite	1.2	-	-	-	-	0.7	2.2	4.6
epidote	1.2	0.8	1.8	2.2	1.0	0.6	1.5	2.1
opaque	tr	tr	0.7	-	-	-	-	-
sphene	-	-	tr	-	-	-	-	-
zircon	tr	tr	tr	tr	tr	tr	tr	tr
apatite	tr	tr	tr	tr	tr	tr	tr	tr
allanite	0.3	tr	tr	-	-	-	-	-
carbonate	tr	-	-	-	-	-	-	-

Table 2: Chemical analyses of selected samples from the area of the field trip.

Rock Type	WABIGON METAVOLCANIC ROCKS				WINNIPEG RIVER BELT METAVOLCANIC ROCKS					WINNIPEG RIVER BELT DIKE ROCKS				WINNIPEG RIVER BELT TONALITES						MARGINAL GRANODIORITE	
Sample Number	83GPB-323	84GPB-8007	84GPB-8009	83GPB-8010	83GPB-275A	83GPB-279	83GPB-290A	83GPB-309	83GPB-312	83GPB-270	83GPB-272	83GPB-304A	83GPB-321B	83GPB-262	83GPB-290B	83GPB-310	83GPB-320A	83GPB-320B	83GPB-321A	83GPB-300	83GPB-313
SiO ₂	54.0	51.2	50.1	62.5	50.2	51.2	50.4	50.9	50.8	54.2	55.3	63.9	51.2	69.9	64.0	73.2	72.0	72.1	72.7	73.1	74.9
Al ₂ O ₃	15.5	13.7	14.4	14.3	13.8	13.9	13.6	13.6	14.3	12.5	13.7	16.0	12.8	16.1	15.9	14.1	14.5	16.3	16.1	14.9	14.5
Fe ₂ O ₃	1.22	1.49	2.04	0.9	3.96	2.30	1.93	3.02	1.61	2.37	2.93	1.31	3.18	0.40	1.52	1.12	0.29	0.06	0.04	0.36	0.22
FeO	9.09	9.67	9.31	5.31	8.00	8.36	10.8	9.96	9.60	5.09	4.95	2.98	11.3	1.31	3.35	1.45	2.69	1.02	1.16	1.09	0.87
MgO	4.23	7.73	7.51	5.09	6.86	8.39	6.64	5.78	6.47	9.55	7.50	2.67	5.65	0.73	1.55	0.43	0.70	0.20	0.45	0.30	0.18
CaO	9.73	11.9	11.3	4.33	10.4	11.50	9.74	10.7	11.9	10.2	8.52	5.02	10.3	3.20	4.49	3.42	3.57	3.02	2.88	2.00	1.70
Na ₂ O	2.68	1.52	2.21	4.07	2.69	1.56	2.26	2.05	1.78	2.87	2.76	3.53	1.74	4.48	3.41	3.71	3.77	5.20	4.61	3.48	3.80
K ₂ O	0.10	0.04	0.12	0.45	0.81	0.21	1.30	0.81	0.43	0.79	1.37	1.82	0.38	1.37	1.83	0.93	1.00	1.03	1.39	3.67	2.97
TiO ₂	1.17	0.89	1.27	0.81	0.97	0.77	1.19	1.17	0.93	0.57	0.70	0.47	1.29	0.20	0.85	0.28	0.38	0.16	0.16	0.21	0.12
P ₂ O ₅	0.04	0.02	0.04	0.09	0.04	0.02	0.04	0.05	0.03	0.22	0.15	0.09	0.05	0.01	0.19	0.03	0.06	0.00	0.00	0.01	0.00
MnO	0.23	0.19	0.20	0.11	0.26	0.18	0.21	0.25	0.21	0.15	0.14	0.07	0.25	0.04	0.07	0.03	0.05	0.02	0.03	0.03	0.04
CO ₂	.24	0.07	0.22	0.11	0.07	0.08	0.07	0.12	0.12	0.08	0.08	0.08	0.06	0.24	0.12	0.06	0.06	0.05	0.10	0.07	0.06
H ₂ O	.76	0.56	0.87	0.52	1.01	0.76	0.70	0.68	0.98	1.04	1.05	0.58	0.87	0.47	1.06	0.30	0.29	0.14	0.42	0.36	0.72
Rb	8	8	9	33	16	11	48	12	19	17	39	51	10	43	94	46	41	25	51	76	85
Sr	110	83	120	170	155	86	76	125	100	610	650	560	105	435	315	250	160	625	350	165	130
Li	8	10	10	16	12	16	22	6	22	18	56	17	20	20	34	22	24	20	28	12	58
Ba	90	70	90	160	130	80	110	120	120	390	770	800	90	370	660	380	260	240	570	840	590
Y	16	12	20	13	16	13	22	18	18	9	13	8	24	3	11	6	7	6	2	3	4
Zr	85	67	86	120	88	64	87	92	78	125	155	155	94	115	225	195	180	120	90	120	95
Co	49	42	41	29	45	43	43	42	41	34	35	17	43	6	14	6	8	<5	<5	5	<5
Cr	304	387	365	118	213	254	129	133	253	570	183	65	116	11	25	7	9	<5	10	7	<5
Cu	172	114	8	50	43	98	89	66	113	22	13	44	70	11	24	10	18	<5	10	8	6
Ni	106	81	95	69	103	128	65	69	98	117	80	38	45	6	9	<5	5	<5	5	<5	<5
Pb	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	11	<10	10	<10	10	12	<10	14	22	18
Zn	92	82	72	82	122	91	107	118	97	76	82	65	126	41	103	124	60	34	26	38	44
La	5	3	6	15	6	4	13	9	7	51	36	26	4	4	53	30	21	21	3	31	15
Ce	12	8	13	29	13	9	10	15	12	17	74	55	12	8	103	59	43	40	10	55	28
Nd	9	6	10	14	9	6	10	10	7	34	36	24	10	3	36	21	14	12	5	19	9
Sm	2	2	5	3	3	2	4	3	1	4	7	8	5	1	8	3	2	3	1	4	1
Sm	2	2	5	3	3	2	4	3	1	4	7	8	5	1	8	3	2	3	1	4	1
Eu	0.9	0.6	0.9	0.9	0.8	0.6	0.8	0.9	0.6	1.0	1.6	1.0	1.1	0.5	1.4	0.8	0.6	0.6	0.3	0.4	0.4
Gd	4	3	5	3	3	3	3	3	3	4	5	3	4	1	6	2	2	2	1	2	1
Dy	2.4	2.0	4.0	2.4	2.9	2.2	3.5	3.3	2.3	1.3	2.7	1.4	4.0	1.2	2.7	0.9	1.2	1.6	0.5	0.6	1.1
Yb	1.9	1.5	2.3	1.3	1.8	1.6	2.4	2.1	2.0	0.9	1.3	0.8	2.7	0.4	1.0	0.6	0.8	0.8	0.2	0.3	0.5

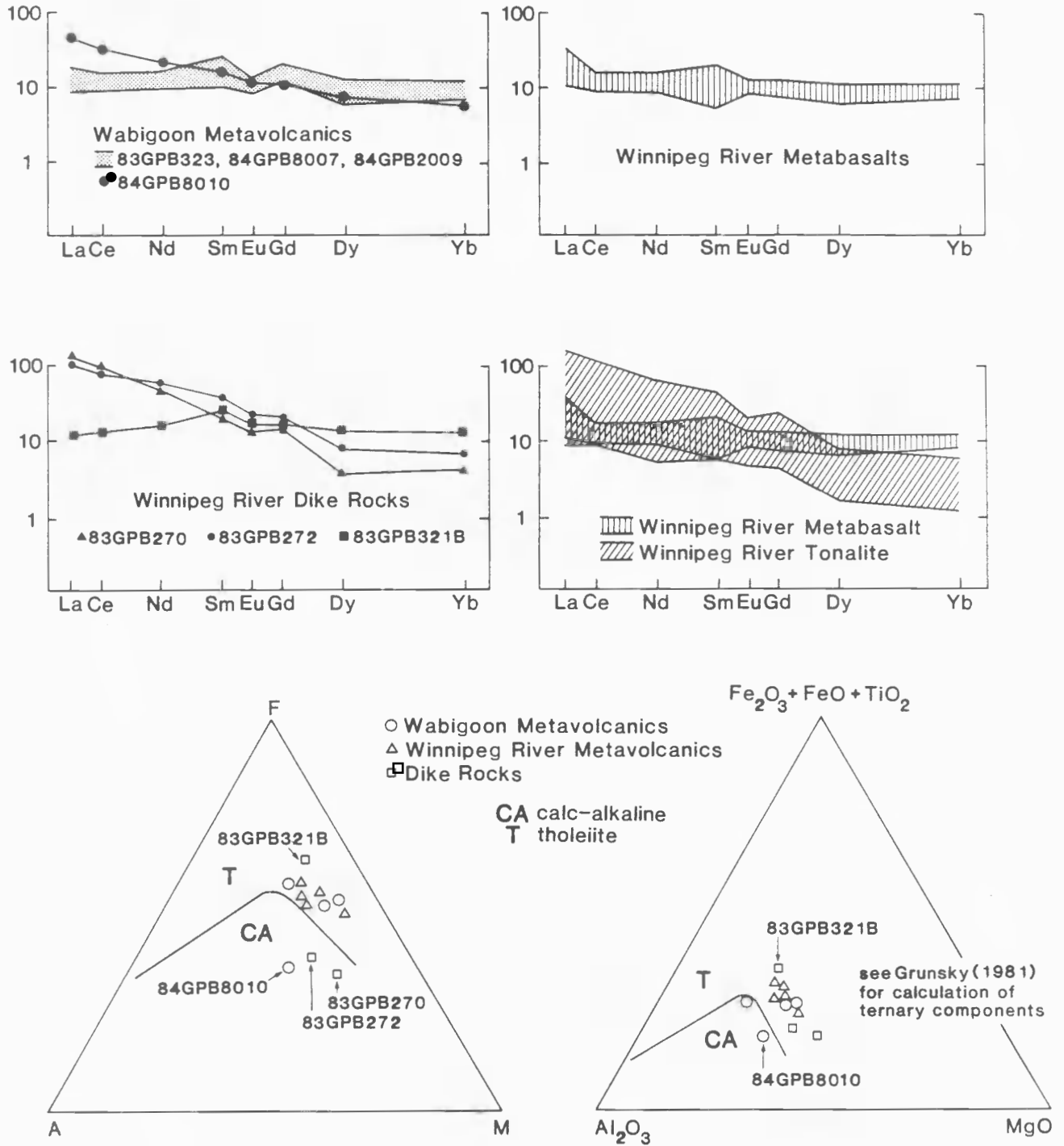


Figure 12. Chondrite normalized rare earth element abundances and major element ternary classification diagrams for samples from the field trip area.

A Volcanic Facies
Interpretation of the Berry River Formation¹

Introductory Discussion and Field Guide
31st Annual Institute on Lake Superior Geology
Kenora, Ontario

by

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INTRODUCTION

The study of Archean pyroclastic rocks has evolved from a simple tuff-agglomerate subdivision to the ternary classifications of Fisher (1966) and Schmid (1981). Now, an important tool in the study of Archean pyroclastics is facies analysis (Ayres, 1977; Tasse et al., 1978 and Lichtblau and Dimroth, 1980). This field trip will examine a deformed amphibolite metamorphic grade Archean subaqueous pyroclastic sequence (the Berry River formation) which the author has subdivided into volcanic facies.

The purpose of this field trip is to examine the evidence for the subdivision of the Berry River formation (Warclub group) into volcanic facies. It is important to study Archean pyroclastic rocks using facies analysis because a facies model, when developed, will lead to an understanding of volcanic stratigraphy and depositional environment and thus the placement of known volcanogenic ore bodies.

The formation was originally described during a 3 year 1:15,840 detailed mapping programme carried out by the Ontario Geological Survey. Results of this mapping (Johns and Richey, 1982; Johns and Davison, 1983 and Johns et al., 1984) are available as preliminary maps from the Office of the Mining Recorder, Ministry of Natural Resources, Kenora. Additional investigations were carried out during the summer of 1984 and the facies model was developed. A preliminary version of the model is outlined in Johns (1984).

The Berry River formation occurs 48 km southeast of Kenora (figures 1 and 2) outcropping along the shore of Long Bay and Lobstick Bay, Lake of the Woods. The formation is part of the Warclub group, a predominantly metasedimentary sequence, found in the western portion of the Archean Wabigoon subprovince of the Superior Province.

SUPERIOR PROVINCE

The Superior Province, comprised of Archean age rocks with regional easterly trending structures, extends in Ontario from the Manitoba border east to the Quebec border and is overlain to the north by Paleozoic sediments and to the south by Proterozoic metasediments and metavolcanics. The Superior Province has been subdivided into subprovinces based on differences in structural styles and lithology (Stockwell et al., 1970).

These subprovinces show some tendency to decrease in age from north to south (Krogh and Davis, 1971 and Krogh et al., 1984) and are referred to as the Sachigo subprovince, Berens River subprovince, Uchi subprovince, English River subprovince (subdivided into the Ear Falls - Manigotagan subprovince and the Winnipeg River subprovince, Beakhouse, 1977), Wabigoon subprovince, Quetico subprovince and Abitibi subprovince. The Berens River and Winnipeg River subprovinces are composed predominantly of granitoid rocks and the Ear Falls - Manigotagan and Quetico subprovinces are metasedimentary terranes. The remaining subprovinces are characterized by comparatively low metamorphic grade metavolcanics and subordinate metasedimentary rocks intruded by granitoid batholiths.

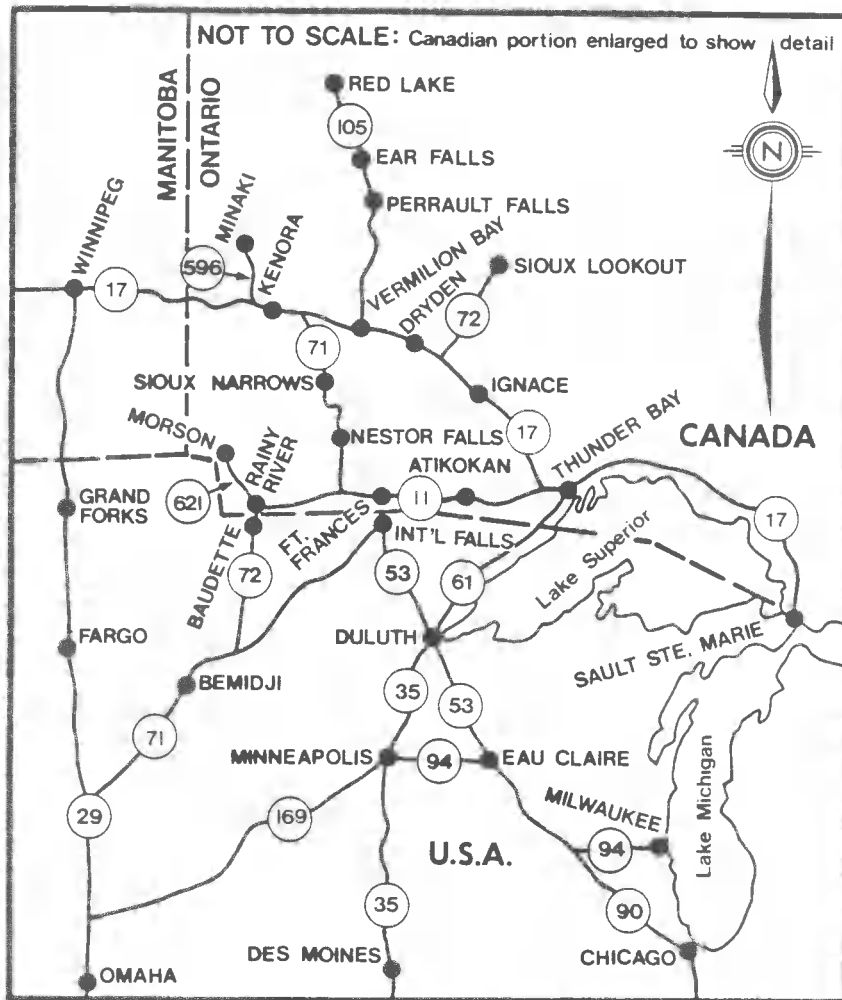


Figure 1: Location of Sioux Narrows, Ontario with respect to that portion of the United States centered on Duluth, Minnesota.

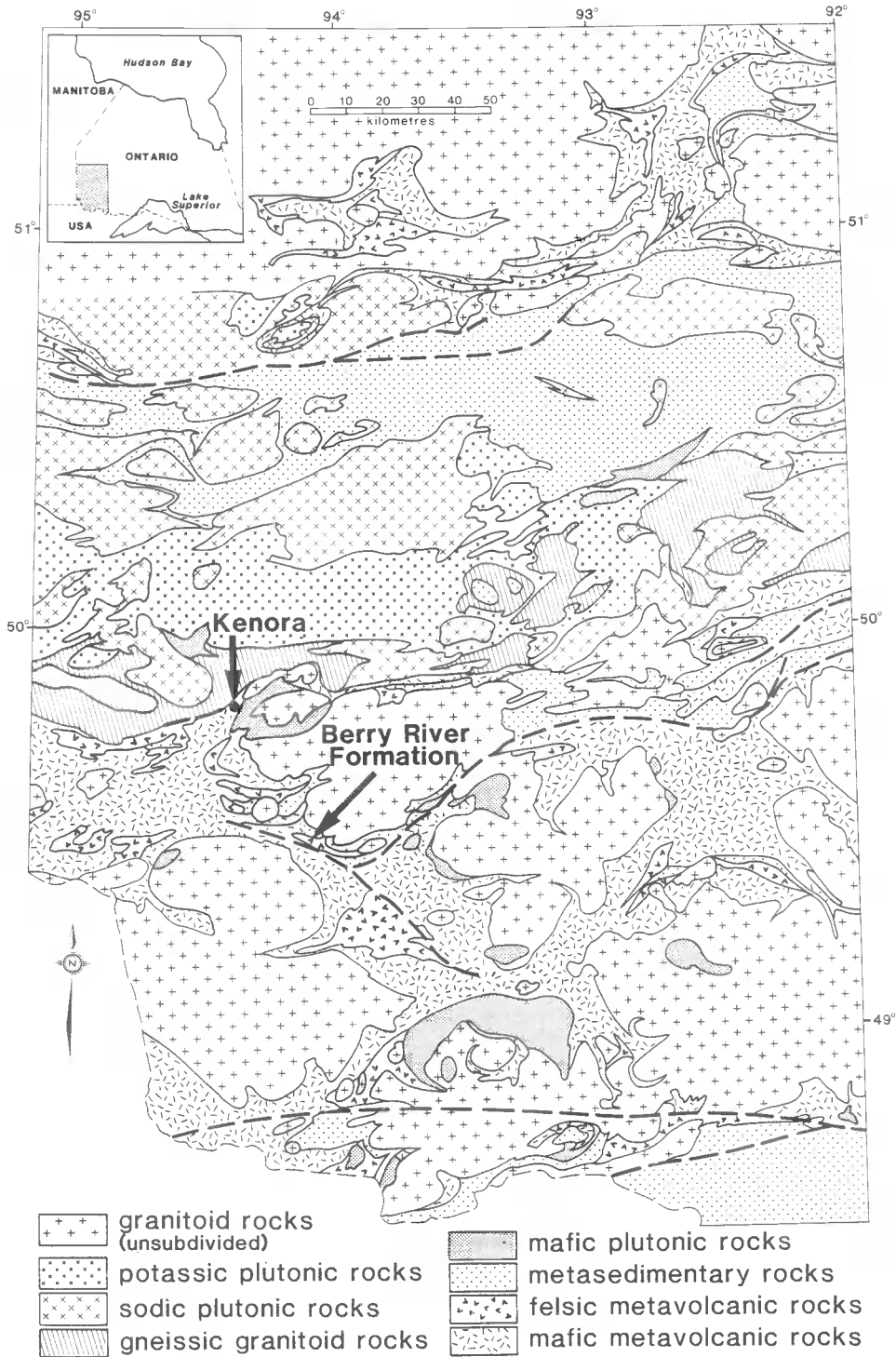


Figure 2: Location of the Berry River formation with respect to the general geology of the western Wabigoon subprovince.

WABIGOON SUBPROVINCE

The Wabigoon Subprovince is an east to northeast trending metavolcanic-metasediment-granitoid terrane. It is bounded on the north by the Winnipeg River subprovince and on the south by the Quetico subprovince. It is approximately 200 km in width and is at least 700 km in length, extending beneath Paleozoic cover of the James Bay lowlands and Williston basin to the east and west respectively.

Trowell et al. (1980) studied the Wabigoon Subprovince between Kakagi (Crow) Lake and Savant Lake and made the following observations. 1) The metavolcanic-metasedimentary belts generally face inwards. 2) Thick basal high Mg tholeiitic basalt assemblages occur at the outer edges of the greenstone belt. 3) Mixed mafic to felsic sequences overlying the basal assemblage occur in the internal portion of the greenstone belts and contain thick sequences of mafic flows that are mostly toward the top of the mixed assemblage. 4) Clastic metasedimentary rocks are associated with the mixed mafic to felsic portion of the metavolcanic sequence. The metavolcanics and metasediments have been isoclinally folded into synforms and antiforms.

Available high precision, U-Pb zircon geochronological investigations of intermediate to felsic metavolcanic rocks in the Wabigoon subprovince suggest that volcanism occurred over a limited time interval between 2745 Ma and 2703 Ma (Davis and Trowell, 1982; Davis and Edwards, 1982, Davis et al., 1982). Davis and Edwards (1982) have determined a U-Pb zircon age of $2713.9 \pm 6.0/-4.4$ Ma for the Berry River formation. The Warclub group, of which the Berry River formation is a part, can be correlated with the Abram group in the Dryden-Sioux Lookout area where a preliminary age of 2716 Ma on a tuffaceous unit (Don Davis, personal communication, 1983) is similar to that age determined for the Berry River formation. These observations suggest that the Warclub group is among the youngest volcano-sedimentary sequences in the Wabigoon subprovince.

LONG BAY - LOBSTICK BAY AREA

The Long Bay - Lobstick Bay area is within the western portion of the Wabigoon Subprovince and is situated between the Savant Lake - Crow (Kakagi) Lake area studied by Trowell et al. (1980) and the classical Lake of the Woods area studied by Lawson (1885) (Figure 2). The metavolcanic - metasedimentary rocks are bounded by the Dryberry Batholith on the northeast and the Aulneau Batholith on the southwest. The forceful intrusion of these two batholiths and the later internal stocks have produced a complex structural pattern (Figure 3).

A general stratigraphy has been developed for the area and is outlined in Figure 4. Two stratigraphic domains occur; one southwest of the Pipestone-Cameron Fault Zone and the other to the northeast (Blackburn, 1981). To date no correlation has been attempted between them.

Southwest of the Pipestone-Cameron Fault Zone the northeast facing Snake Bay formation consists of fine- and medium-grained mafic flows, pillowed and amygdaloidal flows with horizons of autoclastic breccia and pillow breccia and hyaloclastite intruded

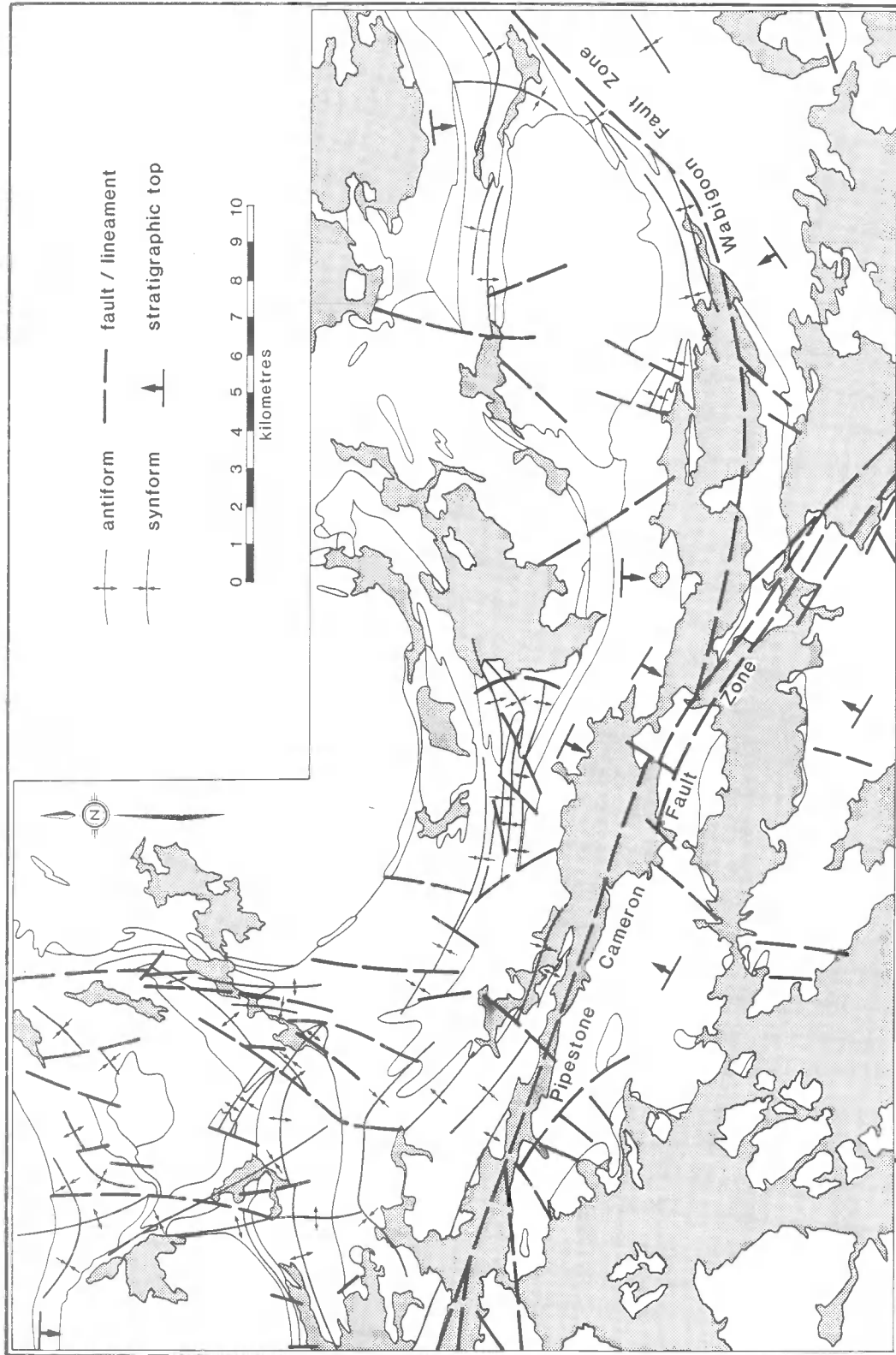


Figure 3: Simplified structural geology map of the Long Bay - Lobstick Bay area. Refer to Figure 4 for the corresponding litho-stratigraphic map.

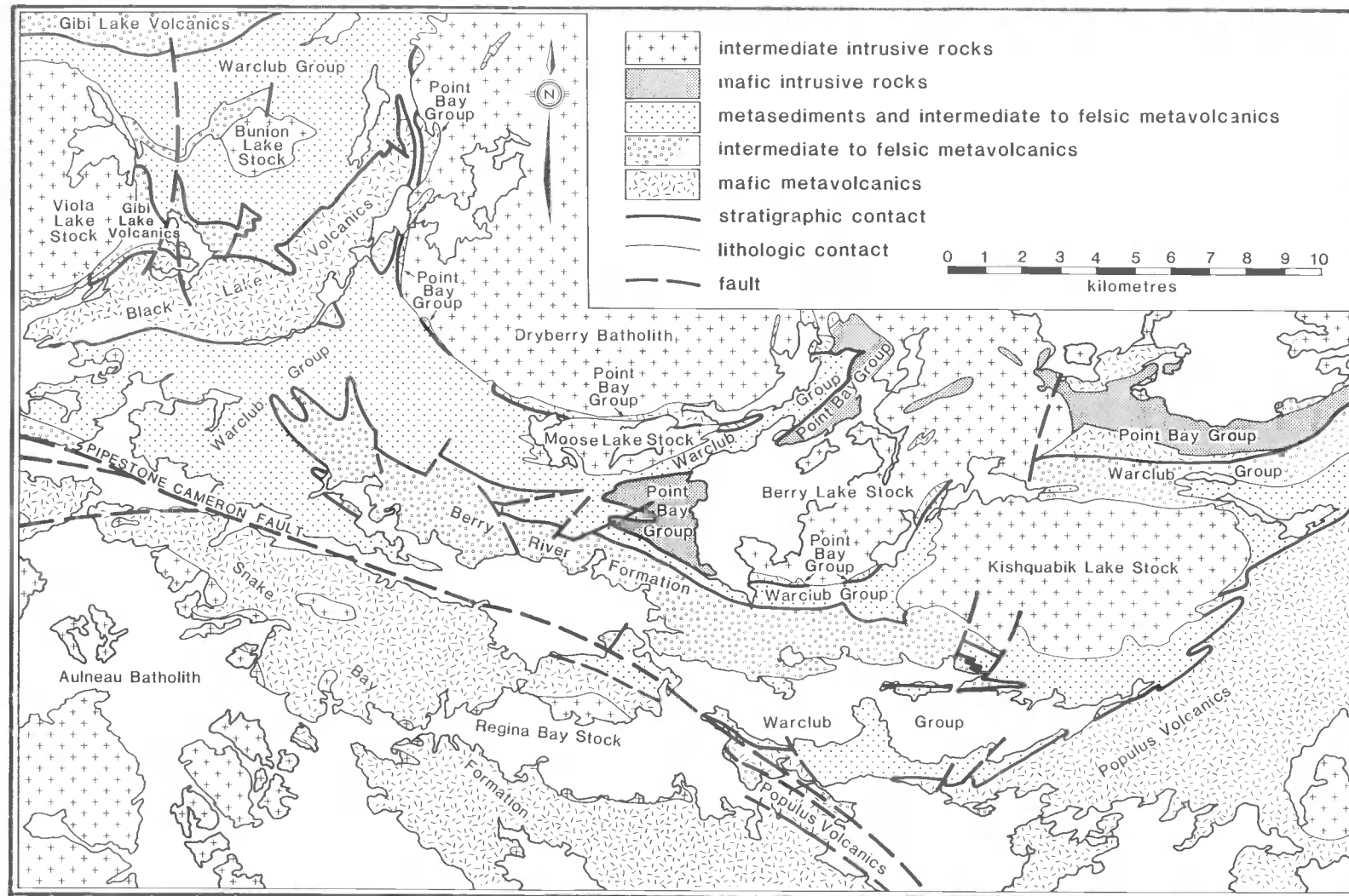


Figure 4: Stratigraphic map of the Long Bay - Lobstick Bay area. Refer to Figure 3 for the corresponding structural interpretation.

by synvolcanic gabbro sills.

The supracrustal rocks northeast of the fault zone are subdivided into 5 informal metavolcanic-metasedimentary groups (Figure 4). They are the Point Bay group, Populus volcanics, Black Lake volcanics, Gibi Lake volcanics and the Warclub group which includes the Berry River formation.

The oldest unit is the Point Bay group which occurs around the margin of the Dryberry batholith. It consists of mafic and intermediate to felsic metavolcanics and metasediments cut by thick differentiated ultramafic to mafic sills. It is overlain by the Warclub group and its base has been totally removed by the intrusion of the Dryberry Batholith. Roof pendants and xenoliths of the Point Bay group are found within marginal phases of the batholith.

The Populus volcanics, Black Lake volcanics and Gibi Lake volcanics are all younger than the Point Bay group and are overlain by the Warclub group (Figure 4). Their mutual stratigraphic relationships are uncertain. The Populus volcanics and the Black Lake volcanics are predominantly mafic metavolcanic sequences. While both sequences contain abundant massive and pillowed flows, the Populus volcanics also contain mafic pillow breccia, hyaloclastite, mafic pyroclastics and intermediate flows and pyroclastics. The Gibi Lake volcanics are composed of intermediate and mafic pyroclastics with minor mafic flows (Trowell, in prep.).

The Warclub group (Figure 4) overlies the units discussed above and is structurally and stratigraphically complex. This group can be traced 71 km northeast to Dryden where it can be correlated with the Abram group. It consists predominantly of metasediments with interbedded intermediate to felsic pyroclastics. The metasedimentary portion of the Warclub group is compositionally varied. North of the Black Lake volcanics wackes are interbedded with arenites while south of the Gibi Lake volcanics thin bedded arenite and siltstone predominate. South of the Black Lake volcanics wacke predominates. Thin beds of magnetite ironstone are found both north and south of the Black Lake volcanics. The variety of lithologies represent various depositional environments and more work is required to reconstruct the paleoenvironment.

The Berry River formation (Figure 4) is a 2 km thick sequence of intermediate to felsic pyroclastic rocks within the Warclub group. Except for the western portion, the formation faces uniformly south (Figure 3). It overlies, with slight unconformity, the wacke in the lower part of the Warclub group and is overlain by a 20-100 m thick mafic metavolcanic unit and wacke of the Warclub group (Johns et al., 1984). The isoclinally folded west end of the formation appears to both overlie and grade laterally into the wacke of the Warclub group. The bedding and foliation planes have near-vertical dips throughout the strike length.

BERRY RIVER FORMATION

The setting of the Berry River formation has been outlined in the previous section and shown in Figure 4. It is a distinct

unit of subaqueously deposited pyroclastics 20 km long and 2 km thick. Rock units were mapped using the size classification of Fisher (1966), making note of volcanic structures and textures. The predominant clast and matrix lithology is both quartz-feldspar porphyry and feldspar porphyry of dacite to rhyolite composition. Thin, fine-grained, metasedimentary horizons composed of reworked pyroclastic material are interbedded with the pyroclastics.

Size classification and sorting alone did not provide sufficient information to properly subdivide the sequence. Volcanic textures and structures, in addition to size and sorting, were used to develop subunits within the formation, as outlined in Table 1. These subunits consist of diagnostic vertical sequences of rock types (after Fisher, 1966) and primary structures which permit definition of distinct depositional events. For example, using the model of Fiske and Matsuda (1964, their Figure 6), the sequence of primary structures expected in a subaqueous pyroclastic flow are: a massive lower portion, which fines upward in terms of non-vesicular material and an upper laminated portion, which also fines upward. The two portions are often referred to as a doubly graded sequence.

Examination of a sequence in the Berry River formation mapped as pyroclastic breccia with tuff interbeds in a purely size and sorting classification (Fisher, 1966) outlined the following vertical sequence from base to top: 2.5 m of homolithic (quartz-feldspar porphyry clasts) pyroclastic breccia grading into lithic lapilli tuff; clasts are subangular and quartz-feldspar tuff matrix supported; this is then overlain by 3 m of moderately sorted, thickly laminated quartz feldspar tuff; bedding is distinct and regular in this portion. This coarse to fine sequence is then overlain by another similar assemblage. The above vertical sequence of primary structures and rock type compares favourably with Fiske and Matsuda's (1964) model and the sequence is most likely a subaqueous pyroclastic flow (Figure 5).

Individual subunits cannot be traced beyond single outcrops but they do form unique packages that can be traced for considerable distance. Subunits representing a single depositional process include debris flow deposits, subaqueous pyroclastic flow deposits, block and ash flow deposits, fall-out tuff deposits and lapilli tuffs and ash flow tuff. Table 1 lists the criteria used to identify subunits within the Berry River formation, and a brief description of each subunit in the formation is listed below.

Debris Flow Deposits (Stop 7)

Debris flow deposits within the Berry River formation are compositionally bimodal (clasts of feldspar porphyry; quartz-feldspar porphyry) to heterolithic, (clasts of feldspar porphyry; quartz-feldspar porphyry; fine grained massive intermediate volcanics; felsic volcanics; mafic volcanics and sediments) matrix supported tuff breccias to lapilli tuffs. These subunits are thick, massive, and poorly bedded. The matrix in any one location corresponds in composition to the majority of framework clasts. The debris flow deposits are mostly tuff-breccia to

Table 1: Subunits identified in the Berry River formation and their characteristics.

Subunit Characteristics	Debris Flow Deposits	Subaqueous Pyroclastic Flow Deposits	Block and Ash Flow Deposits	Fall Out Deposits	Ash Flow Tuff	Reworked Debris Flow Deposits
Bedding	indistinct	distinct	distinct	distinct	absent	distinct
Grain/clast size	coarse	coarse to fine	coarse	medium to fine	fine	coarse to fine
Sorting	poor	moderate to poor	poor	moderate to good	moderate to poor	moderate to good
Grading	absent to normal	normal and reverse	absent to normal	normal to reverse	absent	normal
Composition	bimodal to heterolithic	bimodal	homolithic	bimodal	quartz-feldspar porphyry	heterolithic
Pumice	no	yes	yes	yes	yes	no
Angularity	subround	subround	subangular to angular	subround to subangular	subangular	subround to round
Unique features		double grading	gas escape features	regularity of bedding		interbedded with wacke

lapilli-tuff, but tuff in gradational contact with the coarse part of debris flow deposits forms a fine-grained top in distal debris flow deposits. The debris flows described here are similar to those described by Tasse et al. (1978) in the Abitibi subprovince.

Subaqueous Pyroclastic Flow Deposits (Stops 9 and 11)

Subaqueous pyroclastic flow deposits similar to those originally described by Fiske and Matsuda (1964) occur in the Berry River formation. Massive, compositionally homolithic to bimodal, tuff breccia bases are overlain by caps of thickly laminated feldspar or quartz-feldspar crystal tuff. The massive portion of the flow is graded and the thickly laminated tuffs are subtly graded. These subunits are distinctly bedded with well-preserved base and upper contacts.

Block and Ash Flow Deposits (Stops 3 and 7)

A block and ash flow deposits, a product of explosive volcanism, consists of unsorted ash deposits containing large, generally non-vesicular, jointed cognate blocks which can exceed 5 m in diameter (Fisher, 1982). Block and ash flow deposits in the Berry River formation are thickly-bedded, homolithic pyroclastic breccias that exhibit grading. Thin tuff zones are occasionally preserved as part of these block and ash flow deposits. These tuff zones represent either ash cloud deposits or ground surge deposits. The depositional characteristics displayed by these deposits can be explained in terms of the depositional model proposed by Sparks et al. (1973).

Fall-Out Deposits (Stop 9 and 11)

Fisher and Schmincke (1984, their Figure 7-1) outlined the characteristics of submarine fall-out tephra. Characteristics of the deposits formed in such an environment preserved in the Archean are: thickness of single layers commonly less than 50 cm; sorting good to poor; plane parallel beds; normal grading from crystal and lithic rich bases to shard rich tops; and grading may be inverse if pumice is present.

Fall-out deposits in the Berry River formation are thick sub-units of interbedded, thickly laminated to thinly bedded crystal tuff and pumice crystal lapilli tuff. Although bedding is regular and distinct, contacts between individual beds are gradational. The pumice is recessively weathered and darker than the crystal matrix.

Ash Flow Tuff Deposits (Stops 4 and 10)

Ross and Smith (1961) have defined ash flow tuff as consolidated deposits of volcanic ash resulting from the ash flow eruptive process. They define the ash flow process as a turbulent mixture of gas and pyroclastic materials of high temperature ejected explosively from a crater or fissure, that travels swiftly down the slopes of a volcano or along the ground surface.

Ash flow tuff deposits of varied thicknesses are found throughout the Berry River formation. They are massive, homogeneous, fine-grained quartz-feldspar porphyry. Minor variations

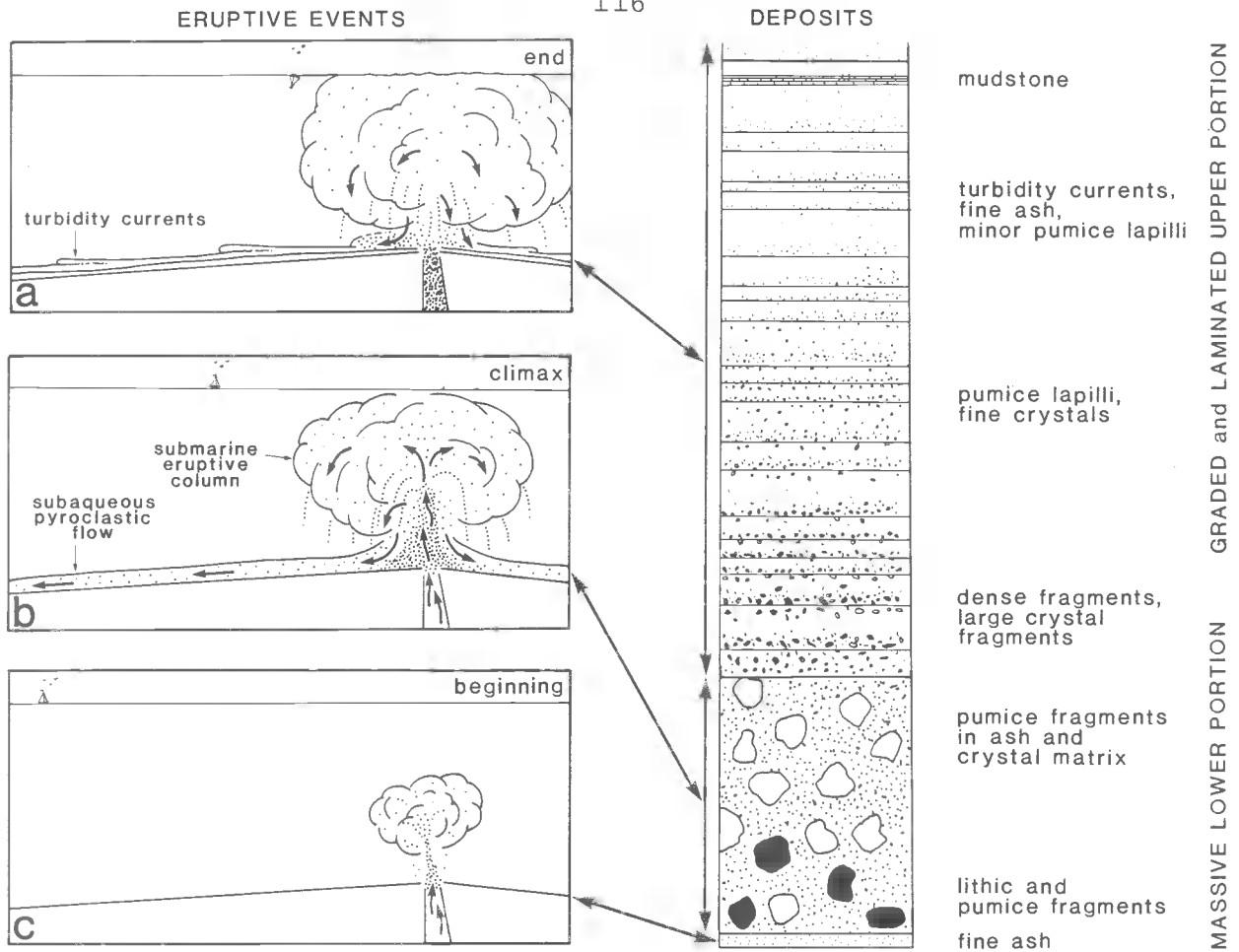


Figure 5:

Schematic drawings of a submarine eruption producing subaqueous pyroclastic flows, and subsequent appearance of the deposits of such an eruption (Easton and Johns, in press).

C) Beginning of eruption. Vesiculating magma is erupted into sea water. Some fine ash may be deposited near the vent.

B) Climax of eruption. Submarine eruption column carried much debris high into suspension. Sorting splits the debris into various fractions. Buoyant pumice floats; dense fragments, large crystals and compact pumice lapilli settle around the vent, and are transported laterally in a subaqueous pyroclastic flow. Most ash remains in suspension.

A) End of eruption. Steady pyroclastic flow ceases as amount of erupted material decreases and is replaced by turbidity current flow. Later turbidity currents contain finer and less dense ash that has settled from suspension.

As shown in the right-hand side of the figure, an important characteristic of subaqueous pyroclastic deposits are their doubly-graded nature. Each bed is graded, and the beds at the base of the sequence contain coarser and denser ash than the beds at the top of the sequence. Modified from Fiske and Matsuda (1964) and Fiske (1969).

within a thicker layer indicates deposition by a number of separate events forming a single cooling unit.

Petrography

Thin section examination of finer grained portions of the formation shows recrystallization to various degrees and much alteration with secondary carbonate, sericite and epidote. The angularity of ash-sized crystals and lithic fragments can often be observed but generally the recrystallization and schistosity has destroyed all primary features.

Geochemistry of the Berry River Formation

Table 2 lists 20 whole rock analyses of finer grained clastic samples of the Berry River formation carried out by the Geoscience Laboratories, Ontario Geological Survey, Toronto. According to the AFM ternary diagram of Irvine and Baragar (1971) and the Al-Fe-Ti-Mg molecular proportion ternary diagram of Jensen (1976), the pyroclastic rocks have calc-alkalic affinities (Figure 6). The single high-magnesium tholeiitic basalt is from the thin metavolcanic unit capping the formation. The classification schemes of Irvine and Baragar (1971) (Table 2) and Jensen (1976) (Figure 6) respectively class the samples as dacite and rhyolite. No discernable trend was noted in the major element chemistry with respect to stratigraphic height. The close clustering of the points plotted in Figure 6 may indicate a similar magma source for all the pyroclastic sub-units.

Figure 6 includes an AFM ternary diagram (Irvine and Baragar, 1971) and an Al-Fe-Ti-Mg molecular proportion ternary diagram (Jensen, 1976) of metasedimentary and metavolcanic rocks from the rest of the Warclub group within the Long Bay - Lobstick Bay area. Data for these plots are found in Johns (in prep). These plots are included to allow comparison between the Berry River formation and the metasediments and metavolcanics of the rest of the Warclub group.

FACIES MODELS

Historically, mapping of Archean pyroclastic rocks has supplied insufficient data to understand complex stratigraphy. Figure 7 shows that the complexity of volcanic stratigraphy depends on the location of the section. Mapping by clast size parameters only will not prove useful in developing map units and facies models. It is difficult to compare ancient and modern pyroclastic sequences without a clear idea of stratigraphy and using facies analysis a stratigraphic context can be developed.

Discrete packages of pyroclastic subunits can be traced and correlated allowing for the development of facies models. Through development of stratigraphy, facies, and subunits within Archean pyroclastic sequences comparisons can be made with other Archean pyroclastic sequences and they can then be compared with younger pyroclastics thus assisting in the development of paleo-environmental models for Archean volcanism.

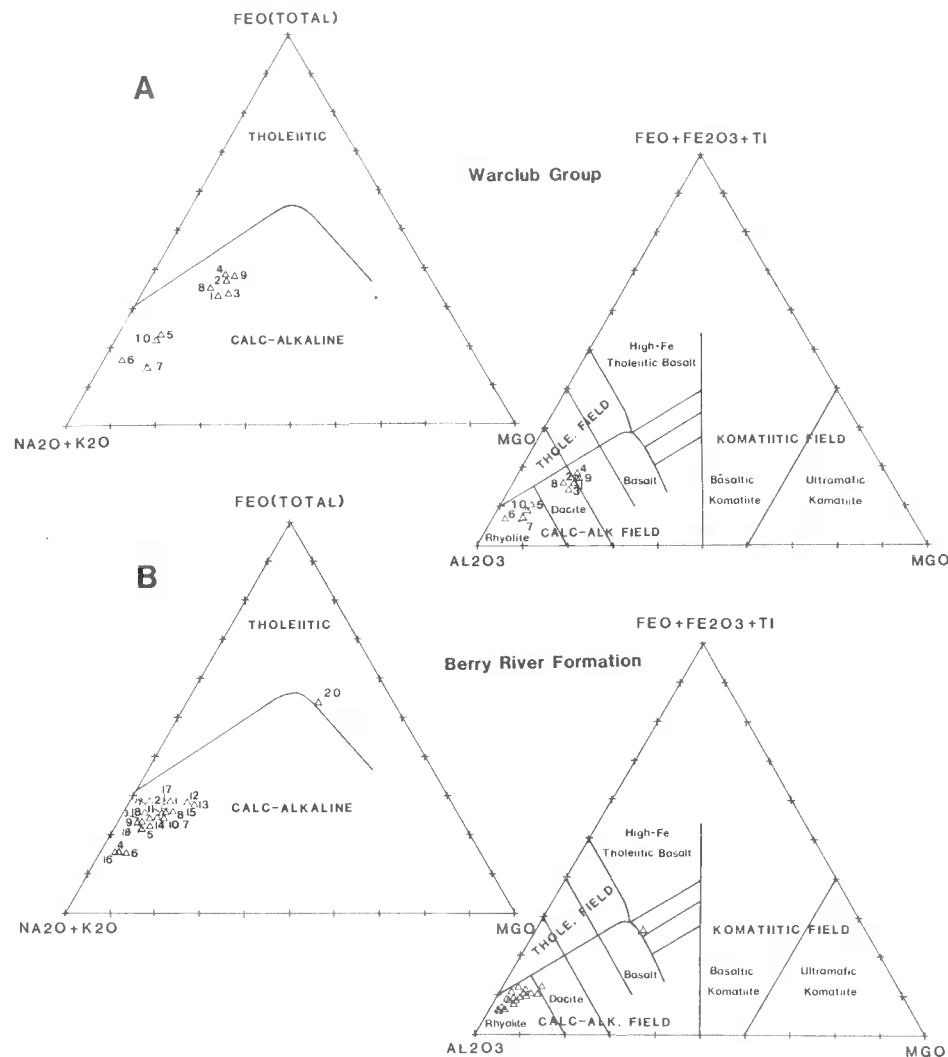
Facies are based on a number of parameters described in Fisher and Schmincke (1984): position relative to source, environment of deposition and primary composition.

Table 2: Chemical analyses of samples from the Berry River formation (Geoscience Laboratories, Ontario Geological Survey, Toronto). The Irvine and Baragar (1971) rock name and the volcanic facies designation are at the bottom of the table.

Sample No.	81-GWJ 2050	81-GWJ 2054	82-GWJ 169	82-GWJ 491	82-GWJ 601	82-GWJ 611	82-GWJ 662	83-GWJ 4172	83-GWJ 4173	83-GWJ 4174
Latitude	49.48657	49.45833	49.43382	49.43269	49.44003	49.44484	49.42917	49.43213	49.42902	49.42833
Longitude	94.12201	94.07147	93.97664	93.94540	93.98581	93.99126	93.94431	93.94319	93.94560	93.94582
SiO ₂	61.20	67.70	69.30	70.30	66.40	69.10	68.50	68.10	67.40	73.20
Al ₂ O ₃	18.40	16.10	16.80	17.10	18.30	17.90	17.30	16.90	16.60	16.50
Fe ₂ O ₃	0.39	0.76	0.73	0.34	0.80	0.42	0.77	0.92	0.72	0.33
FeO	2.58	1.85	1.39	0.95	1.53	1.02	1.17	1.23	1.30	0.89
MgO	0.95	0.40	0.49	0.34	0.80	0.55	0.76	0.41	0.73	0.37
CaO	5.17	3.78	2.34	2.27	2.55	1.67	4.03	3.03	5.10	1.87
Na ₂ O	4.44	4.52	4.95	4.90	6.05	5.33	4.04	4.49	4.16	1.26
K ₂ O	1.98	1.38	1.32	1.61	1.08	1.86	1.00	2.01	0.73	2.95
TiO ₂	0.35	0.29	0.29	0.33	0.42	0.37	0.29	0.36	0.34	0.81
P ₂ O ₅	0.08	0.06	0.03	0.03	0.06	0.04	0.03	0.08	0.07	0.08
MnO	0.06	0.07	0.05	0.06	0.05	0.03	0.03	0.06	0.04	0.04
CO ₂	1.92	1.06	1.06	0.65	0.71	0.44	0.87	0.86	1.43	0.18
S	0.01	0.01	0.01	0.02	0.00	0.12	0.00	0.01	0.01	0.01
H ₂ O ⁺	1.02	0.44	0.98	0.32	0.62	0.54	0.63	0.95	0.96	1.51
H ₂ O ⁻	0.11	0.11	0.12	0.07	0.07	0.06	0.06	0.03	0.04	0.05
LOI	2.90	1.60	1.60	1.00	1.20	1.00	1.10	1.70	1.90	1.20
Total	98.66	98.53	99.86	99.29	99.44	99.45	99.48	99.44	99.63	99.55
Ba	320	370	320	320	350	360	260	810	270	620
Be					1					
Ce								35	60	
Co	10	9	5		7	6	6	5	5	
Cr	8	6	8	8	13	8	7	7	5	8
Cu	16	9	28	5	15	8	9	6	8	
Li	12	10	12	12	17	10	17	14	18	10
Nb			5	4	6	7	7	6	4	6
Nd	6					5	6	5		
Ni										
Pb										
Sc		420	4	4	6	6	5	4	5	51
Sr	500		490	370	430	305	355	285	370	125
Rb								40	20	60
V	55	35	35	30	40	45	30	35	35	22
Y		7	2	2	4	2	4			
Zn	52	38	48	24	84	49	36	30	28	32
Zr	55	45	45	65	70	75	45	45	45	45
Irvine and Baragar rock name	Andesite	Dacite	Dacite	Dacite	Dacite	Rhyolite	Dacite	Dacite	Dacite	Rhyodacite
Volcanic Facies	Proximal-Distal	Proximal-Distal	Ash Flow Tuff	Ash Flow Tuff	Proximal-Distal	Ash Flow Tuff	Proximal	Ash Flow Tuff	Proximal	Proximal

Table 2 cont.

Sample No.	83-GWJ 4176	83-GWJ 4177	83-GWJ 5279	83-GWJ 5282	83-GWJ 5283	83-GWJ 5286	83-GWJ 5294	83-GWJ 5295	83-GWJ 5296	83-GWJ 5297
Latitude	49.42677	49.42537	49.43636	49.43636	49.44095	49.44597	49.44426	49.44313	49.44200	49.4369
Longitude	93.95789	93.96184	93.93486	93.92873	93.99055	93.98924	93.98683	93.98486	93.98421	93.00328
SiO ₂	67.10	50.20	68.90	66.00	69.30	66.30	65.40	70.00	64.80	64.10
Al ₂ O ₃	16.00	14.50	16.50	15.80	15.70	16.90	17.80	16.80	16.30	15.20
Fe ₂ O ₃	0.64	1.84	0.82	1.57	0.29	0.53	1.11	0.47	1.24	1.90
FeO	1.85	8.43	1.44	1.65	2.13	1.65	1.78	0.82	1.58	0.96
MgO	0.83	5.56	0.78	1.41	1.30	0.60	1.19	0.28	0.83	0.52
CaO	3.09	10.20	3.41	3.77	3.80	4.71	3.78	3.36	5.14	4.85
Na ₂ O	4.60	2.98	3.87	5.06	3.61	4.72	4.81	4.68	4.80	3.16
K ₂ O	2.08	0.15	1.74	1.29	1.37	1.30	1.94	1.86	1.59	4.10
TiO ₂	0.36	1.27	0.39	0.40	0.40	0.45	0.41	0.41	0.42	0.38
P ₂ O ₅	0.10	0.12	0.08	0.13	0.08	0.10	0.08	0.08	0.09	0.14
MnO	0.05	0.18	0.03	0.06	0.05	0.05	0.06	0.03	0.10	0.07
CO ₂	1.34	2.00	0.72	1.56	0.30	1.06	0.08	0.54	2.02	3.26
S	0.02	0.21	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01
H ₂ O ⁺	0.92	1.10	0.86	0.75	1.07	0.56	0.78	0.73	0.77	1.08
H ₂ O ⁻	0.06	0.06	0.06	0.05	0.09	0.02	0.04	0.02	0.07	0.03
LOI	2.20	2.70	1.30	1.80	1.20	1.70	0.60	1.20	2.40	4.10
Total	99.04	98.80	99.61	99.51	99.51	98.96	99.27	100.08	99.76	99.76
Ba	580	90	390	320	240	320	330	380	350	940
Be	1	1		1						1
Ce	75	135	55	111	65	60	55	45	70	110
Co	6	37	7	7	7	6	7	5	6	6
Cr	6	204	7	10	7	10	7	7	10	6
Cu	5	64	6	5	26	16	12	11	32	10
Li	10	6	38	30	16	9	10	8	14	12
Nb	8	10	7	5	7	7	6	6	6	8
Nd		45		35						35
Ni	5	110	11	6	5	5	6			
Pb										
Sc	4	23	4	4	5	5	5	4	5	4
Sr	364	310	205	430	305	305	385	315	360	400
Rb	50		40	30	40	40	60	30	30	70
V	40	235	45	45	40	55	45	50	50	40
Y		21		8		2		2		5
Zn	45	143	26	38	48	44	47	44	56	46
Zr	50		50	70	35	14	13	32	35	55
Irvine and Baragar rock name	Dacite	Basalt (tholeiitic)	Dacite	Dacite	Dacite	Dacite	Dacite	Dacite	Dacite	Dacite
Volcanic Facies	Ash Flow Tuff	Proximal	Proximal	Proximal	Proximal- Distal	Proximal -Distal	Ash Flow Tuff	Ash Flow Tuff	Ash Flow Tuff	Ash Flow Tuff



Point No.	Sample No.
1	81GWJ-2044 Int.,Tuff
2	81GWJ-2045 Wacke,Feld.
3	81GWJ-2049 Int.,Tuff
4	81GWJ-2051 Wacke,Feld.
5	81GWJ-2053 Porph.,Feld.
6	81GWJ-2055 Porph.,Qtz/Feld.
7	81GWJ-2048 Arenite
8	82GWJ-0008 Int.,Tuff
9	82GWJ-0691 Wacke
10	82GWJ-0092 Wacke

Point No.	Sample No.
1	81GWJ-2050 Int.,Tuff
2	81GWJ-2054 Porph.,Qtz/Feld.
3	82GWJ-0169 Int.,Tuff
4	82GWJ-0491 Tuff Breccia
5	82GWJ-0601 Tuff
6	82GWJ-0611 Porph.,Feld.
7	82GWJ-0662 Lapilli Tuff
8	83GWJ-4172 Porph.,Qtz/Feld.
9	83GWJ-4173 Int.,Tuff
10	83GWJ-5279 Int.,Tuff
11	83GWJ-5282 Porph.,Qtz/Feld.
12	83GWJ-5283 Int.,Tuff
13	83GWJ-5286 Tuff,Feld.
14	83GWJ-5294 Tuff
15	83GWJ5295 Int.,Tuff
16	83GWJ-5296 Tuff,Feld
17	83GWJ-5297 Porph.,Qtz/Feld.
18	83GWJ-4174 Tuff
19	83GWJ-4176 Tuff Breccia
20	83GWJ-4177 Mafic Flow

Figure 6: AFM ternary diagram (Irvine and Baragar, 1971) and Al-Fe-Ti-Mg molecular proportion ternary diagram (Jensen, 1976) of wackes and pyroclastics from the Warclub group (top) and Berry River formation (bottom).

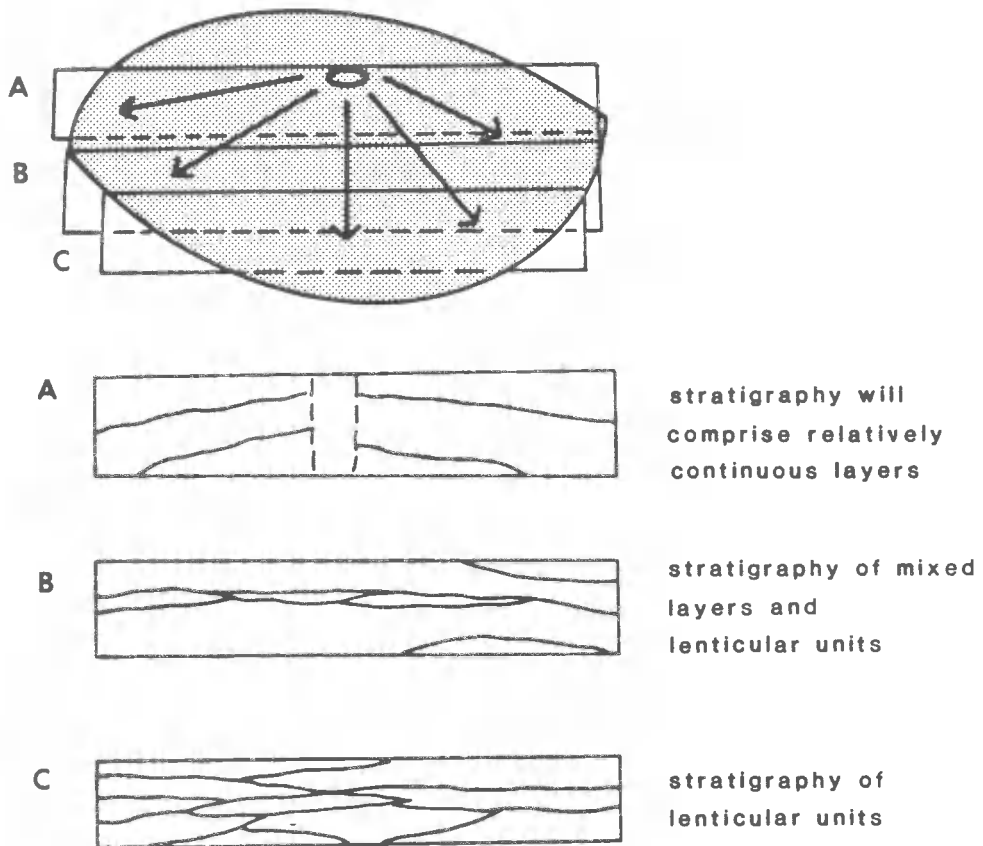


Figure 7: Stratigraphic sections through a volcano illustrating the continuity and complexity of stratigraphy in different sections.

Position Relative to Source

Fisher and Schmincke (1984) have subdivided this parameter into: near-source facies; intermediate-source facies and distant facies. These are relative terms and distance values vary with topography, eruption strength, subaerial vs. subaqueous environment etc.. Williams and McBirney (1979) have provided numerical estimates for these terms; central facies (0.5-2 km from vent); proximal facies (5-15 km from vent) and distal facies (greater than 5-15 km from vent). Providing allowances are made for differences in method of deposition and in the mode of preservation, these concepts can be applied to Archean pyroclastic deposits (Table 3). The products listed in table 3 for each facies/zone can be related to a large central vent composite volcano (Figure 8).

Environment of Deposition

This method of facies analysis considers all physical, chemical and geological features that affect deposition. Such facies models have yet to be fully exploited in studies of Archean pyroclastic rocks. The facies model of the Berry River formation presented later attempts to postulate an environment of deposition.

Primary Composition

In complex areas, such as near vent, a petrologic-chemical approach may be used to separate assemblages into broad compositional categories to aid the determination of stratigraphic relationships. Note however that if this were the sole approach here then the entire Berry River formation would be classed as one facies (Figure 6); calc-alkaline dacite.

FACIES MODEL FOR THE BERRY RIVER FORMATION

Figure 4 portrays the stratigraphic setting of the formation within the Long Bay-Lobstick Bay area. Various structural complications are shown in Figure 3. Subunits based upon recognition of depositional processes were erected within the formation as previously described. These subunits were categorized in terms of distance from source, environment of deposition and degree of reworking based on primary structures and lithological association.

The Berry River formation developed during two episodes of explosive volcanism. The first episode resulted in subaqueous deposition of the proximal-distal deposition facies along with thin ash flow tuffs (Figure 9). During the second episode involving the same volcano (Figure 6), eruptions may have been subaerial but deposition was subaqueous in the proximal deposition facies. A thick ash flow tuff is also associated with this eruptive event (Figure 9). A subvolcanic quartz-feldspar porphyry was intruded into the vent complex of the proximal deposition facies. Adjacent to this is a facies of reworked coarse to fine pyroclastics interbedded with and prograding into wackes of an epiclastic facies (Figure 9).

Table 3: Products associated with the four main volcanic facies of a central vent as shown in Figure 8. Johns et al. (1983). Adapted from Williams and McBirney (1979).

VENT FACIES

(0.5-2 km from vent)

Depositional	{	dikes, sills and domes co-ignimbrite deposits phreatomagmatic deposits talus breccia, megabreccia
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PROXIMAL FACIES

(2-15 km from vent)

Depositional	{	air fall deposits (tuffs) pyroclastic flows subaqueous pyroclastic flows lava flows and domes
--------------	---	--

Redeposited	{	<table border="0"> <tr> <td style="vertical-align: middle;">Recognizable as pyroclastics</td> <td style="vertical-align: middle;">{</td> <td style="vertical-align: middle;">Lahars pyroclastic flows tuffs</td> </tr> <tr> <td style="vertical-align: middle;">Recognizable as volcanic sediments</td> <td style="vertical-align: middle;">{</td> <td style="vertical-align: middle;">debris flows arenites wackes</td> </tr> </table>	Recognizable as pyroclastics	{	Lahars pyroclastic flows tuffs	Recognizable as volcanic sediments	{	debris flows arenites wackes
Recognizable as pyroclastics	{	Lahars pyroclastic flows tuffs						
Recognizable as volcanic sediments	{	debris flows arenites wackes						

DISTAL FACIES

(>5-15 km from vent)

Depositional	{	air fall deposits (tuffs) pyroclastic flows subaqueous pyroclastic flows lava flows
--------------	---	--

Redeposited	{	<table border="0"> <tr> <td style="vertical-align: middle;">Recognizable as pyroclastics</td> <td style="vertical-align: middle;">{</td> <td style="vertical-align: middle;">lahars pyroclastic flows tuffs</td> </tr> <tr> <td style="vertical-align: middle;">Recognizable as volcanic sediments</td> <td style="vertical-align: middle;">{</td> <td style="vertical-align: middle;">debris flows arenites wackes siltstones</td> </tr> </table>	Recognizable as pyroclastics	{	lahars pyroclastic flows tuffs	Recognizable as volcanic sediments	{	debris flows arenites wackes siltstones
Recognizable as pyroclastics	{	lahars pyroclastic flows tuffs						
Recognizable as volcanic sediments	{	debris flows arenites wackes siltstones						

EPICLASTIC FACIES

Redeposited	{	<table border="0"> <tr> <td style="vertical-align: middle;">Talus debris flows sediments</td> <td style="vertical-align: middle;">{</td> <td style="vertical-align: middle;">in crater lakes (active, extinct) perched ponds (Santiaguito) alluvial fans (Caribbean)</td> </tr> </table>	Talus debris flows sediments	{	in crater lakes (active, extinct) perched ponds (Santiaguito) alluvial fans (Caribbean)
Talus debris flows sediments	{	in crater lakes (active, extinct) perched ponds (Santiaguito) alluvial fans (Caribbean)			

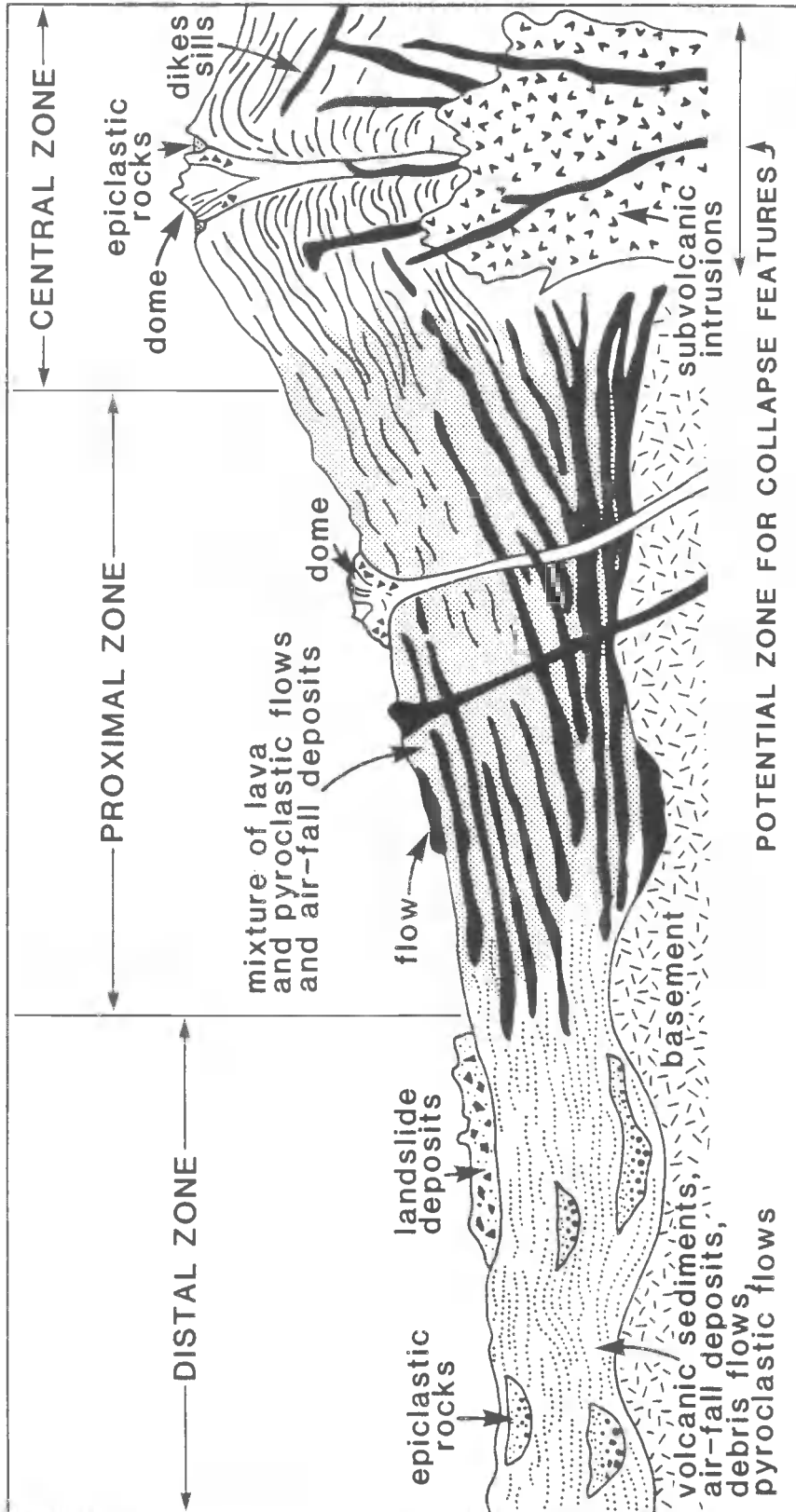


Figure 8:

Principal facies variation in volcanic rocks related to a large central vent composite volcano. Central zone is also known as the vent facies. Epiclastic facies can occur in all three zones. Products of each zone/facies are listed in Table 3. Easton and Johns (in press) modified from Williams and McBirney (1979).

Description of Volcanic Facies Components

Each volcanic facies within the Berry River formation has a unique character, as described below. Figure 10 illustrates the various styles of deposition for the facies of the Berry River formation.

Proximal - Distal Deposition Facies (Stops 9 and 11)

The designation proximal-distal deposition facies does not indicate a facies progressing from a proximal setting to a distal setting. It instead refers to a facies association without sufficient distinct primary features which permit assignment to one environment. Therefore this facies represents both proximal and distal environments. It is a unit formed by deposition directly from the volcano without significant reworking. The proximal-distal deposition facies extends the length of the formation (Figure 9). While the eastern and western portions are similar, the central portion is of a different nature.

In the east and west the deposits are massive, interbedded, poorly-sorted, coarse, medium and fine, compositionally homolitic to bimodal debris flows. The bedding planes between deposits are generally indistinct indicating rapid accumulation. Interbedded with these massive deposits are discontinuous ash flow tuff. These discontinuous tuffs are in similar stratigraphic positions (Figure 9) indicating either erosion after deposition or topographic control of deposition. Figure 10, stage A is a sketch of these depositional events.

The central portion, particularly south-southwest of Berry Lake consists of doubly graded subaqueous pyroclastic flow deposits (see Figure 5), massive coarse and medium clast sized debris flow deposits, thickly laminated tuffs and pumiceous lapilli-tuff fall-out deposits and massive tuffs (see Figures 18 and 19, Stop 9 and Figure 20, Stop 11). Tuffs deposited by ash flow mechanisms are interbedded with these lithologies. Bedding planes between all units are generally sharp to distinct. The deposits are compositionally homolitic to bimodal, matrix supported and show normal and reverse grading. Pumice is a common constituent in the finer deposits. Figure 10 stage B outlines the mechanism of deposition for this portion of the proximal-distal deposition facies which is in a shallow basin within the more massive units.

Ash Flow Tuff Deposits (Stop 10)

Ash flow tuff occurs as discontinuous units within the proximal-distal deposition facies and as a thick continuous unit between the proximal-distal and proximal deposition facies (Figure 9). The ash flow tuff deposits are massive, broken and euhedral crystal supported quartz-feldspar tuff to feldspar tuff. Distinct lithic clast rich zones of the base or pumiceous upper parts were not observed in the thinner units. Clastic zones were noted in the thicker unit indicating the possibility of several events. These clastic zones, discontinuous along strike may represent the base of ash flow deposits. The lack of distinct beds may be accounted for by rapid accumulation of several ash flows cooling as one unit. Minor lithic lapilli were noted throughout the thick cooling unit. These ash flow tuffs

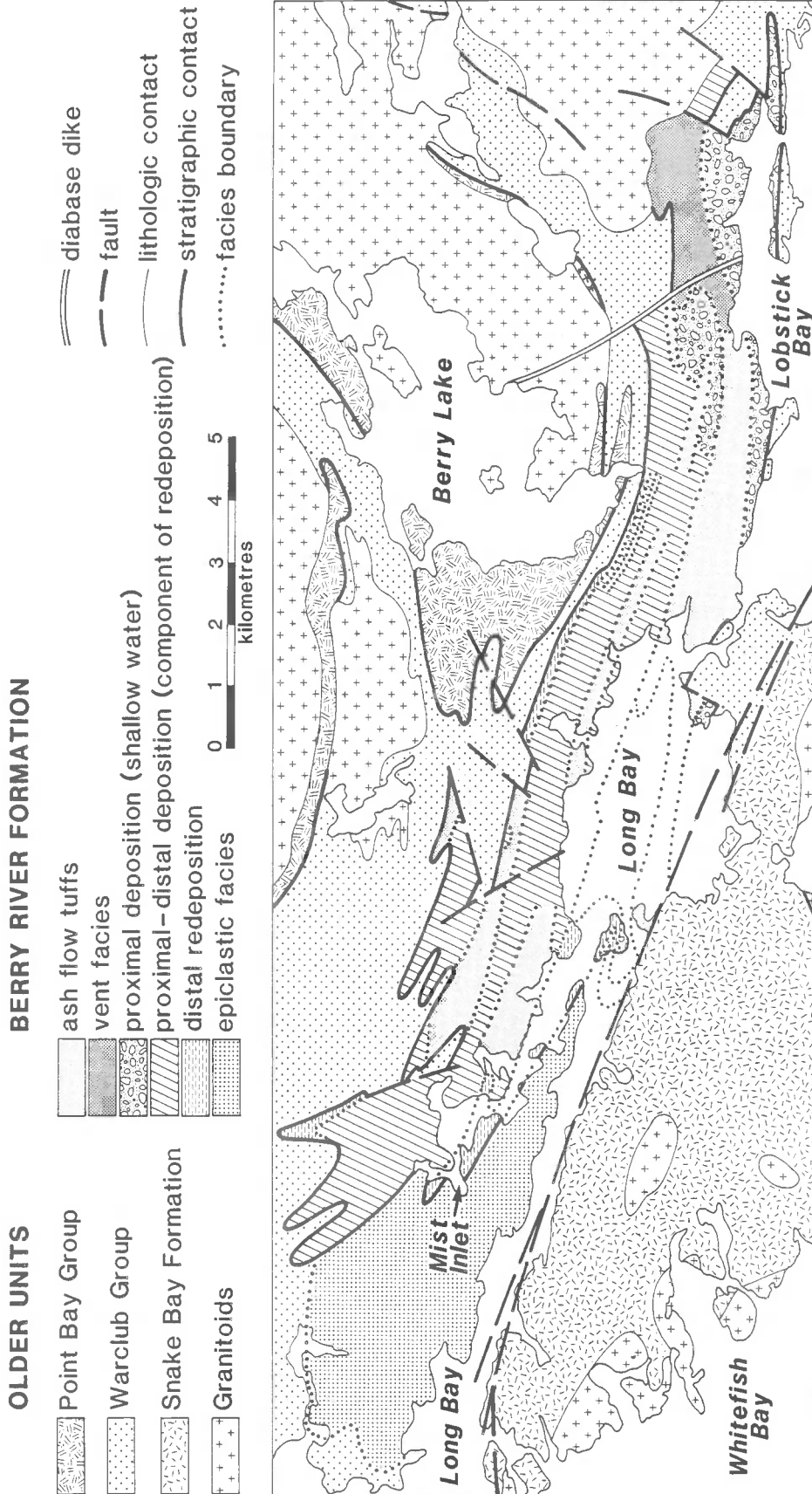


Figure 9: Schematic diagram of the facies distribution of the Berry River formation. Vent facies is a subvolcanic quartz-feldspar porphyry intrusive into the base of a volcano. Proximal deposition facies is composed of coarse homolithic pyroclastics. Proximal-distal deposition facies contains a variety of pyroclastic deposit types. Distal redeposited and epiclastic facies are derived from the previous facies.

were deposited during stage A, B and C, Figure 10.

Proximal Deposition Facies (Stops 3, 4, 7 and 8)

The proximal deposition facies occurs at the east end of the Berry River formation north of Lobstick Bay and in the central part of the formation within Long Bay (figure 9). It is compositionally homolithic pyroclastic breccia to tuff breccia with mostly quartz-feldspar porphyry clasts and the matrix is quartz-feldspar tuff. The facies represents a proximal environment and direct deposition by volcanic processes.

This facies is composed of matrix supported, thickly-bedded, lapillistone to pyroclastic breccia deposited by the block and ash flow process (Sparks, et al. 1973). Bedding planes range from gradational to sharp indicating rapid accumulation. Thin tuff beds separate some block and ash flow deposits and grain size gradations occur within these beds. These thin tuff interbeds are the surge deposit portion of block and ash flow subunits. Thin tuffs overlying the main body of the block and ash flow deposits exhibit grain gradation, suggesting they are products of cloud surge deposition rather than ground surge (stop 3). Primary features such as gas escape structures are found within block and ash flows (stop 7). These features are limited to the top of cooling units.

Figure 10 stages C and D outline the depositional environment envisaged for this facies.

Vent Facies

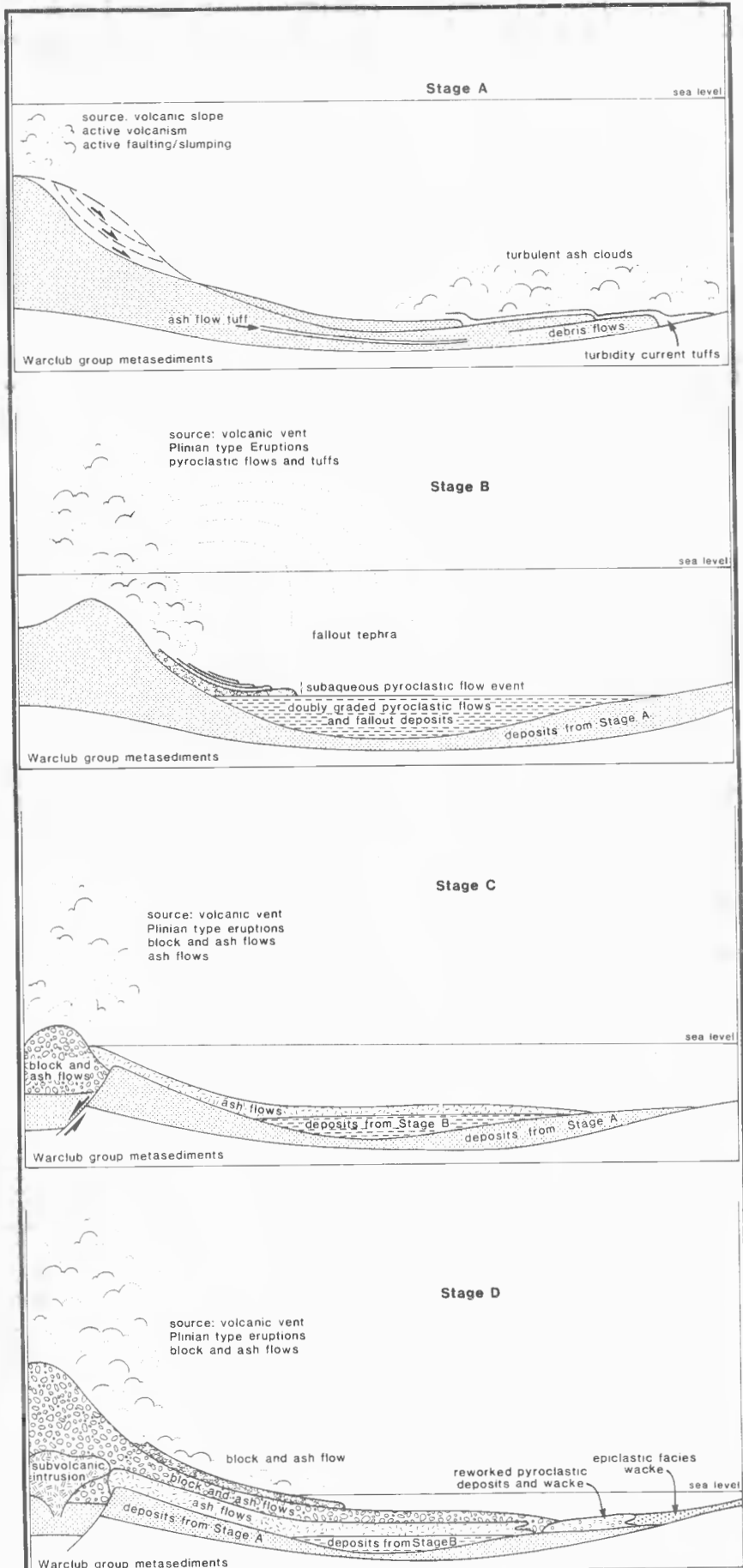
Vent facies rocks are restricted to the eastern end of the Berry River formation (figure 9). The vent facies is represented by a subvolcanic quartz feldspar porphyry intrusion cutting the proximal deposition facies. Portions of the porphyry are massive, while others are subtly clastic or brecciated. The porphyry is similar in petrography to the clasts of the proximal deposition facies. Rafts and xenoliths of the proximal unit are found within the vent facies intrusive.

The similarity between clasts of the proximal deposition facies and the porphyry as well as the presence of rafts and xenoliths suggests that the porphyry was emplaced into the lower part of a growing volcanic edifice and may have been the magma chamber feeding the block and ash flows overlying the thick ash flow tuff (Figure 10, Stage D).

Distal Redeposited Facies (Stops 6, 9 and 11)

Rocks of the distal redeposited facies occur in two stratigraphic positions. They are found at the base of the Berry River formation south-southwest of Berry Lake and north of Mist Inlet and at the top of the formation in Mist Inlet (Figure 9). The distal redeposited facies exhibits primary structures and rock associations (Table 3) typical of a distal environment. Evidence for re-working is shown on Table 1.

South-southeast of Berry Lake the distal redeposited facies has varied thickness (see Figure 18, Stop 9 and Figure 20, stop 11) and consists of moderately sorted, thin to thick bedded, cross bedded, graded reworked tuffs and lapilli tuff. Evidence



- Figure 10: Proposed model for the development of the Berry River formation.
- Stage A: Proximal - Distal Deposition Facies: Debris flows and tuffs.
A volcanic edifice is constructed by explosive volcanism. Building and oversteepening of the cone results in slumping of coarse pyroclastic material into a basin in the underlying sediments of the Warclub group. Interbedded tuffs in this sequence are deposited from fine ash clouds elutriated from the turbulent mass flows and from less violent eruptions. Although these thickly bedded deposits thin laterally, they do not appear to grade laterally.
These deposits are best exposed at the east and west ends of the Berry River formation.
- Stage B: Proximal - Distal Deposition Facies: Subaqueous Pyroclastic Flow Deposit and Fall-Out Deposits
The growing edifice approaches sea level. The eruption pattern changes from single, violent explosive events to a relatively continuous series of explosions that wane in energy. These eruptions are similar to those described by Fiske and Matsuda (1964) producing doubly graded subaqueous pyroclastic flow deposits which fill a shallow basin in the underlying massive deposits. The eruptive column becomes subaerial and fine tephra falls out through air and water, producing thin to thick bedded tuff and pumiceous lapilli tuff. Thin beds of ash flow tuff and lapilli tuff were deposited at this time. Slumping from the edifice also produced massive, medium to coarse debris flow deposits. These deposits are confined to the central portion of the Berry River formation and are best exposed between Highway 71 east to the Berry River.
- Stage C: Proximal Deposition Facies: Block and Ash flow Deposits and Ash Flow Tuffs
The voluminous eruptions producing the proximal-distal deposition facies rocks deflate the magma chamber resulting in local volcano-tectonic subsidence. Homolithic, massive block and ash flow deposits were deposited within the depression. A sequence of thick, massive ash flow tuffs, forming a cooling unit were erupted, perhaps subaerially, from the vent. This depression later becomes the center of renewed volcanism in a shallow water environment.
- Stage D: Proximal Deposition Facies and Redeposited Facies
While explosive eruptions continue to build the edifice a subvolcanic quartz-feldspar porphyry body intrudes the base.
The eruptions produce subaerial block and ash flow deposits that prograde laterally into a shallow water environment.
Surge deposits are recognized between block and ash flow deposits. These proximal deposits extend a fair distance from the vent as the eruptions were subaerial and more violent.
Reworking of these deposits and underlying units resulted in production of the coarse distal redeposited facies rocks and the lateral epiclastic facies wackes.
Erosion and degradation of the subaerial edifice produce the overlying wacke of the Warclub group.

for reworking in this portion of the facies include the moderate sorting and presence of subrounded quartz and feldspar crystals and lithic clasts. Along the basal contact of this facies thin beds show evidence for prograding from both east and west (see Figure 18, stop 9).

North of Mist Inlet at the top of the Berry River formation (Figure 9) distal redeposited facies rocks consist of interbedded, heterolithic, graded tuff breccia to tuff. The presence of volcanic clasts found in the Berry River formation indicates that this facies is associated with the prograding Berry River formation.

Evidence for reworking in the Mist Inlet area includes: sub-round to round clasts, poor to moderate sorting, normal grading, heterolithic nature and presence of wacke interbeds. This tuff to pyroclastic breccia unit is heterolithic, poor to well sorted and graded with subrounded clasts. Wacke beds are interbedded with these reworked pyroclastic rocks (see Figure 16, stop 6).

Figure 10 stage D outlines the model for the deposition of these rocks.

Epiclastic Facies

Quartz-feldspar wackes interdigitate with and overlies proximal-distal deposition facies rocks in the western portion of the Berry River formation (Figure 9) west of Mist Inlet. The wackes are thick to very thick bedded, graded (AE Bouma sequence) and moderately sorted. They resulted from both reworking of pyroclastic deposits and from the distal turbiditic deposition of block and ash flows (Figure 10, stage D).

SUMMARY

The Berry River formation has been placed in a facies context and as such a stratigraphy has been developed. During the field trip we will examine the evidence for the proximal-distal deposition, proximal deposition, ash flow tuff and distal redeposition facies.

ACKNOWLEDGEMENTS

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FIELD TRIP GUIDE

The field trip is designed to be completed in one day with 11 stops; 4 on Lake of the Woods and 7 with road access. Figures 11 and 12 show the locations of the stops.

The 1985 I.L.S.G. trip will commence with stop 3 but if conditions do not permit the use of boats then stops 1 and 2 will be visited. The road guide starts at the Sioux Narrows bridge in the village of Sioux Narrows on Highway 71, 60 km south of Highway 17.

Several points should be made regarding the area:

- 1) Lake of the Woods can be treacherous with respect to navigation and weather. It is advisable to have a navigation map of the lake published by the Canadian Hydrographic Service, Department of Fisheries and the Environment. These maps are readily available from local outfitters.
- 2) The Maybrun Road is a "Forest Access Road" used by heavy trucks. Please park well off the right of way.
- 3) Outcrops are slippery when wet.
- 4) This is WOOD TICK country.

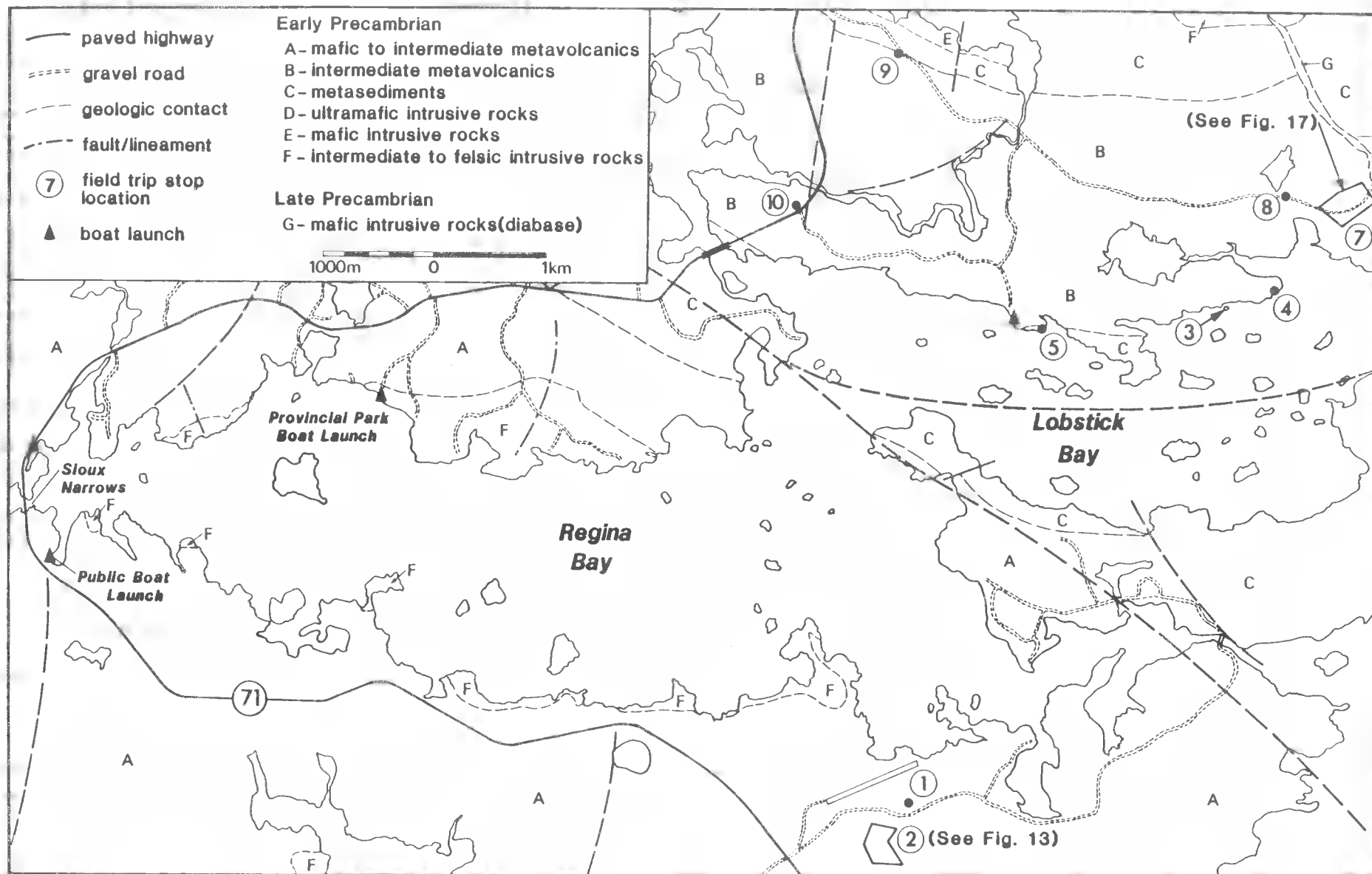


Figure 11: Navigation guide to the field trip stop locations in the eastern part of the field trip area.

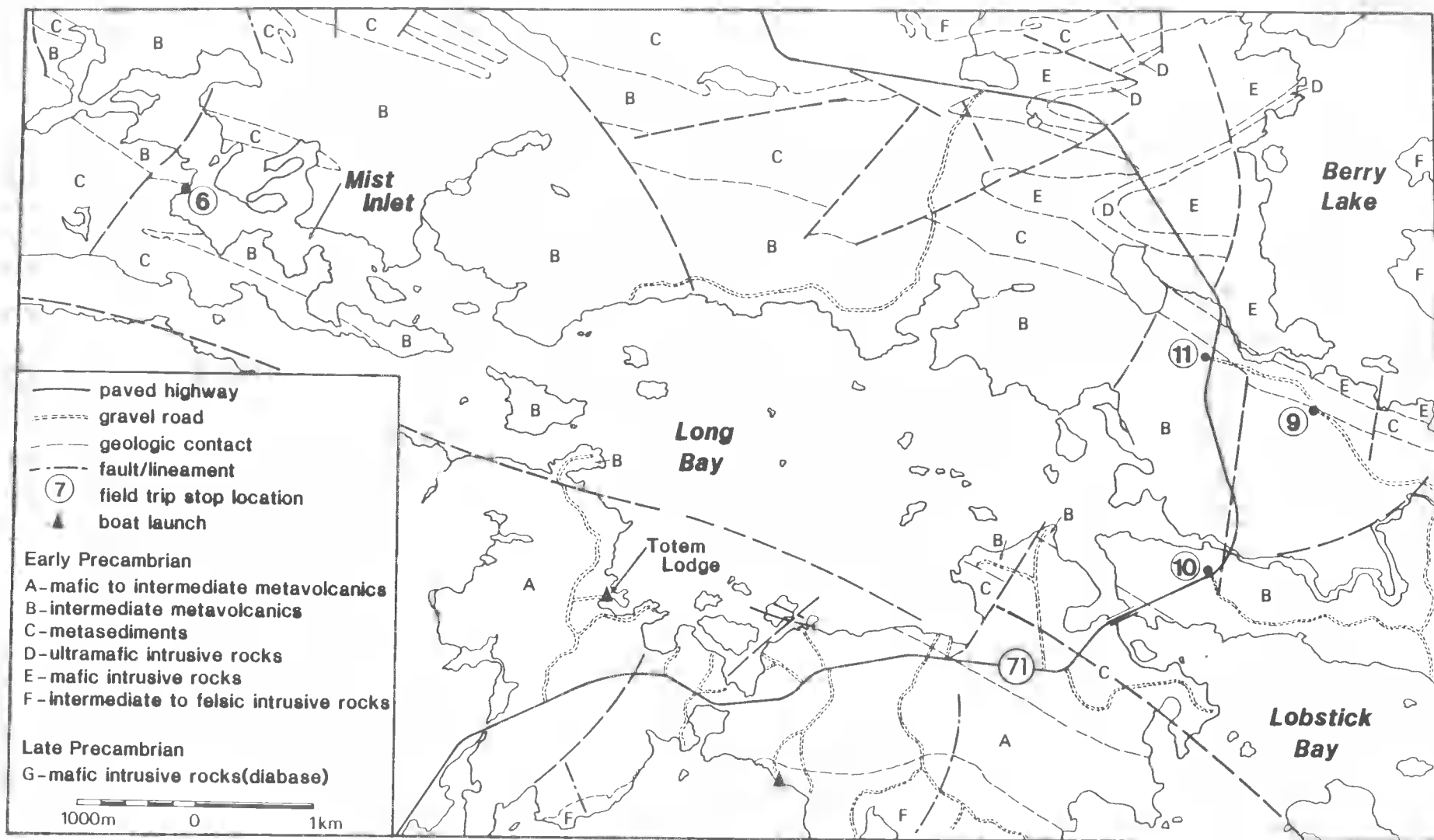


Figure 12: Navigation guide to the field trip stop locations in the western part of the field trip area.

- 0.0 km South end of the Sioux Narrows Bridge on Highway 71 (Figure 11).
Drive south on Highway 71 through Sioux Narrows for 8.2 km to the Gaudry Road.
- 8.2 km Gaudry Road (see Figure 13 for stop locations) Turn left onto the Gaudry Road and drive towards the Whitefish Bay Indian Reservation Village for 0.9 km.
- 9.1 km STOP 1: Park above the gravel pit on the left side of the road and walk to outcrop 'A' (Figure 13).

Formation: Snake Bay formation
Lithology: Mafic metavolcanics
Setting: Strike-southeast, dip-vertical,
stratigraphic top-northeast.
Feature(s): Feeder dike, hyaloclastite, pillows.

The best exposures are in the cut between the two outcrops where the gravel has been removed. The feeder dike, in Figure 14, strikes at 125° and dips vertically. It consists of two symmetrical parts; a marginal zone chilled against the country rock and a central zone chilled against the marginal zone.

The marginal zone is 7 cm thick with a 1.5-2.0 cm thick darker coloured chilled zone against the country rock. Concave up, sub-parallel, fractures 1-7 cm apart occur within this zone and dip 20 degrees into the outcrop. These fractures, 0.2-1 cm thick in the center are carbonate filled and rarely occur in the central zone. The central zone, approximately 4 cm wide, has a 2 mm chilled margin against the marginal zone. The dike can be traced for 26 m to the southeast.

The dike formed in two stages. The first stage froze as a massive fine grained dike. Reactivation forced new magma through the center of the old dike producing fractures in the older marginal zone. The orientation of the fractures indicates that magma movement (in present orientation) had an upward vector perpendicular to the fracture system.

The host of the feeder dike is a brittly fractured, massive, fine-grained, mafic metavolcanic. The carbonate-filled fractures may be early - there is little evidence of penetrative deformation within the outcrop. The exposure on the north side of the gravel pit is fine grained mafic flows with very large ameoboid shaped pillows with 1-2 cm thick selvages and chilled margins that have carbonate filled concave (smile) fractures. A 70 cm thick hyaloclastite to pillow breccia (Carlisle, 1963) zone occurs within the megapillow sequence.

Stratigraphically overlying the megapillows (east end of outcrop) fine-grained, massive flows have thick, brittle fractures filled with soft black (oxidized red)

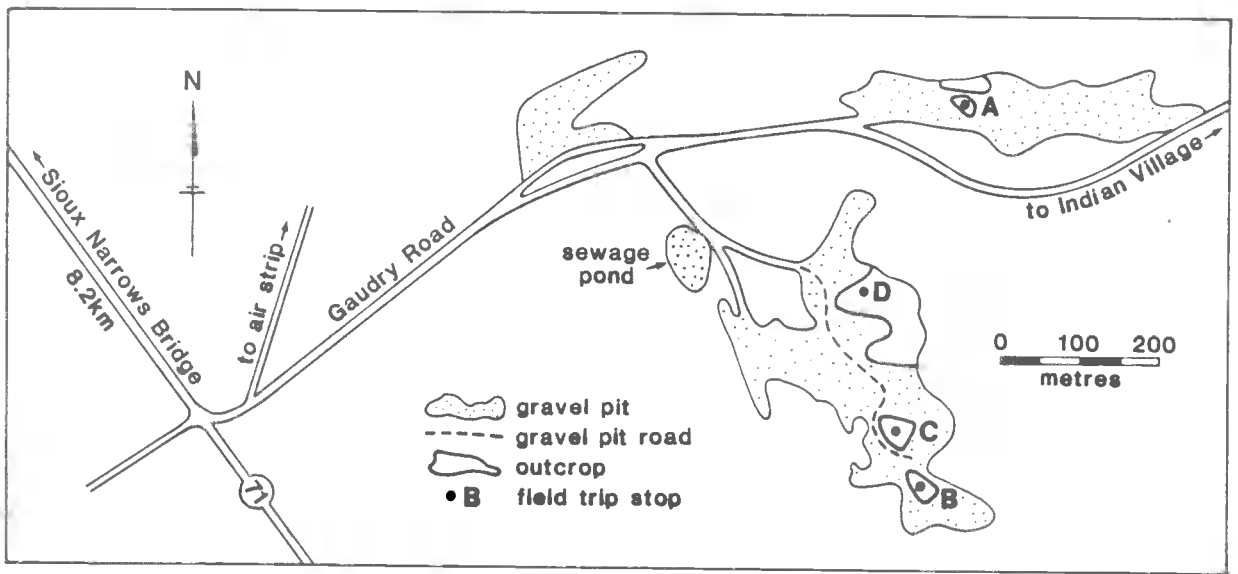


Figure 13: Locations of field trip Stop number 1(A) and Stop numbers 2(B, C and D)

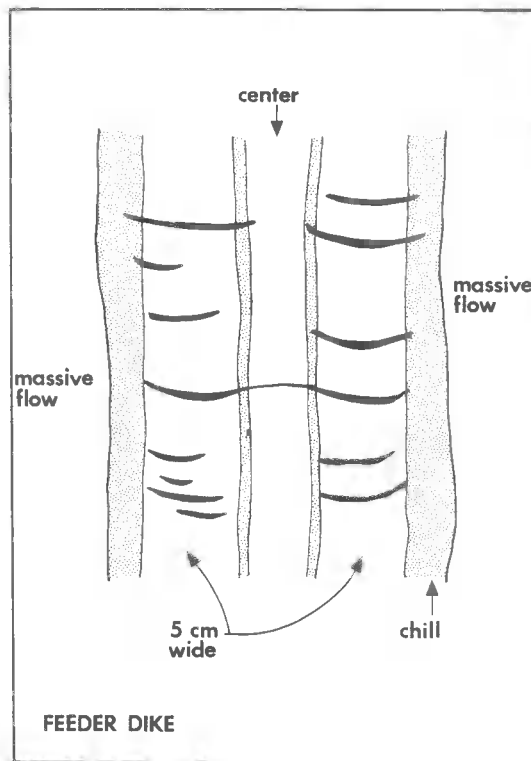


Figure 14: Sketch of the mafic feeder dike at Stop 1 (A on Figure 13). The dike is bilaterally symmetrical forming during two periods of magma emplacement.

hematitic mud and scattered carbonate filled amygdules less than 4 mm in diameter. A sample of the hematitic fracture filling contained 0.05-0.50 percent copper as well as greater than 10 percent iron (30 element quantitative spectrographic analysis by the Geosciences Laboratory, Ontario Geological Survey, Toronto).

- 0.0 km Return to vehicles and drive towards Highway 71 (Figure 13) for 240 m.
- 0.24 km Turn at first left and drive to the "Y" junction about 150 m.
- 0.39 km If road (Figure 13) is blocked, park here taking care not to block access to the sewage pond (follow your nose) or road to gravel pit.

STOP 2: Walk through the gravel pit to outcrop B (Figure 13). This is an ongoing (1985) operation and if the pit is active at the time of your visit please seek permission to enter.

STOP 2B:

Formation: Snake Bay formation

Lithology: Mafic metavolcanics

Setting: Strike-southeast, dip-vertical, stratigraphic top-northeast.

Feature(s): Flow contact, pillows, hyaloclastite, autoclastic breccia

At this stop we will examine an east-south-easterly striking mafic flow that displays a variety of structures. The flow grades from a massive, fine-grained, crystalline base through poorly defined pillows into a hyaloclastite to broken pillow breccia top (Figure 15). Amoeboid pillows rimmed with 1 mm carbonate filled amygdules appear 1.3 m below the flow top. The pillows grade sharply into a variable thickness of hyaloclastite and broken pillow breccia. The pillow breccia contains irregular pillow fragments generally <64 mm in a fine-grained chloritic matrix. The elongation of pillows and embayment of hyaloclastite 30° to bedding represents a flow front advancing to the west-northwest (right to left on figure 15).

The contact with the overlying flow is sharp and irregular. The base is a 2 cm very fine vesiculated zone overlain by a 60 cm to 1 m massive portion grading into coarse hyaloclastite-broken pillow breccia - poorly formed pillows that have been sheared and carbonatized.

STOP 2C: Figure 13 Outcrop "C"

Formation: Snake Bay formation

Lithology: Mafic Metavolcanics

Setting: Strike-southeast, dip-vertical, stratigraphic top - northeast

Feature(s): Pillows, radiating amygdules, epidote alteration.

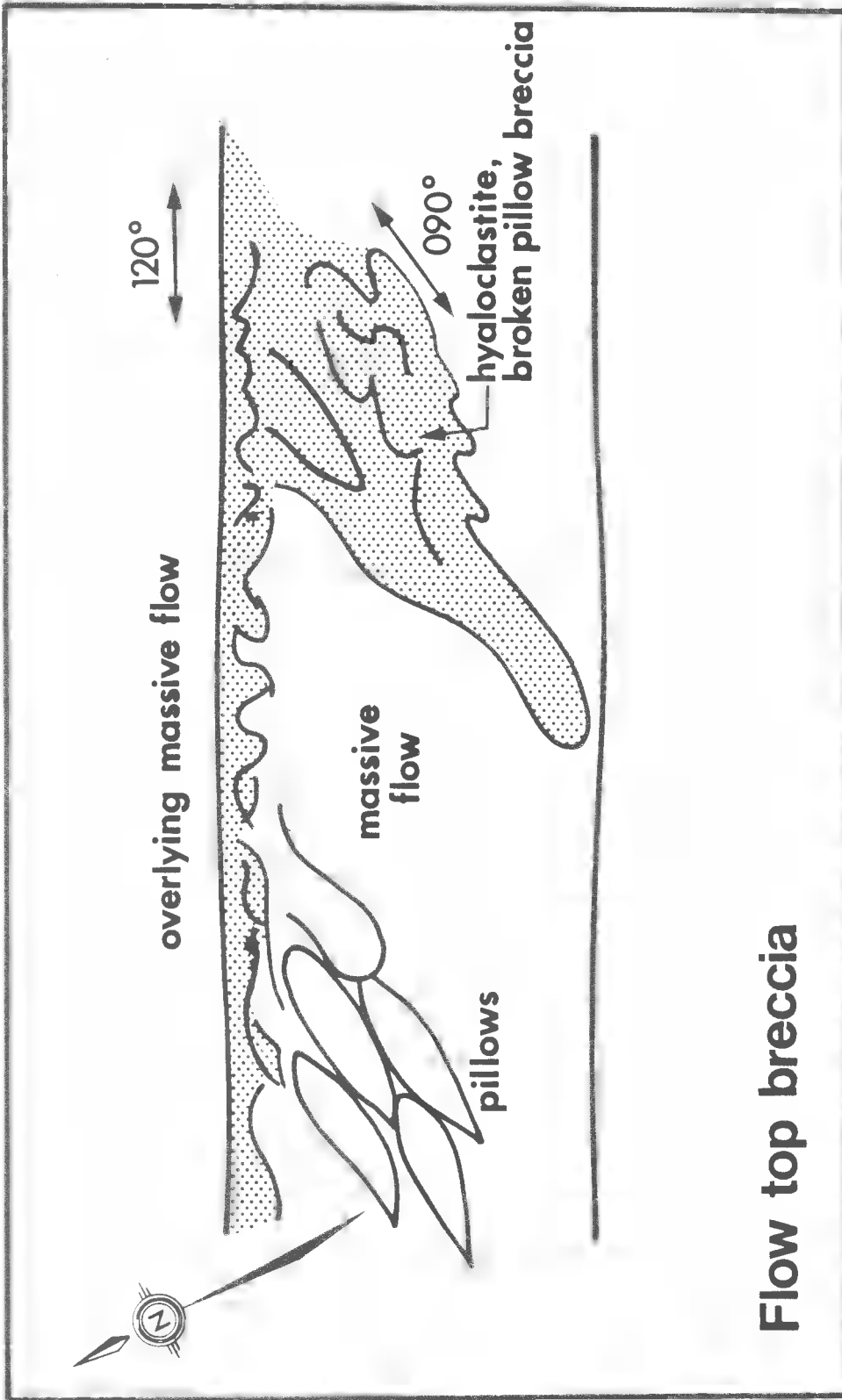


Figure 15: Sketch of the mafic flow top at Stop 2(B on Figure 13). The flow direction is from right to left.

STOP 3: Lobstick Bay. This stop is the larger of two little islands 20 m from the mainland (Figure 11).

Formation: Berry River formation
 Facies: Proximal Deposition
 Setting: Strike-east, dip-vertical,
 stratigraphic top-south
 Feature(s): Block and ash flow; massive portion and
 cloud surge top.

Massive, matrix supported pyroclastic breccia with subangular quartz-feldspar porphyry clasts in quartz-feldspar tuff matrix have poorly defined contacts. Essential porphyry clasts are 10-30 cm and minor accidental mafic clasts are smaller. The fine tuff matrix increases slightly in abundance toward the top. A 21 cm thick quartz-feldspar tuff bed occurs on the south side of the island between two pyroclastic breccia units (Photo 2). The base of the tuff is medium to fine grained and has pronounced laminae.

On the basis of lack of bedding in the coarse units, homolithic subangular clasts and poor sorting this sequence is interpreted as a block and ash flow. The thin tuff interbed based on the association with the block and ash flows and the gradational contact with the underlying breccia is interpreted to be a cloud surge deposit.

Overlying the tuff with a sharp but irregular contact is a similar block and ash deposit.

STOP 4: 400 m east of Stop 3 (Figure 11). Stop on a long outcrop sloping steeply inland. This outcrop has moss on it and can be very slippery.

Formation: Berry River formation
 Facies: Proximal Deposition
 Setting: Strike-east, dip-vertical,
 stratigraphic top-south
 Feature: Pumice in ash flow tuff

Irregular chloritic clasts of a variety of sizes (Photo 3) occur near the topographic top of the outcrop. These quartz-feldspar phyrlic clots, interpreted to be pumice, occur in a fine-grained, massive quartz-feldspar tuff. Two pumice bearing horizons occur in zones of silicification marked by thin quartz veinlets. Subtle lithic clasts are seen. The pumice occurs in the upper portion of two tuff beds (units). Alteration to chlorite and silicification may be pre-metamorphic.

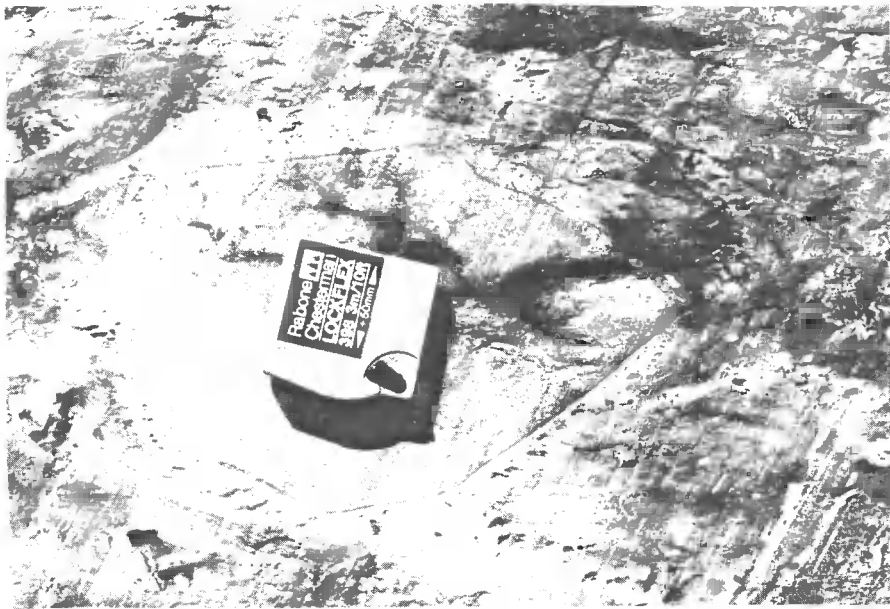


Photo 1: Cored bomb found in a 4 m thick mafic breccia at Stop 2D. A tuff accidental clast is armored with a thin rim of amygdaloidal mafic magma.

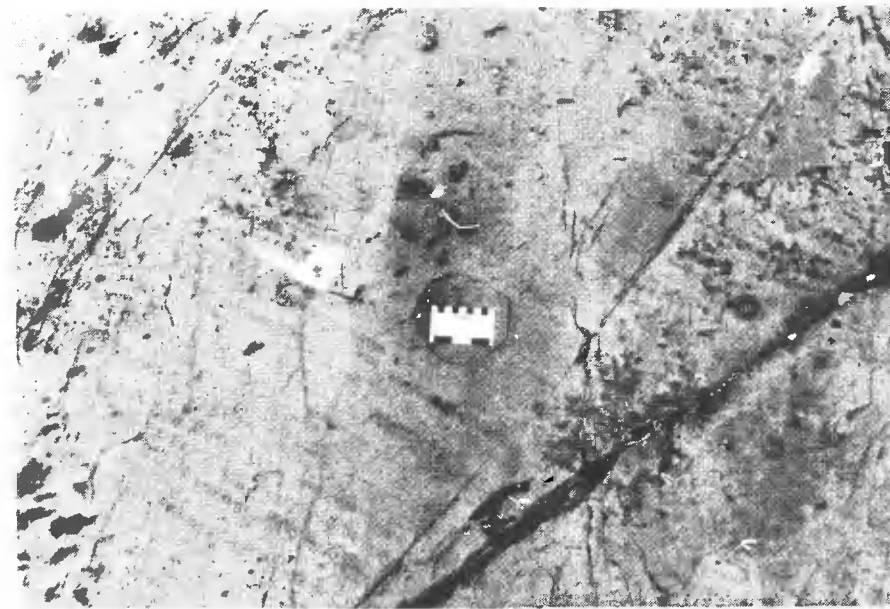


Photo 2: Cloud surge tuff between two pyroclastic breccias at Stop 3. The tuff is the upper portion of a block and ash flow. Stratigraphic top is to the left.



Photo 3: Irregular shaped pumice in a quartz-feldspar ash flow tuff at Stop 4. The pumice now consists of quartz and feldspar phenocrysts in a chloritic matrix.

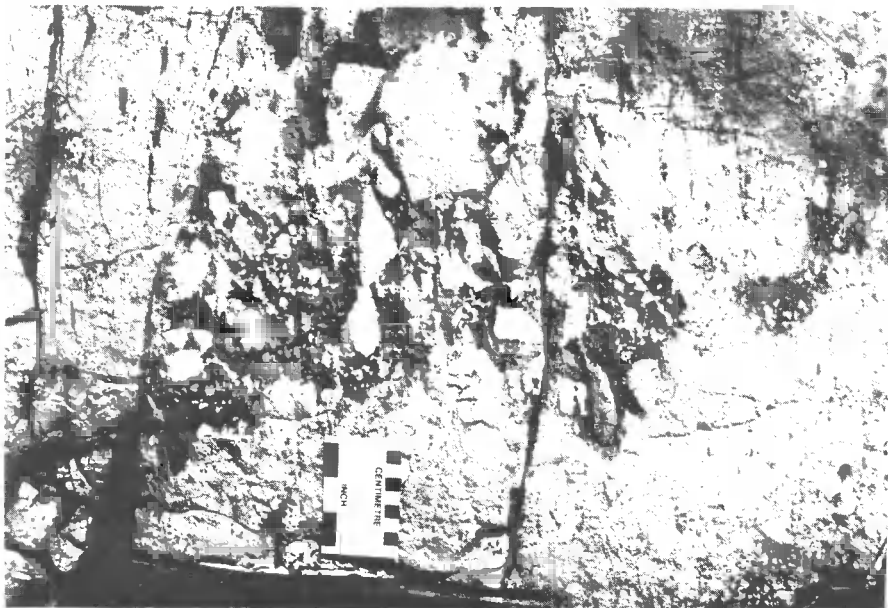


Photo 4: Gas escape structure in the upper portion of a series of block and ash flow deposits at Stop 7. Stratigraphic top is towards the narrow apex of the triangular shaped feature. Gas streaming has altered the matrix to chlorite thus enhancing the clasts.

STOP 5: Approximately 2 km west of Stop 4 on the lakeshore at the west end of Roberts Lodge near the boat dock. This stop at the end of Roberts Road may be reached by vehicle from Highway 71 (Figure 11). Please get permission from the lodge for this stop.

Formation: Berry River formation
 Lithology: Mafic metavolcanics, wacke
 Setting: Strike-east, dip-vertical,
 stratigraphic top-south
 Feature: Top of Berry River formation

The Berry River formation culminated in a 20-100 m thick unit of mafic pillowed flows. The outcrop on the north side of the dock is fine-grained porphyritic (plagioclase) mafic flow with poorly developed pillows. Remnant pillow selvages are seen here. Better developed pillows are found in outcrops among the cabins.

The overlying wackes have 40-90 cm thick graded beds. Grain gradation is well developed with 1 mm subrounded feldspars and quartz grains in a biotite rich matrix grading into 4-5 cm thick argillite top. These wackes, based on the character of the grains, represent reworking and redeposition of the Berry River formation.

STOP 6: This location in Mist Inlet is approximately 11.5 km west of Stop 5 (Figure 12).

Formation: Berry River formation
 Facies: Distal redeposited
 Setting: Strike-east, dip-vertical,
 stratigraphic top-south
 Feature(s): Heterolithic volcaniclastics, wackes

The coarse volcanics (Figure 16) are clast supported beds 1.7 - 4 m thick, predominantly ungraded and contain clasts of feldspar porphyry, intermediate, mafic and felsic metavolcanics and wacke. Size grading is noted in some beds. Clasts, allowing for deformation, are subrounded. The signs of reworking (i.e. compositional difference between matrix and clasts, rounding of clasts, heterogeneity of clast type and the ungraded to poorly graded nature of the unit) suggests this is a debris flow (lahar) (c.f. Fisher and Schmincke, 1984).

The interbedded feldspar tuffs are thinner bedded than the coarse volcaniclastics and contain occasional lithic clasts.

Dark grey, fine grained, argillaceous wacke and lighter grey feldspar and quartz-feldspar wacke units have variable bed thickness.

The heterolithic nature, degree of sorting, and rounding as well as the wacke interbeds suggest the volcaniclastics have been reworked and redeposited. Provenance of the unit was the underlying and lateral parts of the Berry River formation.

MIST INLET: LAKE OF THE WOODS

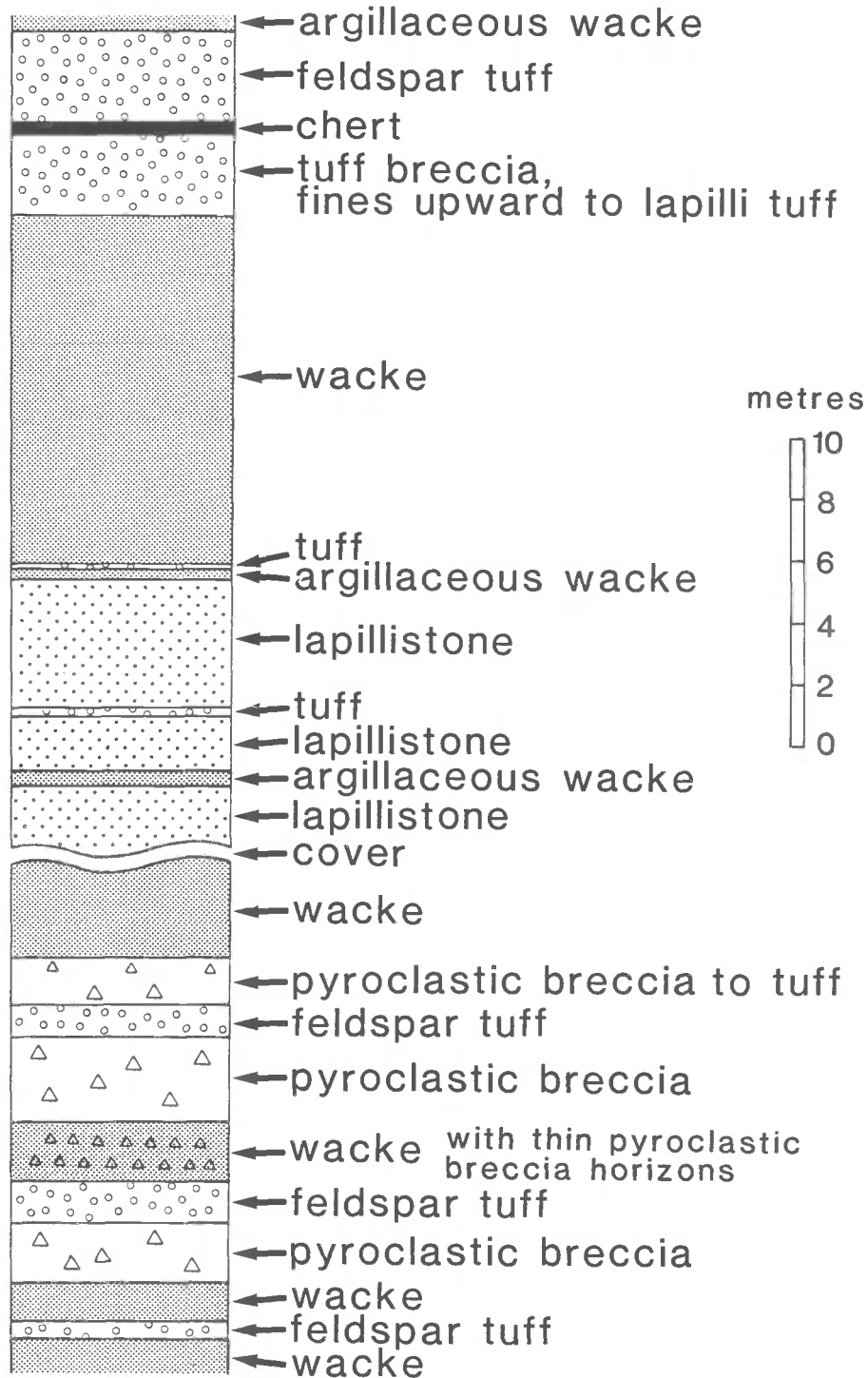


Figure 16: Columnar measured section from a shoreline outcrop in the distal redeposited facies at Stop 6 in Mist Inlet. The base of the column is at the north end of the outcrop area.

Return to vehicles for the remainder of the field trip. The road log will start from the north end of the Sioux Narrows Bridge.

- 0.0 km From the north end of the Sioux Narrows Bridge proceed north on Highway 71 for 10 km to the Maybrun Road. Figures 11 and 12.
- 10 km Turn right (east) on the Maybrun Road for 4.9 km. This mileage is STOP 8 and we will return to it so please remember the location. Continue for a further 0.6 km (Figure 11).
- 15.5 km STOP 7: (Figure 11) is a large area of outcrops on the left (north side of the road) (Figure 17). Please park off the road as it is used by large trucks to haul pulp wood.

Formation: Berry River formation
 Facies: Proximal deposition
 Setting: Strike east, dip-vertical,
 stratigraphic tops-south
 Feature(s): Block and ash flows, debris flows, lahars,
 gas escape structures

Figure 17 shows the location of outcrops as a lithologic map and separately the interpretation of the depositional mechanism. The pyroclastics were deposited as block and ash flows, debris flows and lahars. The lahars were deposited in a channel within debris flows (Figure 17).

The units at the west end of Figure 17 are medium to coarse pyroclastics which are homolithic (quartz-feldspar porphyry clasts), matrix supported, poorly sorted and graded. Up to 8-10 percent pumiceous lapilli occur in units at the western end. Evidence bearing on the origin of these units includes: 1) gas escape structures (Photo 4); 2) their homolithic, poorly graded and poorly bedded character and 3) the presence of pumice. The above are consistent with these units being deposited by block and ash-flow mechanisms (c.f. Sparks et al., 1979).

A gas escape structure (Photo 4) is crudely triangular in shape with the apex towards the stratigraphic top. The matrix has been altered to chlorite, due to gas streaming, and the subround clasts enhanced. Other areas of chlorite alteration, though not as well defined as Photo 4, have been interpreted as gas escape structures.

The presence of the gas escape structures and pumice indicates that these block and ash flows were deposited in a shallow water proximal environment.

In the central and eastern part of the outcrop area one observes a transition to finer heterolithic pyroclastics with thinner, better-developed bedding. The

most common clasts are quartz-feldspar porphyry with with minor siltstone clasts in the northern-most outcrop (Figure 17). Subrounded, fine-grained crystalline mafic clasts also occur. In contrast to the block and ash-flows, sorting is good to poor and graded beds are common. The bedding, sorting, and heterolithic nature of these subunits indicate that they were deposited by debris flows (c.f. Fisher and Schmincke, 1984).

Three coarse volcaniclastic units, each with a siltstone top were deposited in a channel within the debris flows (Figure 17). They are unsorted, heterolithic with elongate, subangular clasts. Most clasts are fine- and medium-grained mafic metavolcanics with approximately 5 percent white aphanitic, felsic lapilli throughout. The matrix is a feldspar-hornblende "tuff". The felsic clasts disappear towards the top and the matrix becomes a light brown silty arenite with cobbles of subround quartz-bearing pyroxene gabbro. The unsorted, heterolithic nature of the volcaniclastic and the presence of the siltstone top suggests these subunits are lahars.

The tuffaceous matrix of the southernmost lahar contains rounded quartz-bearing mafic clasts and quartz-feldspar porphyry clasts. This lahar is rusty, weathering with disseminated pyrite and has cross-cutting cherty veinlets. These features suggest hot emplacement and early alteration.

The complexity present at this stop shows that even though a single facies has been assigned many depositional processes have occurred within this one facies.

0.0 km Turn around and proceed west on the Maybrun road to STOP 8 (you remember the spot?) approximately 0.6 km.

0.6 km STOP 8: (Figure 11). Flat outcrop on the north side of the road.

Formation: Berry River formation

Facies: Proximal deposition

Setting: Strike-east, dip-vertical,
stratigraphic top-south

Feature: Pyroclastic breccia with mafic magma matrix

Homolithic pyroclastic breccia with subangular quartz-feldspar porphyry clasts has a matrix of fine quartz-feldspar tuff. Locally mafic magma occurs as discontinuous blebs with chilled margins wrapping around clasts; clasts and tuffaceous matrix have been bleached for about 1 cm adjacent to the mafic magma. The outcrop has been interpreted to have formed by intrusion of mafic magma into unconsolidated felsic pyroclastic breccia.

This interpretation leads to the conclusion that mafic magmatism must have been concomitant with the

felsic magmatism producing the Berry River formation.

Continue west for 2.6 km toward Highway 71 resetting odometer at the Berry River bridge.

0.0 km Berry River bridge. Continue west for 1.3 km

1.3 km STOP 9 (Figure 11 and 12): included are outcrops on both sides of the road. The road is built up and the sides are soft so care must be taken while parking. Watch for logging trucks!

Formation: Berry River formation
 Facies: Proximal-distal and distal redeposited
 Setting: Strike-southeast, dip-vertical
 stratigraphic top-southwest
 Feature(s): Contact with the Warclub group wackes,
 fall-out deposits, subaqueous pyroclastic
 flows, double grading.

The contact of the base of the Berry River formation and wackes of the Warclub group is exposed in 2 outcrops 30 m apart on the north side of the road. Measured sections of these exposures are shown in the center and right columns of Figure 18. A thin zone of distally redeposited, reworked tuff separates the wacke from the proximal-distal facies rocks. Note that detailed correlation cannot be made between these two closely spaced outcrops (Figure 18).

Figure 19 (left column) is a generalized section of the outcrop on the south side of the road. A thickly laminated to thin bedded sequence of crystal tuff and pumice-crystal lapilli tuff exhibits poorly defined mutual contacts.

The distinct but poorly defined beds are moderate to well sorted and may have been deposited by continuous fall-out through water.

Overlying the fall-out deposits are interbedded, doubly graded pyroclastic flows, tuffs and wacke. These are best exposed 180 m southeast on both sides of the road. The right column (Figure 19) is a generalized section of this exposure.

At this stop the fall-out deposits underlie the coarser flow deposits.

Drive for 2.2 km to Highway 71.

3.5 km Highway 71

0.0 km Turn left and drive to Roberts Road 1.9 km

1.9 km STOP 10 (Figures 11 and 12). The outcrop is on the northwest side of Highway 71 opposite Roberts Road. This stop is on a curve so please park on Roberts Road.

Formation: Berry River formation
 Facies: Ash flow tuff

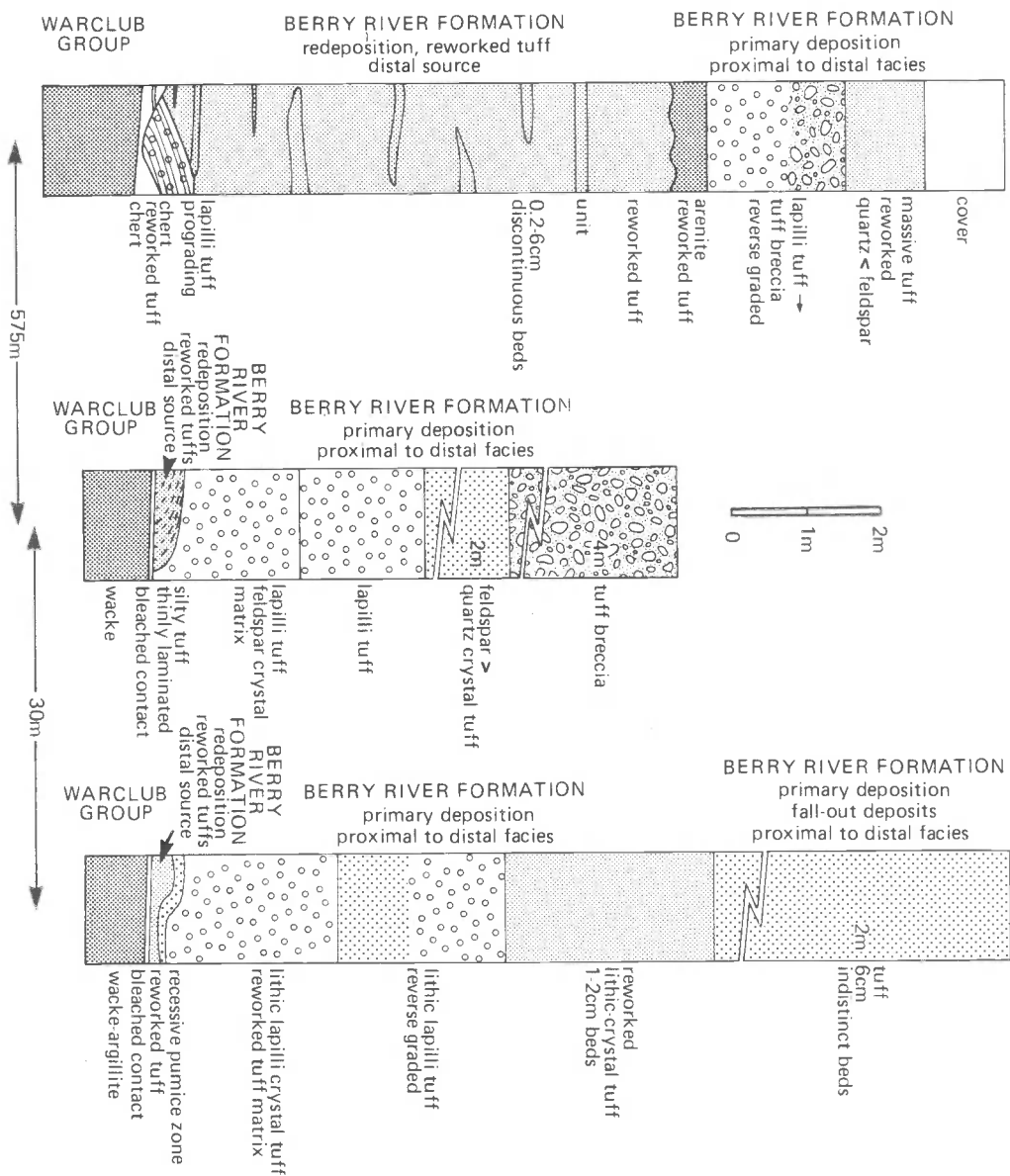


Figure 18:

Three measured sections from separate outcrops found along the Maybrun Road. The contact with Warclub group wackes is exposed in all outcrops. Note the varying thickness of distal redeposited facies and the lack of correlation in the proximal-distal facies volcanics. The centre and right-side measured sections are from Stop 9.

BERRY RIVER FORMATION

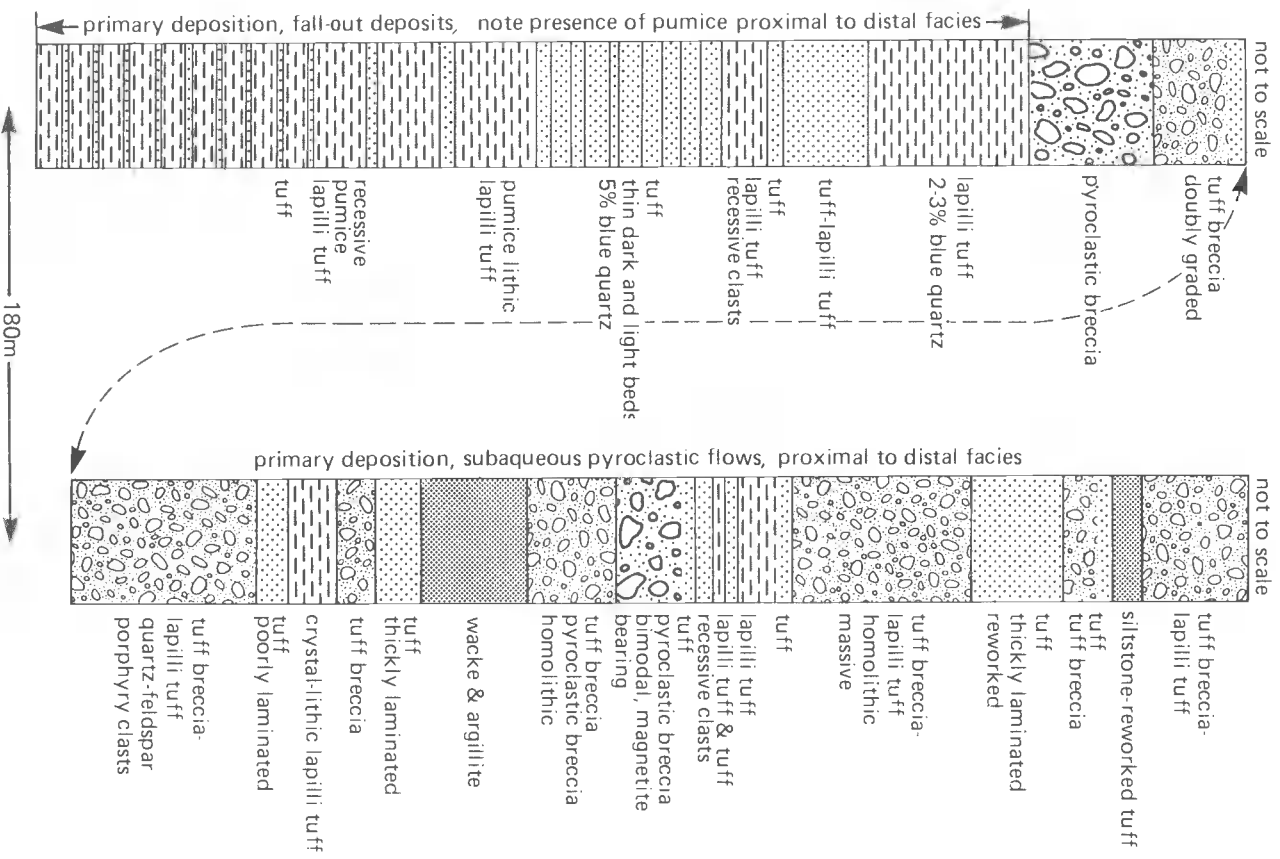


Figure 19:

Two generalized sections from two outcrop areas along the Maybrun road at Stop 9. Rocks in the left column overlie the sections on the center and right side of Figure 18 and the right column stratigraphically overlie rocks described in the left column. Fallout deposits progress up into a mixed assemblage of pyroclastic flow deposits, tuffs and wackes.

Setting: Strike-east-southeast, dip-vertical
 Feature(s): Crystal (quartz-feldspar) tuff, lithic clasts

Anhedra of quartz vary in size and proportion across the outcrop, whereas euhedral to subhedral feldspar phenocrysts do not vary stratigraphically. The phenocrysts are supported in a very fine-grained quartzofeldspathic relict ash matrix. Minor subrounded lithic clasts are found throughout. This stop is just north of the Pipestone-Cameron Fault and much of the "clastic" texture is structural. The variation of phenocryst ratios, tuffaceous character and clastic horizons indicate that the unit was deposited by the ash flow mechanism.

There has been much debate (not quite physical) as to what this outcrop represents. What do you think?

0.0 km Drive north on Highway 71 back to the Maybrun Road 1.9 km.

1.9 km STOP 11 (Figure 12). This is a spectacular outcrop on the west side of the highway opposite Maybrun Road.

Formation: Berry River formation
 Facies: Distal redeposited; proximal-distal deposition
 Setting: Strike-southeast, dip-vertical, stratigraphic tops folded at the north end and south at the south end.
 Feature(s): Reworked tuffs, double grading, reverse and normal grading

Figure 20 is a measured section starting at the small outcrops in the ditch just north of the main exposure.

The base of the section is folded wackes and pyroclastics of the Warclub group.

The distal redeposited facies consists of folded, bedded, reworked tuff and lapilli tuff. The reworked pyroclastics have subround grains and a higher biotite content in the matrix than the non-reworked material. This higher biotite content gives the weathered surface a brownish-buff colour.

Three depositional events not seen at Stop 9 are present here. The first unit (Figure 20) above the reworked tuffs is massively bedded feldspar and quartz-feldspar crystal tuffs. Based upon biotite content and angularity of phenoclasts these are not reworked.

The second depositional event (Figure 20) is the deposition of 5 subunits of doubly graded, subaqueous pyroclastic flows. Each have a coarse clastic base fining south to tuff and/or laminated tuff.

The third event (Figure 20) was the deposition of

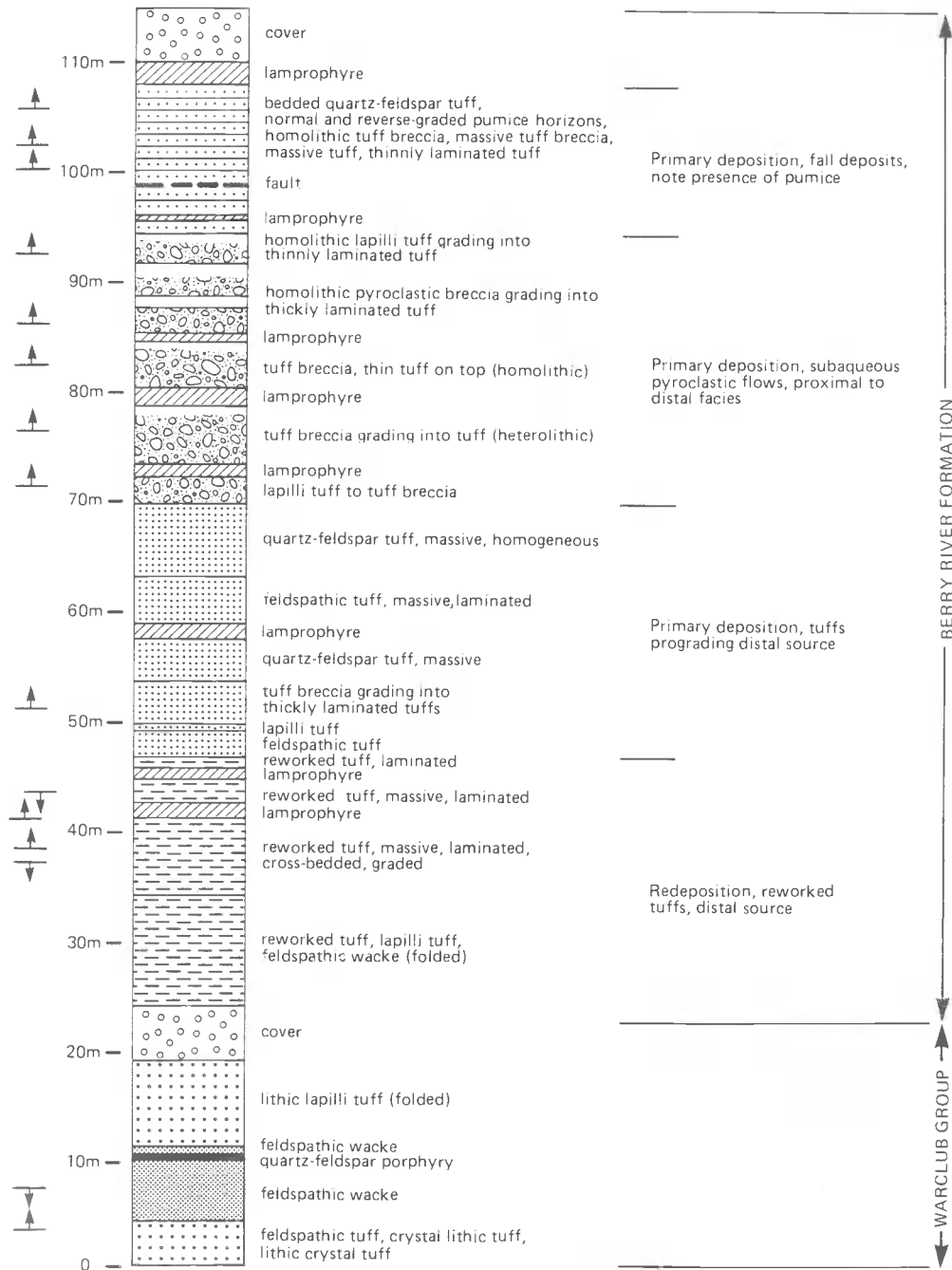


Figure 20: A measured section from Stop 11 on Highway 71 opposite the Maybrun Road. The basal portion of the section has been folded and the rest of the section faces south. The base of the measured section includes the small outcrops exposed in the ditch north of the main exposure. This section includes Warclub group; reworked tuffs in the distal redeposition facies; tuffs, subaqueous pyroclastic flow and fallout deposits in the proximal-distal facies.

thin bedded fall out tuffs and lapilli tuff and tuff breccia deposited by the pyroclastic flow mechanism. The fall-out deposits have normal and reverse graded pumice and normal graded tuffs.

Here fall out deposits overlies subaqueous pyroclastic flows whereas at Stop 9 they underlie them. This is indicative of the difficulty in attempting pyroclastic stratigraphy in the Archean.

TIME FOR A BEER

References

- Ayres, L.D., 1977. Importance of stratigraphy in Early Precambrian Volcanic Terranes: Cyclic Volcanism at Setting Net Lake, Northwestern Ontario; p.243-264 in Volcanic Regimes in Canada, edited by W.R.A. Baragar, L.C. Coleman, J.M. Hall, The Geological Association of Canada Special Paper Number 16.
- Beakhouse, G.P., 1977. A Subdivision of the Western English River Subprovince; Canadian Journal of Earth Science, Volume 14, p.1481-1489.
- Blackburn, C.E., 1981. Kenora-Fort Frances, Geological Compilation Series, Kenora and Rainy River Districts; Ontario Geological Survey Map 2443, 4 miles to the inch.
- Carlisle, D., 1963. Pillow Breccias and their Aquagene Tuffs. Quadra Island, British Columbia; Journal of Geology, Volume 71, p.48-71.
- Davis, D.W., and Edwards, G.R., 1982. Zircon U-Pb Ages from the Kakagi Lake Area, Wabigoon Subprovince, Northwestern Ontario; Canadian Journal of Earth Science, Volume 19, p.1235-1245.
- Davis, D.W., Blackburn, C.E., and Krogh, T.E., 1982. Zircon U-Pb Ages from the Wabigoon-Manitou Lakes Region, Wabigoon Subprovince, Northwest Ontario; Canadian Journal of Earth Science, Volume 19; p.254-266.
- Davis, D.W. and Trowell, N.F., 1982. U-Pb Zircon Ages from the Eastern Savant Lake - Crow Lake Metavolcanic-Metasedimentary Belt, Northwest Ontario; Canadian Journal of Earth Science, Volume 19, p.868-877.
- Easton, R.M., 1984. Reconstruction of Precambrian Volcanoes - A Review of the Canadian Literature; P164-212 in Growth and Evolution of Volcanic Edifices - with Implications for Precambrian Volcanoes edited by R.M. Easton and M. Gaiswinkler Easton, Geological Association of Canada Short Course Notes, Volume 4, 212p.
- Easton, R.M. and Johns, G.W., in press. Volcanology and Mineral Exploration: The Application of Physical Volcanology and Facies Studies; Ontario Geological Survey Miscellaneous Paper.
- Fisher, R.V., 1966. Rocks Composed of Volcanic Fragments and their Classification; Earth Science Reviews, Volume 1, p.287-298.
- Fisher, R.V., 1982. Pyroclastic Flows; in Pyroclastic Volcanism, L.D. Ayres, editor, Geological Association of Canada Short Course Notes, Volume 2, p.111-131.
- Fisher, R.V. and Schmincke, H.U., 1984. Pyroclastic Rocks; Springer-Verlag, New York, 528p.
- Fiske, R.S. and Matsuda, T., 1964. Submarine Equivalents of Ash Flows in the Tokiwa Formation, Japan; American Journal of Science, Volume 262, p.76-106.
- Irvine, T.N. and Baragar, W.R.A., 1971. A Guide to the Chemical Classification of the Common Volcanic Rocks; Canadian Journal of Earth Sciences, Volume 8, p.523-548.
- Jensen, L.S., 1976. A New Cation Plot for Classifying Subalkalic Rocks; Ontario Division of Mines, Misc. Paper 66, 22p.

- Johns, G.W., in prep. Geology of the Long Bay-Lobstick Bay Area, Lake of the Woods, District of Kenora; Ontario Geological Survey Open File Report.
- Johns, G.W., 1984. Kakagi Lake-Rowan Lake Regional Geology; p.25-32 in Summary of Field Work, 1984, Ontario Geological Survey, edited by John Wood, Owen L. White, R.B. Barlow, and A.C. Colvine, Ontario Geological Survey Miscellaneous Paper 119, 309p.
- Johns, G.W., Good, D.J., and Davison, J.G., 1984. Precambrian Geology of the Long Bay-Lobstick Bay Area, Eastern Part, Kenora District; Ontario Geological Survey Map P.2595, Geological Series - Preliminary Map, scale 1:15,840 or 1 inch to 1/4 mile. Geology 1982, 1983.
- Johns, G.W., Thurston, P.C. and Easton, R.M., 1983. Physical Volcanology: 2 Facies Models (Abstract); in Geoscience Research Seminar, December 6-7, 1983, Abstracts, Ontario Geological Survey, 22p.
- Johns, G.W. and Davison, J.G., 1983. Precambrian Geology of the Long Bay-Lobstick Bay Area, Western Part, Kenora District; Ontario Geological Survey Map P.2594, Geological Series - Preliminary Map, scale 1:15,840 or 1 inch to 1/4 mile. Geology 1982.
- Johns, G.W. and Richey, Scott, 1982. Precambrian Geology of the MacQuarrie Township Area, Kenora District; Ontario Geological Survey Map P.2498, Geological Series - Preliminary Map. Scale 1:15,840 or 1 inch to 1/4 mile. Geology 1981.
- Krogh, T.E. and Davis, G.L., 1971. Zircon U-Pb Ages of Archean Metavolcanics in the Canadian Shield; Carnegie Institute Yearbook, Vol.70, p.241-242.
- Krogh, T.E., Davis, D.W. and Corfu, F., 1984. Implications of Precise U-Pb Dating for the Geological Evolution of the Superior Province; Geological Association of Canada Abstracts, Vol.9, p.79.
- Lawson, A.C., 1885. Report on the Geology of the Lake of the Woods Region with Special Reference to the Keewatin (Huronian?) Belt of Archean Rocks; Geological and Natural History Survey of Canada, Annual Report, 1885, Vol.1, Pt.C., 151p.
- Lichtblau, A.P. and Dimroth, E., 1980. Stratigraphy and Facies at the South Margin of the Archean Noranda Caldera, Noranda, Quebec. In Current Research Pt.A, Geological Survey of Canada, Paper 80-1A, p.69-79.
- Ross, Clarence, S. and Smith, Robert L., 1961. Ash Flow Tuffs: Their Origin, Geological Relations and Identification, United States Geological Survey Professional Paper 366, 81p.
- Schmid, R., 1981. Descriptive Nomenclature and Classification of Pyroclastic Deposits and Fragments: Recommendations of the IUGS Subcommittee on the Systematics of Igneous Rocks; Geology, Volume 9, p.41-43.
- Sparks, R.S.J., Self, S., and Walker, G.P.L., 1973. The Ground Surge Deposit: A Third Type of Pyroclastic Rock; Nature, Volume 241, p.63-64.
- Stockwell, C.H., McGlynn, J.C., Emslie, R.F., Sanford, B.V.,

- Norris, A.W., Donaldson, J.A., Fahrig, W.F. and Currie, K.L. 1970. Geology of the Canadian Shield, in Geology and Economic Minerals of Canada, edited by R.J.W. Douglas, Geological Survey of Canada, Economic Geology Report No.1, 838p.
- Tasse, N., Lajoie, J., and Dimroth, E., 1978. The Anatomy and Interpretation of an Archean Volcaniclastic Sequence, Noranda Region, Quebec; Canadian Journal of Earth Science, Volume 15, p.874-888.
- Trowell, N.F., in prep. Geology of the Gibi Lake Area, District of Kenora; Ontario Geological Survey Open File Report.
- Trowell, N.F., Blackburn, C.E., and Edwards, G.R., 1980. Preliminary Synthesis of the Savant Lake - Crow Lake Metavolcanic - Metasedimentary Belt, Northwestern Ontario; and its Bearing upon Mineral Exploration; Ontario Geological Survey, Miscellaneous Paper 89, 30p. Accompanied by Chart A.
- Williams, H., and McBirney, A.R., 1979. Volcanology; Freeman, Cooper and Co., San Francisco 379p.

Granitoid Related
Mineralization in the
Dryden Area

Introductory Discussion and Field Guide
31st Annual Institute on Lake Superior Geology
Kenora, Ontario

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GENERAL GEOLOGY

This field trip examines the diverse granitoid rocks and associated mineralization in the Dryden area. The rocks of this area are transitional from the Wabigoon Subprovince northward to the Southern plutonic domain (Winnipeg River terrane) of the English River Subprovince. All field stops will be made within a boundary zone limited by the Wabigoon Fault on the south and by a line marking the maximum extent northward of abundant supracrustal rocks (Figure 1). The precise position of the subprovince boundary within this zone is debatable and is largely dependent on the lithological, structural or metamorphic criteria used to define it but, for the purposes of this excursion, it is sufficient to note a number of general characteristics of the zone:

- 1) It contains a higher proportion of metasedimentary rocks and derived migmatitic stages (Table 1) than is normal within the main mass of the Southern plutonic domain (Breaks et al., 1978) and the Wabigoon Subprovince to the south. In the Dryden area, the Wabigoon Fault marks the southern extent of these metasedimentary rocks (Zealand Group) which are dominantly of turbiditic origin.
- 2) The proportion of exposed plutonic rocks relative to supracrustal rocks increases northward and northwestward within the zone.
- 3) A steep metamorphic gradient is present within this zone: low grade assemblages in pelitic rocks adjacent to the Wabigoon Fault give way northward over a distance of 7 km to high grade assemblages. Metamorphic pressures are inferred to be relatively low (Bartlett, 1978; Breaks, 1982).
- 4) A suite of gneissic granitoid rocks which represent intensely metamorphosed and deformed polycyclic intrusive phases are characteristic of the main mass of the southern plutonic domain but are absent from the zone under consideration.
- 5) Metavolcanic rocks are particularly abundant in the eastern part of the zone and form parts of two units which extend northeastward out of the area of consideration. The northern unit which extends to Gullwing Lake (Figure 1) forms a portion of the Northern Volcanic Belt (Trowell et al., 1980) which comprises a lower south facing mafic sequence composed largely of magnesian basalts and an upper mixed intermediate to felsic metavolcanic-metasedimentary sequence. The second unit which includes the Brownridge Volcanics (Figure 1) forms part of the Central Volcanic Belt which comprises mafic to intermediate flows with local felsic flows, autoclastic breccias and thin sedimentary intervals (Trowell et al., 1980). A third unit, comprising mixed mafic to felsic metavolcanic rocks occurs in the western part of the zone in the vicinity of Vermilion Bay.

GRANITOID ROCKS

While individual intrusions within the boundary zone commonly display internal compositional variations, it is possible to subdivide the granitoid rocks in a general way into

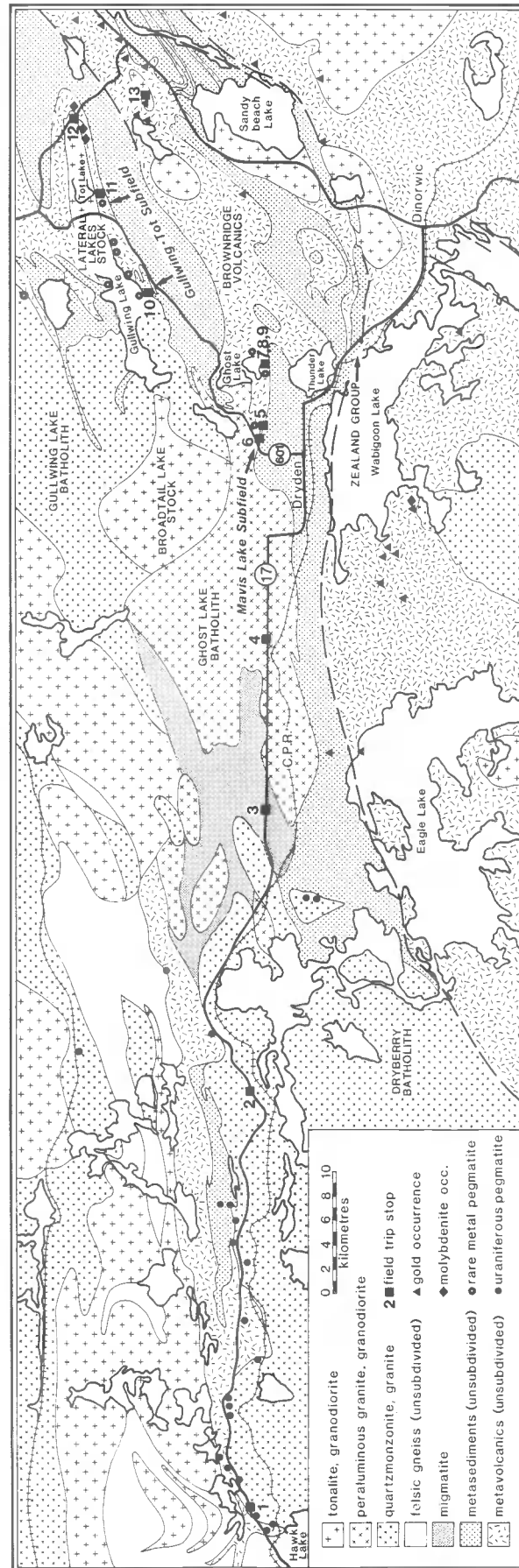


Figure 1: General Geology of the Dryden Area.

Table 1: CLASSIFICATION AND GENERAL FEATURES OF METASEDIMENTARY MIGMATITES ENGLISH RIVER SUBPROVINCE.

MIGMATITIC STAGE	LEUCOSOME: LEUCOSOME & PALEOSOME RATIO	DIAGNOSTIC MIGMATITIC STRUCTURE	GENERAL PROCESSES DOMINANT	GENERAL FIELD FEATURES
Proto-metatexite	<0.1	Locally stromatic Sedimentary bedding or laminae may show preservation	Incipient, 'selective' anatexis. Metamorphic differentiation may be important	Characterized by intercalated fg wacke and mg-cg porphyroblastic pelite components. Podiform and lentic potassic leucosomes exhibit confinement to pelitic horizons. Hydrothermal (qtz. veins) mobilize may be important.
Metatexite	0.1 to 0.6	Stromatic and/or Phlebitic	Local to moderate degree of anatexis	Mobilize developed in situ, possesses biotitic melanosomes along leucosome-paleosome interface. Mobilize (without melanosomes) also commonly injected along prevailing foliation or bedding surface of paleosome.
Inhomogeneous Diatexite	0.6 to 0.9	Schollen structure especially characteristic. Locally schlieritic or nebulitic	Anatexis relatively extensive, mobilize essentially autochthonous	Repletion of disoriented paleosome and metatexite inclusions and melanosome clumps is diagnostic.
Homogeneous Diatexite	>0.9	Homogeneous usually massive. Locally schollen, schlieritic or nebulitic structures apparent	Very advanced fusion; mobilize probably allochthonous	Usually holo-leucocratic and severely seriate. Rarely more mafic variants (CI 15 to 30) may represent cases of complete anatexis effecting resorption of melanosome component into magmatic phase

Abbreviations: fg - fine grained; mg - medium-grained; cg - coarse grained; CI - Colour Index; > - greater than; < - less than

three suites which reflect not only compositional variations but their timing relative to other tectonic events:

- i) a pre- to syn-tectonic sodic suite
- ii) a syn-tectonic peraluminous suite
- iii) a syn- to post-tectonic potassic suite.

The essential features of each suite are described below with reference to the plutonic rock nomenclature of Streckeisen (1976).

Sodic Suite

The phases within these intrusions are dominantly trondhjemite and granodiorite with rare granite. The rocks are quite uniform in composition and commonly possess a metamorphic foliation. The rocks are generally medium grained and contain biotite as the dominant mafic mineral. In the Dryden area this suite is represented by small plutons and sills such as the Goldlund trondhjemite within the Central Volcanic Belt and by stocks such as the Lateral Lake body composed of foliated biotite-bearing granodiorite (McCarter, 1980) and the Broadtail Lake body composed of massive to foliated biotite and hornblende-biotite granodiorite.

Peraluminous Suite

A dominant granitoid element in the Dryden area is the Ghost Lake Batholith. This peraluminous, two-mica granitoid was previously mapped, in part, by Moorhouse (1939) and Satterly (1941) and completely delineated by Breaks et al. (1976) with more recent detailed mapping completed by Breaks (1984). The batholith is 280 km², is elongated and is broadly concordant with east to northeasterly striking structural trends in adjacent rocks. The batholith is distinguishable from virtually all other plutonic masses in the region by virtue of a unique accessory mineral suite including muscovite, cordierite, sillimanite, Mn-garnet, tourmaline and rare dumortierite and beryl.

The main exposed mass of the batholith covers a 40 km long area between Eagle River and Ghost Lake which contains inhomogeneous to homogeneous diatexite composed of coarse grained granite and granodiorite with variable contents of biotite and muscovite. The western two-thirds of this mass probably represents a lower structural level of the intrusion in that the granitoid rocks are less homogeneous, contain less tourmaline and muscovite and form transitional contacts with migmatized metasedimentary rocks displaying assemblages typical of the second sillimanite isograd. The eastern portion of the mass shows characteristics of the upper level of the intrusion as it has an abrupt contact with metasediments, few sedimentary inclusions, abundant muscovite and tourmaline (locally 10 to 20 percent) and potassic pegmatites which typify the cupola zone of batholiths parental to rare element pegmatites (Varlamoff, 1972). Along the southern contact, the batholith lies in abrupt contact with low- to medium-grade metasediments with no appearance of assemblages of the second sillimanite zone in metapelitic layers which militates against in-situ partial melting in this part of the body.

Potassic Suite

This suite is represented in the Dryden area by two large masses, the Dryberry Batholith and the Gullwing Lake Batholith (Figure 1). The phases within these bodies are predominantly quartz monzonites and granites which are commonly massive; homogeneous and medium to coarse grained. Unlike the sodic suite, these rocks are generally discordant with their hosts and dikes and sills of similar composition may be very abundant in adjacent country rocks. The greatest volume of rock in these batholithic complexes characteristically lacks a metamorphic foliation although contact-parallel foliations near the margins related to emplacement, are common features.

ECONOMIC GEOLOGY

Gold Deposits

Gold occurrences in the Dryden area are concentrated in the Central Volcanic Belt in the eastern part of the boundary zone. In a general way, the geological setting of gold mineralization in this area is similar to that in many other gold camps in Superior Province (Colvine et al., 1984). Some of the features of this gold-district are:

- i) On a regional scale, gold is localized at or near a major volcanic-sedimentary interface (Figure 1) though most individual deposits occur within volcanic and granitic rocks.
- ii) A wide variety of deposit morphologies is present including veins and stockworks in volcanic rocks, en-echelon vein arrays in felsic intrusive bodies, carbonate zones in volcanic and sedimentary rocks and silicified shear zones in tuffs and lavas (Blackburn and Janes, 1983).
- iii) Gold mineralization is hosted by structures bounded by two major faults which roughly parallel stratigraphic trends.
- iv) Gold-quartz veins post-date all rock types in the area with the possible exception of some post tectonic granitoid intrusions.
- v) Several occurrences are spatially related to small bodies composed of quartz and feldspar porphyry.

The most significant occurrence of gold in this area is the Goldlund deposit. It comprises en echelon sets of extensional veins which transect a body of trondhjemite which is broadly concordant with adjacent volcanic rocks. The veins appear to be the product of hydraulic fracture of the intrusion and the relationship between gold and the granitoid rock is likely passive and simply the result of brittle deformation of a competent unit. This situation is similar to that at other deposits in Superior Province, most notably at the Sigma Mine, Val d'Or, Quebec (Robert et al., 1983). An important feature of Goldlund is the presence of numerous dikes of quartz and feldspar porphyry which cut the trondhjemite but not the gold-quartz veins. In some cases vein-bearing structures in the trondhjemite extend unmineralized into porphyry dikes. Porphyries of this type are common features of many gold deposits (McIntyre, Hollinger, Dome) but their genetic relationship to gold mineralization is poorly understood.

Molybdenite Deposits

The Lateral Lake granodioritic stock hosts several occurrences of molybdenite at its contact with rocks of the Northern Volcanic Belt. This mineralization has been described by McCarter (1980), Colvine and McCarter (1977) and Page (1984). Molybdenite occurs in quartz and pegmatitic veins which are particularly well developed in aplitic dikes which cut the granodiorite. The molybdenite occurs i) as euhedral grains with K-feldspar and phlogopite in pegmatites, ii) as rare euhedral grains in quartz veins, iii) along vein margins, iv) as isolated grains in quartz-vein stockworks; v) as narrow bands and lenses parallel to foliation in wall-rocks and, vi) along fractures not parallel to foliation in wallrocks (Colvine and McCarter, 1977). Muscovite and pyrite are common accessories to the molybdenite mineralization. Colvine and McCarter (1971) relate the mineralization to a hydrothermal system developed during the later stages of crystallization of the Lateral Lake stock while Page (1984) relates it to post-tectonic emplacement of pegmatites which are unrelated to the Lateral Lake intrusion and controlled by the nearby Kathlyn Lake Fault.

Rare-Metal Pegmatites

Rare-metal pegmatites in the Dryden area were first described by Mulligan (1965) who termed their concentration the "Dryden Pegmatite Field". Regional zonation within this field was first recognized by Breaks (1980), as outlined in Table 2, resulting in a subdivision into two distinct pegmatite groups (Terminology after Cerny, 1982, p.8-9) at Mavis Lake and Gullwing-Tot Lakes separated by a distance of 9.7 km.

The geology of the Mavis Lake Pegmatite Group is dominated by the Ghost Lake Batholith which represents the most plausible source area for the rare-metal pegmatites based upon the following petrologic features and chemical characteristics:

- i) the advanced degree of geochemical fractionation and difference in plutonic units such as GLB4, 5, 6 (Table 3) relative to the main western mass of the batholith which is dominated by plutonic unit GLB-1 (Breaks, in prep.). The former plutonic units tend to exhibit increased levels of muscovite, tourmaline and garnet with sparse to nil biotite in contrast to plutonic unit GLB-1;
- ii) sporadic presence of rare-metal mineralization (i.e. beryl) within some internal pegmatites (Dryden Airport beryl occurrence) and in contact pneumatolytic tourmaline - quartz deposits (Petrunka Tungsten Property at STOP 5);
- iii) contact metasomatic selvages, developed in mafic metavolcanic host rocks, which exhibit anomalous trace level enrichment in Li, Cs, Be, Rb, Sn and F (Petrunka Tungsten Property at STOP 5).

Significant petrochemical zonation within the Ghost Lake Batholith is revealed by certain trace elements, in particular B, Ba, Li, Rb, Sr, Zr and REE (Breaks, in prep.). Trends of increasing geochemical fractionation involving enrichment of B, Be, Ga, Li, Nb, and Rb coupled with depletion of Ba, Sr, Zr, and total REE have been delineated within the batholith from the main

Table 2: Petrological and Geochemical Features of Rare-Metal-Bearing Pegmatites from the Dryden Field.

Area/Occurrence	Pegmatite Structure	Characteristic Pegmatite Mineral Assemblage ¹	Geochemical Association ²	Degree of Late Stage Replacement Albitization
Mavis Lake Pegmatite Group				
Dryden Airport Occurrence, Iseland Township	Unzoned, internal, usually barren potassic pegmatites	Garnet + Muscovite + Tourmaline + Albite + Quartz + Blocky K-feldspar + (Limegreen Beryl)	B-(Be-Cs)	Absent
Taylor Occurrence, Concession VII, Iseland Township	Unzoned potassic pegmatites developed marginal to Ghost Lake Batholith	Beryl + Tourmaline + Albite + Quartz + Graphic K-feldspar + Blocky K-feldspar + (Columbite + Garnet + Muscovite + Green Apatite)	B-Be-(Sn-Nb)Ta-P	Incipient. Small pods of fine-grained, saccharoidal apite (+ Chromium Mica + Quartz + Albite)
Petrunka Occurrence, Brownridge and Iseland Townships	Unzoned, tourmaline-enriched fracture/pillow selvage-controlled replacement in Brownridge mafic metavolcanics	Quartz + Plagioclase + Biotite + Scheelite + Tourmaline + (Pyrite + Holmquistite + Fluorite?)	W-B-P-(Li-Sn-Be)	Not Observed
Fairservice Property, Brownridge Township Main and South Zones	Unzoned to crudely zoned external pegmatites. Contains randomly oriented, green primary spodumene phenocrysts (Type 1 spodumene of Heinrich, 1975)	Beryl + Green Muscovite + Albite + Spodumene + Blocky K-feldspar + Quartz + (Blue Apatite + Garnet + Tourmaline + Tantalite)	Li-(Be-Rb-Ta)Nb-B-Sn	Moderate development of albitization (+ Garnet + Tantalite + White Beryl + Green Muscovite + Quartz + Cleavelandite)
East Zone	Unzoned external pegmatites	Primary mineralogy largely replaced. Remnants of blocky K-feldspar, spodumene, and quartz still discernible	Ta>Nb-Be-B-(Li)	Pervasive albitization. Most of pegmatite replaced by + White Beryl + Tantalite + Green Muscovite + Quartz + Cleavelandite
Gullwing Lake-Tot Lake Pegmatite Group				
Mica Point Gullwing Lake, Drope Township	Crudely zoned, external potassic pegmatite	Muscovite + Biotite + Albite + Quartz + K-feldspar + (Molybdenite + Columbite-Tantalite)	(Nb>Ta-Mo)	Incipient
Gullwing Lake Pegmatite, Webb Township	Complex external pegmatite. Most of dike consists of muscovite-biotite pegmatitic granite. Contains green Type 1 spodumene of Heinrich (1975)	Green Spodumene + Muscovite + Albite + Quartz + Blocky K-feldspar + Albite + Green Muscovite + Quartz + Blocky K-feldspar	Li-(Ta<Nb-Mo-Sn)	Moderate albitization developed in 2 stages: 1. early medium-grained, equigranular + Biotite + Tantalite + Muscovite + Quartz + Albite 2. later, fine-grained Lepidolite-Albite Complex + Tantalite + Beryl + Garnet + Lepidolite + Muscovite + Cleavelandite)
Coates Pegmatite, Webb Township	Unzoned, external potassic pegmatite	Muscovite + Albite + Quartz + Graphic K-feldspar + (Molybdenite + Pyrite + Garnet + Samarskite + Sphalerite + Chalcocopyrite + Beryl)	Mo-(Nb>Ta-Cu-Zn-S-Be-REE)	Incipient replacement by fine-grained albite along edges of K-feldspar. Most molybdenite and samarskite associated with albitized areas
Pegmatite 215 m Southwest of Coates Pegmatite, Webb Township	Unzoned, external potassic pegmatite	Muscovite + Albite + Quartz + Blocky K-feldspar + (Columbite-Tantalite + Garnet + Molybdenite)	Nb>Ta?-(Mo)	Incipient, film-like albitization and associated columbite-tantalite in coarse blocky K-feldspar
Tot Lake Pegmatite, Webb Township	Complex external pegmatite. Contains pink Type 2 spodumene of Heinrich (1975)	Pink Spodumene + Pollucite + Quartz + Albite + Blocky K-feldspar + (Fluorapatite + Garnet + Tourmaline)	Li-Cs-Rb-F-P (Ta>Nb-Be)	Extensive albitization developed in 2 stages: 1. saccharoidal sodic apite (Blue Apatite + Tourmaline + Quartz + Albite) 2. Lepidolite-Albite Complex (+ White Beryl + Garnet + Tantalite + Sericite + Lepidolite + Quartz + Cleavelandite)

¹ Accessory minerals placed in brackets.² Minor elements placed in brackets.

Table 3: Summary of Petrologic Features of Intrusive Phases of the Ghost Lake Batholith.

Intrusive Phase	Rock Type	Petrologic Features	Intrusive Relations
GLB-1	Muscovite-biotite and biotite-muscovite granite and rare granodiorite. Locally cordierite-bearing	Massive to weakly foliated. Coarse-grained to pegmatitic, inequigranular, generally hypidiomorphic-granular, white-weathering	
GLB-2	Muscovite-biotite granite to granodiorite	Massive, occasionally weakly foliated fine- to medium-grained, allotriomorphic-granular. Distinctive medium grey colour	Occurs as small masses and dikes cross-cutting GLB-1.
GLB-3	Muscovite-biotite and muscovite-biotite granite + garnet	Massive to weakly foliated and occasionally lineated, medium-grained, allotriomorphic-granular. Unique presence of fibrolite + quartz + muscovite faserkiesel	No intrusive relations with adjacent GLB-1 or 2 observed. GLB-3 contains layers and small pods of gradational coarse-grained to pegmatitic granite
GLB-4	Muscovite and biotite-muscovite pegmatitic leucogranite + garnet + tourmaline	Massive to occasionally vague layered hypidiomorphic granular	Observed as veins and dikes cutting GLB-1 and 2.
GLB-5	Fine-grained garnet-muscovite-tourmaline leucogranite	Massive to occasionally layered, fine-grained, allotriomorphic-granular, usually equigranular	Noted only in eastern part of Ghost Lake Batholith; intruded by GLB-6 and 7.
GLB-6	Garnet-biotite-tourmaline-muscovite pegmatitic leucogranite	Massive, hypidiomorphic-granular, with distinctive randomly oriented bladed aggregates rich in green muscovite	Cross-cuts GLB-5
GLB-7	Muscovite-tourmaline potassic pegmatite + garnet	Massive, hypidiomorphic granular, may be associated with fine-grained, banded sodic aplite. Contains rare green syngenetic beryl.	Cross-cuts GLB-1, 4, 5, and 6 Youngest granitic phase in Ghost Lake Batholith. May be correlative with exocontact, pegmatitic dikes having tourmaline haloes and zoned garnet-tourmaline-green muscovite dike on Highway 17 10.2 km west of Dryden.
GLB-8	Tourmaline-quartz veins	Small veins concentrated towards eastern part of Ghost Lake Batholith. Relatively common in metasedimentary rocks between Dryden and Thunder Lake. Featured by locally impressive metasomatic tourmaline haloes in host rocks.	

northwestern mass of GLB-1 towards the narrow eastern lobe adjacent to the Mavis Lake Pegmatite Group. A second trend of geochemical fractionation is also developed within the batholiths and directed south to southwest towards plutonic unit GLB-3.

The Mavis Lake Pegmatite Group is almost entirely contained within the Brownridge volcanics and is 8 km in strike length and .8 to 1.4 km in breadth. A well developed lateral zonation of pegmatites is apparent over this strike length. The zonal pattern of typomorphic mineral assemblages, internal structures, degree of replacement stage albitization and characteristic geochemical associations changes progressively at increasing distances from the parental Ghost Lake Batholith. This succession of rare-metal pegmatites is similar, for example, to that described in the Yellowknife-Beaulieu district (Hutchinson, 1955), Preissac-Lacorne area of Quebec (Mulligan, 1965), and the Issia et Touvre area of the Ivory Coast (Varlamoff, 1972). The degree of late-stage albitization increases from west to east in the Mavis Lake Pegmatite Group such that thin (1-3 m width) pegmatite sheets exposed at the East Zone of the Fairservice Property consist essentially of secondary mineral assemblages. These mainly consist of white sodic beryl + green muscovite + quartz + cleavelandite + tantalite. Primary relict spodumene, largely converted into fine-grained albite + green mica, ragged blocky K-feldspar and quartz are, nevertheless, still recognizable. Further west at the Main Zone albitization is much less pronounced, consuming between 20 and 50 percent of primary pegmatite units. In many instances here the albitization appears directly related to a rudimentary quartz-rich core zone as at STOP 4.

Metasomatic release of Li, K, Cs, Sn, B, Rb and F can develop impressive endogenic dispersion patterns in mafic metavolcanic rocks enclosing albitized spodumene pegmatites as exemplified by the Main and South Zones of the Fairservice Property. These elemental concentrations are mineralogically attributable to secondary biotite, tourmaline, and holmquistite which are intensely developed within layers and pods in proximal mafic host rocks. Five analyses of these metasomatic assemblages, (termed glimmerite in Russian literature) from immediate contacts of Mavis Lake and Gullwing Lake - Tot Lake rare-metal pegmatites are presented in Table 4. Be, Cs, F, Li, Rb and Sn all have obvious utility in lithogeochemical surveys, however, in reconnaissance surveys designed to locate rare-metal pegmatite fields, lithium represents the most useful element (Ovchinnikov, 1976). Pryslak (1981) documented several impressive lithium aureoles on the Main Zone with the largest measuring 60 - 185 m by 1220 m and containing maximum lithium concentrations of 4095 ppm.

The Gullwing Lake-Tot Lake rare-metal Pegmatite Group (0.8 to 2.2 x 8.6 km in area) is situated between the northeast end of Gullwing Lake and Tot Lake and is mainly confined to a medium-grade, highly deformed, southwest-striking mafic metavolcanic unit which envelopes the Lateral Lake Stock (Breaks et al., 1978). Six rare metal pegmatites have been discovered to date (Table 2) which are distinctive in several ways from those

Table 4: Selected Chemical Analyses¹ of Metasomatic Selvages Developed at Contacts of Some Rare-Metal Pegmatites from the Dryden Pegmatite Field.

Spodumene Pegmatite No. 2 Fairservice- Property	B	Be	Cs	F	Li	Rb	Sn	Ta	Nb	K ₂ O
	450	22	920	7650	6050	3790	68	<30	13	3.69%
(tourmaline-holmquistite-biotite glimmerite)										
Pegmatite No. 2 Taylor Beryl Occurrence	5	20	1150	7250	2300	2210	762	<30	70	4.58%
(biotite-rich glimmerite)										
Gullwing Lake Spodumene Pegmatite	<5	6	1140	18400	4600	7580	167	60	925	6.39%
(biotite-rich glimmerite)										
Tot Lake Spodumene- Pollucite Pegmatite 25		12	1.60%	5930	1900	4820	23.9	N.D.	N.D.	4.78%
(biotite-rich glimmerite)										
Gullwing Lake Batholith at STOP 6.	N.D.	6	755	17,800	1400	1860	142	<10	7	3.70%
(biotite-tourmaline glimmerite)										
Petrunka Tungsten 4200 Property at STOP 5 (tourmaline + biotite metasomatized pillow selvage immediately below scheelite-tourmaline sheet)		3	90	1010	1240	50	3.7	N.D.	35	0.13%

¹ in ppm unless otherwise shown
N.D. = Not Determined

in the Mavis Lake Pegmatite Group:

- 1) Most pegmatites are significantly discordant to the general 070° structural trend of host rocks.
- 2) Presence of internally zoned pegmatites exhibiting more complex geochemical specialization i.e. Li-Cs-Rb-Be-Ta>Nb and Li-Ta>Nb-Be-Mo.
- 3) no evidence of associated scheelite mineralization;
- 4) absence of tourmaline replacement of host rocks, and much more confined development of metasomatic glimmerite assemblages.
- 5) regional zonation of rare-metal pegmatite types not as clearly developed.
- 6) parental granitic source area is not clearly apparent.

Most pegmatites of this group are oriented sub-perpendicular to the curvilinear contact between the mafic metavolcanic host unit and Lateral Lake Stock. Exceptions to this include the Mica Point columbite pegmatite (concordantly emplaced in the mafic metavolcanic unit where it deflects north to northeast around western terminus of the Lateral Lake Stock) and the Drope Township columbite-molybdenite pegmatite exposed at the extreme southern known limit of the pegmatite group. The latter pegmatite was emplaced concordant to foliation of clastic metasedimentary host rocks.

The source area for rare-metal pegmatites of the Gullwing-Tot Lakes Group is not known at present. The Coates Bay Stock, located near the southwestern limits of the pegmatite group, could represent a parental source region for the pegmatites but has not been adequately studied. The easternmost part of this granitic stock consists of garnet-muscovite pegmatitic leucogranite characterized by plumose aggregates of fine-grained green muscovite and quartz.

The relationship of rare-metal pegmatites of this pegmatite group to molybdenite-rich granitic pegmatites of the Lateral Lake area, approximately 4.8 km to the east of the Tot lake pegmatite, is presently uncertain. Although these molybdenite-rich pegmatites were not included in the study of Breaks (in preparation), a genetic relationship could very well be involved since molybdenite is an accessory in most of the rare-metal pegmatites of the Gullwing-Tot Lakes Pegmatite Group.

Uraniferous Pegmatites

Uranium mineralization within the boundary zone is concentrated in the Vermilion Bay area. These occurrences are generally concentrated along the margins of potassic intrusions associated with the Dryberry Batholith. The mineralization occurs in coarse grained to pegmatitic granitoid rocks and contains uraninite, uranothorite and allanite as common minerals. Most uranium occurrences in the northern part of the English River Subprovince occur in inhomogeneous diatexites where boundaries with supracrustal rocks are gradational into pods of uraniferous granitoid material. In the Vermilion Bay area however, supracrustal host rocks are metavolcanic flows, tuffs and associated sedimentary rocks. Radioactive mineralization in this case is confined to discrete pegmatitic leucocratic dikes

and sills which are clearly intrusive into the host rocks.

Building Stone

Some phases of the late-tectonic potassic suite are suitable sources of monumental and building stone in that they are homogeneous, lack metamorphic foliation and locally contain few joints. The Nelson Granite Ltd. quarry near Vermilion Bay (STOP 2) is developed on a plug of granite of this type within the Dryberry Batholith (Storey, 1983).

THE FIELD TRIP

The field trip will commence at the Inn of the Woods, Kenora. Proceed approximately 37 km (23 miles) east along the Trans-Canada Highway (Hwy 17) to a point of intersection with the C.P.R. rail line (overpass). Proceed an additional 650 m east of the overpass and walk approximate 250 m south from the highway to an area of stripping and trenching.

STOP 1: BYBERG (HAWK LAKE) URANIUM OCCURRENCE

Uranium mineralization is concentrated in masses and stringers of coarse magnetite within a wedge-shaped mass of pegmatite (Figure 2). The pegmatites are contained within a package of mafic volcanic rocks, now largely hornblende schists. These volcanic rocks, along the contact with the main pegmatite mass, are very iron rich, locally magnetic, and layered in appearance.

The main showing has been explored by stripping and 10 trenches, one open cut 130 ft. long, scintillometer surveys, and 2 diamond drill holes. None of the uraniferous zones are more than 1.6 m wide or 9 m long. Earlier work by J. Satterly of the Ontario Department of Mines suggests that although an average of 7 grab samples from the narrow magnetite-rich sections of the pegmatite ran 1.9 lbs U_3O_8 per ton, the main mass of pegmatite will probably run less than 1 lb U_3O_8 per ton. Diamond drilling by Kerr Addison Mines Ltd. in 1975 intersected one zone that ran 0.41 lb U_3O_8 per ton over 49 feet. A bulk sample collected in the early 50's had an average content of 1.4 lbs U_3O_8 (chem. equiv.) per ton.

Uranium minerals which have been identified are uraninite, thorite, uranothorite, uranophane, and B-uranotil. In 1977, J. Douglas Scott, formerly of Eldorado Nuclear, working with magnetite-rich material from the Byberg deposit, identified what may be either a new mineral or a peculiar altered allanite.

Return to highway and proceed eastward approximately 43 km (27 miles) to the Nelson Granite Quarry.

STOP 2: GRANITE QUARRY

Examine the highway exposures. The "granite" at this locality is distinctive in its uniform pink colour, consistent texture and freedom from fracturing as evidenced by the highway cut. This is in distinct contrast with the surrounding gneissic rocks exposed along the highway on either side. It is intrusive into the gneissic granitoid rocks and xenoliths of these can be found near the contacts of the body. Only the central part is satisfactory as dimension granite due to these xenoliths. This granite is currently being quarried as monumental and building stone by two companies, Nelson Granite Ltd. on the south side of the highway and Granite Quarries (GQI) Inc. (formerly Scotstown Granite and Universal Granite Centre (1976)

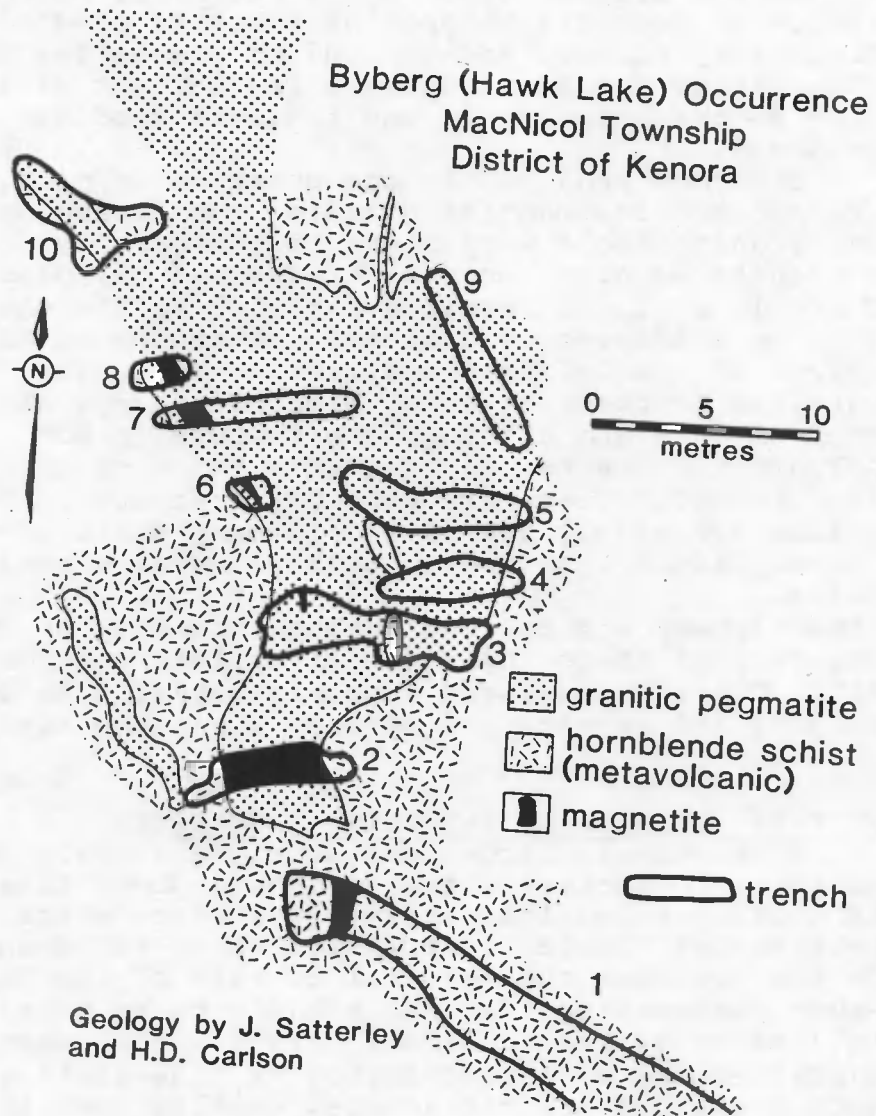


Figure 2: Geology of the Byberg Uranium Occurrence.

Ltd. on the north side.

The Nelson Granite Ltd. quarry was first opened in 1981, producing stone largely for monuments. More recently, however, the production is more evenly split between monumental and construction uses. Annual production is on the order of 50,000 cubic feet per year.

At the present time, rough blocks of approximately 100 cubic feet are shipped by truck to plants in New Brunswick, Quebec, and the United States for finishing. The company has constructed a small plant at the quarry site so that sawn slabs and finished products may be produced.

The rock produced in the quarries is pink, medium grained massive biotite granite: Modal analyses plot in the granite field very close to the granite - quartz monzonite boundary on the Streckeisen QAP diagram. There is a slight layering parallel to the sheeting. This is evidenced by a slight lightening of the pink colour of the feldspar and biotite is roughly aligned parallel to these layers. Microcline crystals reach up to 1 cm in width although the texture is not porphyritic. A few minor quartz veins up to 2 cm wide are present. There are very few fractures. Vertical joints are widely spaced and in many parts of the outcrop are not evident. Subhorizontal sheeting is thick.

Return to the highway and proceed through the town of Vermilion Bay approximately 27 km to the Eagle River intersection (Hwy 17 and Hwy 596). Continue eastward approximately 0.5 km further along Highway 17 and examine the outcrop along the north side of the road.

STOP 3: GHOST LAKE BATHOLITH MIGMATITES

A relatively large zone of migmatization is situated adjacent to northwest flanks of the Ghost Lake Batholith exhibiting migmatitic stages (protometatexite, metatexite) (Table 1) very similar to those encountered in the northern supracrustal terrain of the English River Subprovince (Ear Falls-Manigotagan belt) delineated by Breaks et al., (1978). Protometatexite is characterized by intercalation of fine-grained wacke and medium-grained pelitic layers, usually less than 15 cm in thickness, in which the pelitic component contains sparse non-amalgamated leucosome patches. Note foliation lying at an angle to bedding of about 20° and coplanar with axial surfaces of small scale relatively open, similar type S-folds. Development of metatexite in which layers of thin hololeucocratic mobilizate have been injected concordant and sub-concordant to bedding is subordinate. Subisoclinal to isoclinal S-folding of pelitic and leucosome layers is also apparent and may represent the first phase of folding, although interfering relations with the more open variety have

not been observed in this particular area. These isoclinal folds exhibit coplanarity of axial planes and bedding. Pelitic units are characterized by the assemblage almandine + K-feldspar + cordierite + biotite + quartz + plagioclase. Cordierite porphyroblasts can be recognized as light brown pits on weathered surfaces.

Note also presence of sillimanite, mauve cordierite, and rare blue dumortierite in some of the white intrusive anatectic material.

Proceed 14.8 km (9.25 miles) east along Highway 17b to village of Oxdrift.

STOP 4: GHOST LAKE BATHOLITH

All Highway 17 roadside exposures of white granitoid rocks between the Aubrey Creek bridge east to the Dryden cemetery are situated within the Ghost Lake Batholith. This roadcut is located in the western part of the batholith (Figure 1) and features three plutonic phases (GLB 1, 2 and 4 in Table 3) of this regionally extensive S-type granitoid complex. Most of the outcrop consists of inhomogeneous diatexite (GLB-1), characterized by a significant proportion of metasedimentary inclusions and restite (melanosome) patches. The latter may locally contain significant amounts of sillimanite and/or muscovite. These minerals may also appear within enclosing white, inequigranular, medium- to coarse-grained granite leucosome. Almandine garnet, blue-green apatite, and rare deep-blue cordierite may also be recognizable.

Dikes of fine-grained, grey, muscovite-biotite granite (GLB-2) and coarse-grained, white, holo-leucocratic granite (GLB-4) represent later intrusive phases of the Ghost Lake Batholith.

Proceed 8.8 km east along Highway 17 to a point where it makes a sharp turn south. Continue through Dryden for 13.6 km. Turn north on Thunder Lake Road and continue for 1.7 km. Turn northwest on the south Ghost Lake access road, travelling for 1.5 km. Turn west on poorly maintained powerline road and travel for about 1.1 km to point where road departs north from powerline. Travel a further 0.6 km to base of small ridge marked by a steel gate (usually locked). Proceed west for 0.27 km and then turn south on one of many exploration access roads constructed by Sanmine Exploration Incorporated and travel for about 0.37 km. Property access roads are best travelled by fourwheel drive vehicles. Consult figure 3 for stop locations.

PLEASE DO NOT HAMMER SURFACE OF STRIPPED OUTCROPS. SELECT MINERAL SPECIMENS FROM LOOSE BLASTED MATERIAL ONLY.

STOP 5: PETRUNKA TUNGSTEN PROPERTY, SANMINE EXPLORATION INC. (EAST END)

Scheelite mineralization occurs in generally flat-lying, undulating tourmaline-rich sheets. The sheets which variably transect foliation trends of host

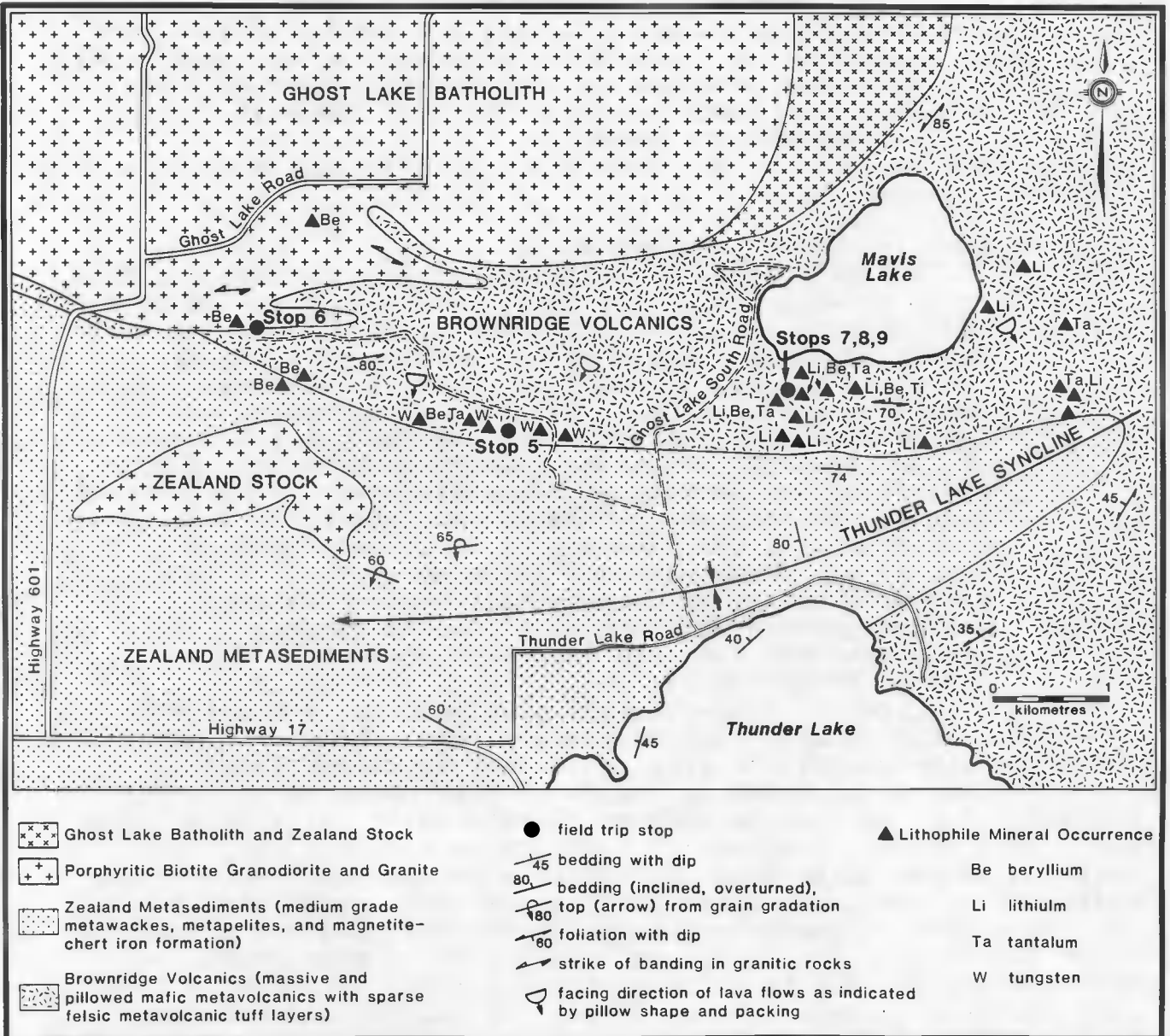


Figure 3: Geology of the Mavis Lake Pegmatite Group.

pillowed mafic metavolcanics are up to 0.3 m in thickness and dip between 11 and 19 degrees. Highest levels of scheelite found to date occur within these tourmaline sheets accompanied by ancillary amounts of biotite, quartz, plagioclase, pyrite and possible lepidolite. An identical mineral assemblage is developed on several spatially associated, ancillary fracture sets with relatively steep dips (69-85°) and along 7 mm - 5 cm wide pillow selvages, the latter of which have been intensely replaced by tourmaline and biotite in the vicinity of the scheelite mineralization and contain trace level anomalous enrichment in Cs and Li (Table 4). Scheelite, up to 5 by 7.5 cm size, is predominantly anhedral, poikilitic, and honey-brown on clean weathered surfaces. Analysis of a 3.5 m channel sample selected from a flat-lying biotite-tourmaline sheet containing the highest scheelite content observed to date is given below (in ppm unless otherwise stated):

W	Be	Cs	F	Li	Sn	Au
950	45	40	9,600	246	193	<2 ppb

Levels of tungsten are similar to those previously documented by Noranda Mines Limited who reported channel samples up to 0.1 percent WO₃ (Blackburn and Hailstone, 1983, p.12-13) and Selco Incorporated who obtained grades up to 0.09 percent WO₃ (A. Pryslak, geologist, Selco Incorporated, personal communication, 1983).

Return to main property access road and travel west for 2.3 km (1.5 miles).

STOP 6: PETRUNKA TUNGSTEN PROPERTY, SANMINE EXPLORATION INC. (WEST END)

This stop is important in that it establishes a genetic connection of rare-metal mineralization and associated boron metasomatism with the Ghost Lake Batholith.

This large stripped outcrop exposes the rarely observed contact of the Ghost Lake Batholith with deformed mafic and intermediate metavolcanics. The contact at the west part of this outcrop is defined by an interlayering of medium- to coarse-grained almandine-muscovite-tourmaline granite with calc-silicate bearing mafic metavolcanics over 35 m. In places a brown weathering 1-8 cm wide biotite-rich metasomatic selvage is developed along the contact.

About 50 m to the east the contact is more abrupt and is characterized by an outermost zone (1-2 m wide) of green muscovite enrichment which reverts inward to a much less micaceous almandine-muscovite-tourmaline granite. Impressive zones of intense black tourmaline deposition occur adjacent to the green muscovite-rich granite. The most extensive tourmaline-rich zone (2 m wide) contains radiating sheaves of tourmaline with individual crystals up to 8 cm in length, brown weathering biotite, and actinolite. These pneumatolytic

deposits exhibit anomalous enrichment in Be, Cs, Rb, Sn, F, and B (Table 3). With increasing distance from the batholithic contact a gradation of the tourmaline-rich zone into albite-quartz-rich pegmatitic pockets is obvious. Accessory, euhedral, light green beryl up to 5.4 x 6.5 cm in basal sections occurs in these quartz-rich masses. Zoned quartz-tourmaline veins, exposed near the eastern limits of the stripped outcrop, are generally subconcordant to host rock foliation and clearly emanate from the Ghost Lake Batholith.

Return to south Ghost Lake access road, turn north, and proceed for 1.5 km. Turn east on a very inconspicuous old drill access road and follow for 0.3 km. Park vehicle and walk on drill road for about 215 m.

**STOP 7: MAVIS LAKE SPODUMENE-BERYL-TANTALITE PEGMATITES
(FAIR SERVICE PROPERTY, MAIN ZONE)**

This 12 x 76 m lensoidal-shaped pegmatite dike occurs in foliated and gneissic mafic metavolcanics and subordinate fine-grained laminated metawackes. It exhibits a vague internal zonation which is generally not apparent in most rare-metal pegmatites of the Mavis Lake Pegmatite Group. A quartz-rich core zone is irregularly and asymmetrically distributed from the dike centre to north contact. These pod-like domains contain conspicuous blocky K-feldspars up to 33 x 41 cm in cross-sectional area, and minor beryl, green spodumene, green muscovite, tourmaline, blue apatite, and orange garnet.

Outer margins of the core zone can grade indistinctly into an albite-spodumene-quartz pegmatite unit containing 37 to 47 percent quartz and 29 to 48 percent green spodumene, the latter being highest levels recorded for any pegmatite of the Mavis Lake Pegmatite Group.

A third zonal unit is developed along the outer margins of the spodumene-rich unit and contains the assemblage green muscovite-spodumene-albite-quartz-K-feldspar. Spodumene contents drop to 7 percent while those of K-feldspar increase to 55 percent. It consists of coarse, generally anhedral, coalescing white K-feldspars with interstitial quartz-spodumene "pockets".

Small, white to buff coloured domains (up to 1 x 3 m) of fine-grained, equigranular, saccharoidal sodic aplite are sparsely scattered throughout the spodumene-rich zone. These domains consist of albite, quartz and accessory green muscovite, apatite and an unidentified brown mineral (possibly a phosphate). This type of aplite appears to be primary in origin as contacts with bounding spodumene-rich units are invariably sharp and exhibit no evidence of replacement textures. At STOP 8, identical aplites are clearly transected by veinlets of coarser albite and green

muscovite (best observed on wetted surfaces) which probably relate to replacement stage aplites.

Albite-rich replacement zones, commonly peripheral to quartz-rich core zones, typify all spodumene pegmatites of the Mavis Lake area. Approximately 10 percent of this pegmatite has been intensely albitized to the point that no vestige of primary mineralogy has survived. Textures are distinctly non-equigranular, fine to medium grained and involve albite, quartz, green muscovite, with accessory white beryl, garnet, and rare ilmenite.

Columbite mineralization in this pegmatite is sporadically distributed and clearly associated with replacement stage albitization, as evidenced by its restriction to albitized rims of some blocky K-feldspar aggregates and spodumenes.

Proceed about 60 m south to side of hill.

STOP 8: TANTALITE-BEARING SPODUMENE PEGMATITE DIKE

The lowermost part of a bifurcating, 268 m long pegmatite dike contains sparse disseminated tantalite, distributed uniformly relative to other pegmatites in the area. Maximum values of tantalum (380 ppm over 0.9 m) and Nb (210 ppm over 0.9 m) for the Main Zone Mavis Lake area pegmatites occur at this locality. The rare lithium amphibole holmquistite $[\text{Li}_2(\text{Mg}, \text{Fe}^{2+})_3 \text{Al}_2\text{Si}_2\text{O}_{22}(\text{OH})_2]$ is subtly apparent in mafic metavolcanics above the upper 50 degree south-dipping contact.

Most of this pegmatite is composed of a K-feldspar-rich zone similar to that at STOP 7 but can be distinguished by the conspicuous development of a symplectic green spodumene + quartz intergrowth. Slender spodumene phenocrysts up to 1 m long and 12-19 cm wide are oriented approximately normal to the lower pegmatite contact and contain fine-grained, equigranular quartz comprising about 30 percent of the host. A 4 to 5 cm wide quartz + green muscovite + albite carapace surrounds most of the spodumene-quartz intergrowths. The entire intergrowth is typically enveloped by coarse-grained K-feldspar.

This pegmatite also contains local development of a quartz-rich core (accompanied by blocky K-feldspar, white euhedral beryl up to 5.5x7 cm in cross-sectional area, and spodumene) and pockets of saccharoidal sodic aplite similar to that at STOP 7.

Return to drill road and walk short distance east to top of hill.

STOP 9: ALBITIZED SPODUMENE PEGMATITE WITH METASOMATIC LAYERS DEVELOPED IN HOST ROCKS

This relatively highly albitized portion of the same dike as that observed at Stop 8 exhibits well

developed, fine-grained saccharoidal albite and quartz. Glimmerite* bands are apparent in the mafic metavolcanics immediately north of the contact and disposed concordant to foliation. The glimmerite in this case is characterized by enrichment in bronze-coloured biotite, black tourmaline, and sporadic holmquistite. A partial chemical analysis is given in Table 4.

Geochemical profiles for distribution of mobile elements K, F, Li, Rb, Cs, Sn and Be in mafic metavolcanic host rocks as a function of increasing distance from northern contact of Pegmatite No. 4, situated 130 m east of this stop is given in Figure 4.

Return to south Ghost Lake access road and travel back to Highway 17. Proceed west to junction with Highway 601 and turn north. Travel for 4.8 km to junction with Ghost Lake Road and turn east. Proceed for 23.5 km and turn west on old gravelled logging road for 1.8 km.

Turn southwest on very poor clay-surfaced, abandoned logging access road and travel in four-wheel drive for 1.7 km. Walk northwest up very poor, partly overgrown logging road for 440 m to top of hill. Then walk for about 260 m north along pegmatitic dike forming a ridge to base of outcrop area on north slope.

STOP 10: GULLWING LAKE SPODUMENE PEGMATITE

This zoned pegmatite, discordant to amphibolitic mafic metavolcanics, is situated at the northern end of a 412 m long, curvilinear, pegmatitic granite dike striking N18°W and dipping 52°-72° east. Most of this pegmatite consists of a blocky K-feldspar + quartz assemblage which represents the coarsest grain size of any pegmatite in the Dryden Field. A small core zone of spodumene-blocky K-feldspar-quartz-albite + muscovite occurs at the lowest exposure. The coarse light-green randomly oriented spodumene phenocrysts measure up to 4 x 100 cm (parallel to 001) and 6 x 40 cm (normal to 001).

Two stages of replacement albitization have been identified within this pegmatite. The most dominant and earliest consists of an equigranular, albite-rich assemblage accompanied by minor quartz, muscovite, and remnants of primary blocky K-feldspar. This albitic replacement unit closely follows external contacts of the Gullwing Lake Pegmatite and those of several mafic and intermediate metavolcanic xenoliths. A variant of this albitic assemblage occurs near the upper pegmatite contact and is featured by an interesting bladed texture formed by randomly oriented, thin biotite and muscovite

* Russian term for a rock resulting from metasomatic transformation by outward dispersion of fluids enriched in such elements as K, B, Li, Cs, F, Rb which were released due to late stage albitization of primary spodumene + K-feldspar bearing assemblages.

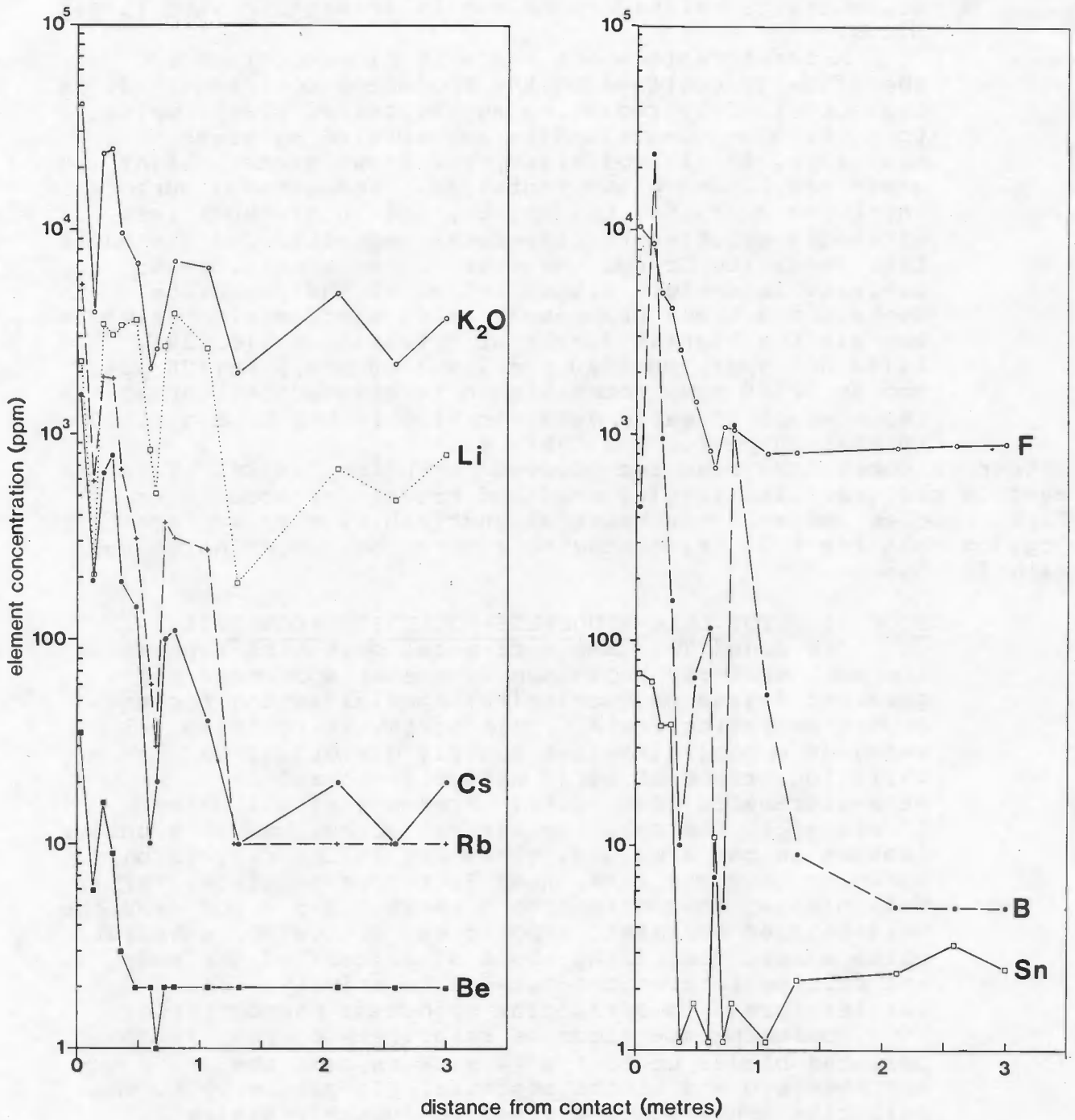


Figure 4: Geochemical profile within mafic metavolcanic rocks adjacent to the contact with a pegmatite dike.

laths up to 10 cm. Sparse molybdenite and platy columbite-tantalite can be subtly intergrown with these micas.

A later replacement stage is finer-grained and specifically confined to the spodumene core zone. It is characterized by radiating aggregates of platy, white to light blue cleavelandite accompanied by green muscovite, local lepidolite, red-brown garnet, light green beryl, and platy tantalite. Endogeneous aureoles involving K, F, Li, Cs, Rb, Be, and Sn are much less extensive relative to rare-metal pegmatites of the Mavis Lake Pegmatite Group. However, intense metasomatic activity is obvious within 3-5 cm of the pegmatite contact and these brown-weathering biotite-rich selvages contain the highest levels of F(1.84%), K_2O (8.82%), Li(11,000 ppm), Rb(7580 ppm), Ta(100 ppm), Sn(320 ppm) and Zn (2560 ppm) found within in glimmerites formed via replacement of mafic metavolcanics in the Dryden Field (Breaks in prep., and Table 4).

Return to Ghost Lake road and proceed north for 1.8 km. Turn east on old gravelled logging road and travel for about 1 km. Park vehicles and walk northeast up undrivable, clay surface logging road for 1.35 km, coming to a junction. Turn north and walk for 345 m.

STOP 11: TOT LAKE SPODUMENE-POLLUCITE PEGMATITE

The zoned Tot Lake rare-metal pegmatite represents the most easterly known occurrence of spodumene and greatest degree of geochemical specialization for the Dryden Pegmatite Field. This northwest-striking 1-6 m wide, 48 m long dike lies sharply discordant to foliation trends of mafic metavolcanic and meta-ultramafic host rocks. Presence of pollucite ($CsAlSi_2O_6$), the only ore mineral of cesium, is a unique feature in the area and, elsewhere in Ontario, is only known at Lily Pad Lake, near Fort Hope (Wallace, 1978). This mineral is confined to a small 1 x 5 m pod near the well-layered southeast exposed end as coarse, anhedral, white masses comprising about 37 percent of the mode, and filling interstices between relatively earlier-formed, bladed, pink spodumene phenocrysts.

Spodumene may occur as relatively coarse, randomly oriented blades up to 1 x 14 x 38 cm near the northwestern end of the exposure. In proximity to the pollucite zone, finer-grained spodumene exhibits sub-horizontal alignment in longitudinally oriented bands with c-crystallographic axis approximately normal to dike contacts. These bands contain about 70 percent spodumene and alternate with albite-quartz-green muscovite aplitic layers containing little spodumene.

A significant portion of the pegmatite has been subjected to late stage fracture-controlled albitization. This moderate to intense albitization was effected by hydrothermal fluids emanating from a now

quartz-filled central branching vein system extending for virtually the complete exposed dike length. Near the northwestern end, metasomatic fronts oriented parallel to thin quartz-filled fractures intersect pink spodumene phenocrysts, partly altering these (behind the front) to fine-grained green micas. A tantalite-rich zone in the same part of the dike exhibits spodumenes and blocky K-feldspars that have been completely replaced by albite-green mica (cymatolite) and albite, respectively. Lepidolite-albite replacement assemblages occur locally at the southeastern end of the dike and are specifically associated with the pollucite zone where spodumenes are clearly partially replaced by fine-grained lepidolite.

Beryl is rarely found in the Tot Lake pegmatite, occurring as white subhedral to anhedral crystals (up to 2 x 4 cm in cross-section) in albitic replacement zones. Holmquistite is noteworthy as fine-grained, randomly oriented acicular purplish crystals on the surface of a massive, medium-grained, mafic metavolcanic host reentrant near the southeast end (best seen on wetted surfaces) and in similar adjacent rocks for distances up to 1 m from the dike contact.

Return to Ghost Lake Road, turn north, and travel for 7.75 km to junction. Turn east along poor gravel road and travel for 5 km to junction with Kathlyn Lake access road. Turn south and travel for about 3.5 km. Proceed a short distance along old road to west to an outcrop area with pits and an adit.

STOP 12: LATERAL LAKE MOLYBDENITE DEPOSIT

This property is held by Pidgeon Molybdenum Mines Ltd. and covers the eastern contact between the Lateral Lake intrusion and intensely foliated metavolcanic and metasedimentary rocks (Figure 5). The property has been explored by an adit and extensive diamond drilling, most recently by Rio Tinto Canadian Exploration Ltd. in 1979-80. At this locality the intrusion is composed of foliated, medium to coarse-grained granodiorite, pink aplite and several varieties of pegmatite. Quartz-filled tension fractures are oriented at a high angle to foliation and cut all of the intrusive phases (Page, 1984). Mineralization consisting of molybdenite with lesser pyrite is preferentially associated with late feldspar and muscovite-rich intrusive phases. Molybdenite is present as scattered flakes in muscovite-bearing pegmatites and vein quartz. Pyrite and muscovite are common associates of molybdenite and rare magnetite, bismuthinite and tourmaline have been reported.

The most recent development on this property has resulted in a published reserve estimate of 14,300,000 tons grading .08% MoS₂.

Return to Kathlyn Lake access road and continue eastward for 8.6 km (5.4 miles) to the intersection with Highway 72. Turn right

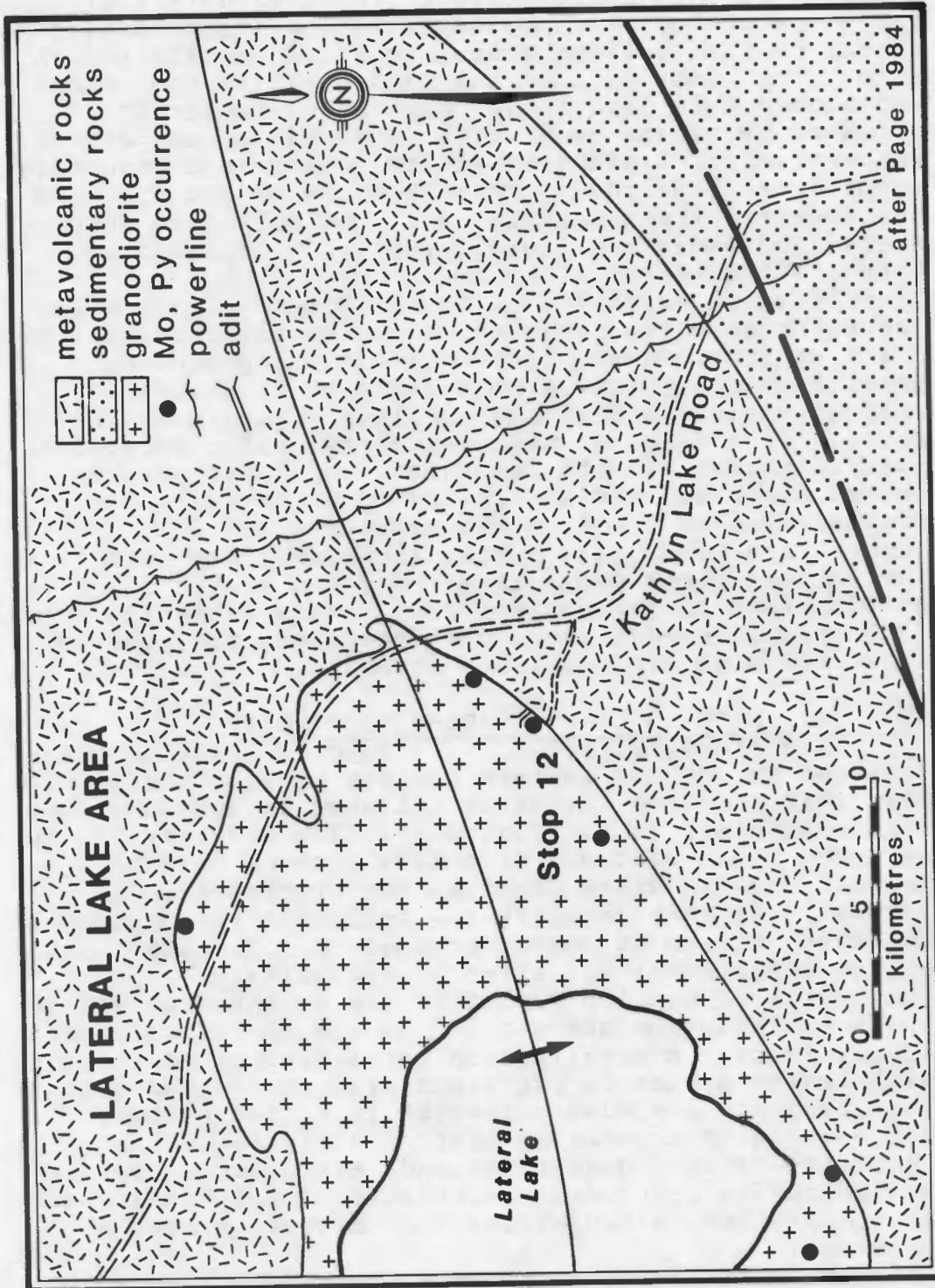


Figure 5: Generalized Geology of the Lateral Lake Area.

(south) and proceed 2.7 km (1.7 miles) to the Goldlund Mine access road. Turn right (west) and proceed along this road approximately 2.4 km (1.5 miles) to the mine.

STOP 13: GOLDLUND MINE

The Goldlund deposit is operated by Campbell Resources Ltd. and contains publishes reserves of 600,000 tons grading .20 oz/ton to the 800 ft. level. It is hosted by a tabular northeasterly striking body of trondhjemite which intrudes and is grossly concordant with mixed mafic to felsic metavolcanic rocks (Figure 6). A large discordant body composed of quartz and feldspar porphyry occurs in the mine area and numerous dikes of similar composition cut the trondhjemite and the volcanic rocks. The ore zones comprise clusters of enechelon extensional veins which transect the trondhjemite: these strike northerly and dip at moderate angles to the west. The veins are fractions of an inch to one foot wide and occur at spacings of inches to several feet.

The veins are composed primarily of quartz, ankerite and pyrite with minor actinolite, biotite, tourmaline, scheelite, sphalerite, galena, chalcopyrite, pyrrhotite, tellurides, ilmenite and native gold. Most veins are enveloped by "bleached" alteration haloes within trondhjemite: these are characterized by enrichments of ankerite, albite and pyrite but their width and intensity of alteration does not correlate directly with gold contents.

The mine has been developed by conventional underground stoping, a decline from surface and open cuts at surface. Depending, on the status of mining operations at the time of the field trip, one or more areas will be visited to examine the host trondhjemite, porphyry dikes, quartz veins and alteration haloes.

Return to Highway 72 and proceed south to the intersection with Highway 17 to Kenora via Dryden and Vermilion Bay.

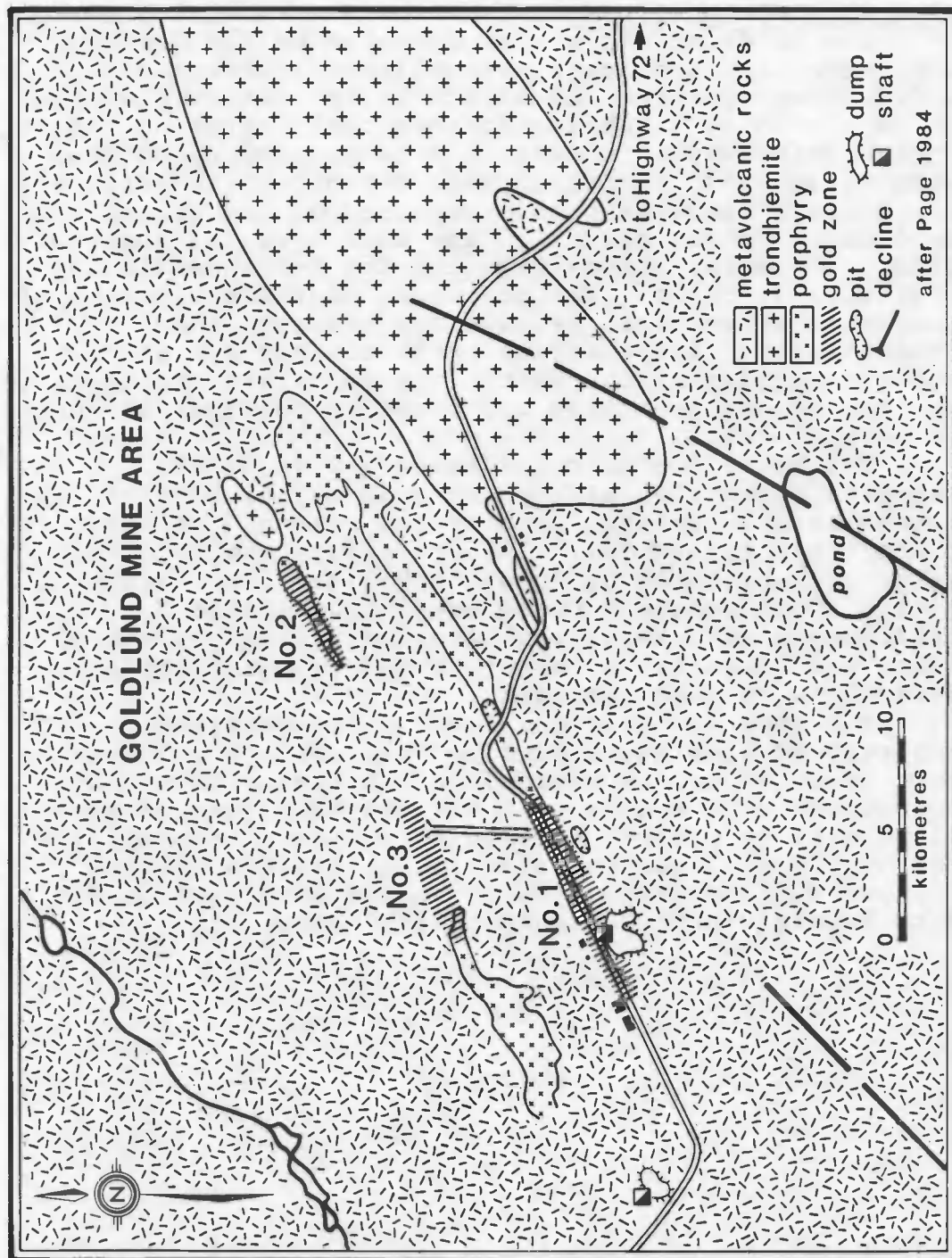


Figure 6: Generalized Geology of the Goldlund Mine Area.

REFERENCES

- Bartlett, J.R., 1978, Metamorphic trends in the metasedimentary rocks north of Eagle Lake, Ontario; unpublished B.Sc. thesis, University of Western Ontario, 73p.
- Blackburn, C.E. and Hailstone, M.R., 1983, 1982, Report of the Kenora Resident Geologist; p.1-19 in Report of Activities Regional and Resident Geologists, 1982, edited by C.R. Kustra, Ontario Geological Survey, Miscellaneous Paper 107, 211.
- Blackburn, C.E., and Janes, D.A., 1983, Gold deposits in Northwestern Ontario; p.194-210 in The Geology of Gold in Ontario; ed. A.C. Colvine; Ontario Geological Survey Miscellaneous Paper 110, 278p.
- Breaks, F.W., 1980, Lithophile Mineralization in Northwestern Ontario: Rare Element Granitoid Pegmatites; p.5-9 in Summary of Field Work, 1980, by the Ontario Geological Survey, edited by V.G. Milne, O.L. White, R.B. Barlow, J.A. Robertson, and A.C. Colvine, Ontario Geological Survey, Miscellaneous Paper 96, 201p.
- Breaks, F.W., 1982, Metamorphism and rare-metal pegmatites of the Dryden area; p.4-12 in Stratigraphy and Structure of the Western Wabigoon Subprovince and its margins, Northwestern Ontario ed. C.E. Blackburn; Geological Association of Canada, Field Trip guidebook No.3, Winnipeg.
- Breaks, F.W., 1984, Eagle River - Ghost Lake area, District of Kenora; Ontario Geological Survey, Preliminary Map P.2623, Geol. Ser., scale 1 in to 1/2 mi.
- Breaks, F.W., in prep., Geology and Geochemistry of the Dryden Pegmatite Field, District of Kenora; Ontario Geological Survey Geological Report.
- Breaks, F.W., Bond, W.D., Westerman, C.J. and Harris, N., 1976a, Operation Kenora-Ear Falls, Dryden-Vermilion Bay Sheet, District of Kenora; Ontario Division of Mines, Preliminary Map p.1203, Geol. Ser., scale 1:63,360 or 1 inch to 1 mile. Geology 1975.
- Breaks, F.W., Bond, W.D., Harris, N., Westerman, C.J. and Desnoyers, D.W., 1976b, Operation Kenora-Ear Falls, Sandybeach-Route Lakes Sheet, District of Kenora; Ontario Division of Mines, Preliminary Map p.1204, Geol. Ser., scale 1:63,360 of 1 inch to 1 mile. Geology 1975.
- Breaks, F.W., Bond, W.D., and Stone, Denver, 1978, Preliminary Geological Synthesis of the English River Subprovince, Northwestern Ontario and its Bearing Upon Mineral Exploration; Ontario Geological Survey Miscellaneous Paper 72, 54p. Accompanied by Map p.1972, Scale 1:253,440.
- Cerny, P., 1982, Anatomy and classification of granitic pegmatites; p.1-39 in Short Course in Granitic Pegmatites in Science and Industry; ed. P. Cerny; Mineralogical Association of Canada, Short Course Handbook, V.8.
- Colvine, A.C. and McCarter, P., 1977, Geology and mineralization of the Lateral Lake stock, District of Kenora; p.205-208 in Summary of Fieldwork, 1977, by the Geological Branch, ed. V.G. Milne et al., Ontario Geological Survey, Miscellaneous Paper 75.

- Colvine, A.C., Andrews, A.J., Cherry, M.E., Durocher, M.E., Fyon, A.J., Lavigne, M.J. Jr., Macdonald, A.J., Marmont, S., Poulsen, K.H., Springer, J.S. and Troop, D.G., 1984, An integrated model for the origin of Archean lode gold deposits; Ontario Geological Survey, Open File Report 5524, 84p.
- Heinrich, E.W., 1975, Economic Geology and mineralogy of petalite and spodumene pegmatites; Indian Journal of Earth Sciences, V.2, No.1, p.19-29.
- Hutchinson, R.W., 1955, Regional Zonation of Pegmatites near Ross Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Bulletin 34.
- McCarter, P., 1980, Geology and mineralization of the Lateral Stock, District of Kenora, Northwestern Ontario; unpublished M.Sc. thesis, Oregon University, 141p.
- Moorhouse, W.W., 1939, Geology of the Eagle Lake Area; Ontario Department of Mines, Vol.48, pt.4, p.1-31. Accompanied by Map No.48d, Scale 1 inch to 1 mile.
- Mulligan, Robert, 1965, Geology of Canadian Lithium Deposits; Geological Survey of Canada Economic Geology Report 21.
- Ovchinnikov, L.N., 1976, Lithogeochemical Methods for Prospecting Rare-Element Pegmatites; Acad. Sci. U.S.S.R. Moscow, 79pp.
- Page, R.O., 1984, Geology of the Lateral Lake Area, District of Kenora; Ontario Geological Survey, Open File Report 5518, 175p.
- Pryslak, A.P., 1981, Rafter Project - Fairservice option No.1, unpublished report; Assessment Files Research Office, Ontario Geological Survey, Toronto.
- Satterly, J., 1941, Geology of the Dryden-Wabigoon area, Kenora District; Ontario Department of Mines, Annual Report for 1941; v.50, pt.2, p.1-57. Accompanied by Map 50e, scale 1:63,360 or 1 inch to 1 mile.
- Storey, C.C., 1983, Preliminary Report of the Building and Ornamental Stone Inventory, Kenora and Rainy River Districts; Ontario Geological Survey Open File Report 5446, 143p., 20 tables and 37 figures.
- Streckeisen, A., 1976, To each plutonic rock its proper name; Earth Science Reviews, v.12, p.1-33.
- Trowell, N.F., Blackburn, C.E. and Edwards, G.R., 1980, Preliminary geological synthesis of the Savant Lake-Crow Lake metavolcanic-metasedimentary belt, Northwestern Ontario and its bearing upon mineral exploration; Ontario Geological Survey, Miscellaneous Paper 89, 30p.
- Varlamoff, N., 1972, Central and West African rare metal granitic pegmatites, related aplites, quartz veins and mineral deposits; Mineralium Deposita, v.7, p.202-216.
- Wallace, H., 1978, Geology of the Opikigen Lake Area, District of Kenora (Patricia Portion); Ontario Geological Survey Report 185, 58p. Accompanied by Map 2379, scale 1 inch to 1/2 mile.