GEOLOGICAL SURVEY OF CANADA

OPEN FILE 2164

Mineral deposits in the western Superior Province, Ontario (Field Trip 9)

edited by

J.M. Franklin
B.R. Schieders
E.R. Koopman

1991

Energy, Mines and Resources Canada
Énergie, Mines et Ressources Canada

Canada

GEOLOGICAL SURVEY OF CANADA

OPEN FILE 2164

MINERAL DEPOSITS IN THE WESTERN SUPERIOR PROVINCE, ONTARIO

[FIELD TRIP 9]

EDITED BY

J.M. FRANKLIN, B.R. SCHNIEDERS and E.R. KOOPMAN

With Contributions from:

F. Balint, F.W. Breaks, P. Brown, S.G. Charlton, A. Chong, J.M. Franklin, G.J. Hudak, E.R. Koopman, B. Kusins, H.C. Lockwood, W.H. McIlwaine, K. McNena, R.L. Morton, T.L. Muir, M.S. O'Brien, M.R. Patterson, K.H. Poulsen, B.R. Schnieders, P.W.A. Severin, R.J. Shegelski, R. Sim, M.C. Smyk, J.S Walker and H.R. Williams

8TH IAGOD SYMPOSIUM

FIELD TRIP GUIDEBOOK

8th IAGOD SYMPOSIUM

FIELD TRIPS COMMITTEE

Chairman and Guidebook Technical Editor:	**W.D. GOODFELLOW**
Guidebook Reviewer:	**K.H. POULSEN**
Field Trip Committee Members:	**B.I. CAMERON** **A. DOUMA** **C.W. JEFFERSON** **A.R. MILLER** **N. PRASAD** **D.G. RICHARDSON** **A.L. SANGSTER** **J.M. SHAW**
Word Processing:	**P. BROWN** **B. GIESE**

Table of Contents

MINERAL RESOURCES OF THE WESTERN SUPERIOR PROVINCE, ONTARIO

J.M. Franklin[1] and B.R. Schnieders[2]

Introduction (J.M. Franklin)

The purpose of this trip is to familiarize participants with some aspects of the metallogeny of western Superior Province, with a brief introduction to a few aspects of the north-western portion of Southern Province. We shall examine two principal types of Archean orebodies, volcanic-associated massive sulphide deposits in the Manitouwadge, Winston Lake and Sturgeon Lake areas, and the disseminated variety of lode gold deposits at Hemlo. In addition, a broad spectrum of Archean rocks will be examined, including some Algoman Iron Formation.

The Superior Province of the Canadian Shield is the largest Archean craton in the world, and contains an abundance of mineral riches. To the east and northeast, it is bounded by the Labrador Trough, a sequence of early Proterozoic sedimentary, volcanic and intrusive rocks, including an ophiolite suite (Cape Smith area, St. Onge et al., 1989) and a foreland thrust belt consisting of platformal to marginal marine sequence of sedimentary and volcanic rocks, including copious iron formation. To the southeast it is in fault contact with the Grenville Province, with enclaves of Mid to Late Proterozoic strata in a complexly deformed terrain of ortho- and paragneiss. To the south it is both overlain by and in tectonic contact with strata of the Southern Province. The latter contains several distinctive geological elements, described later.

SUPERIOR PROVINCE IN NORTHWESTERN ONTARIO

Superior Province has been divided most recently by Card and Ciesielski (1986) into subprovinces, each with distinctive lithological and structural characteristics. They classify these broadly as volcano-plutonic (usually termed "greenstone"), metasedimentary, and plutonic and high-grade gneiss terranes. The latter two are also classified as "gneiss belts" . The provincial boundaries are zones of distinctive change in lithology, and most are high-strain zones. Subprovinces are commonly juxtaposed along boundary faults; strata on either side are rarely correlatable, and each subprovince is quite distinctive from adjacent provinces. Card et al., (in press) suggest that some subprovinces could be compared to "suspect terranes".

Recently, Card et al., (in press) have reviewed the principal metallogenic aspects of Superior Province. Four principal groups of ore deposits have been identified (Card et al., in press; Franklin and Thorpe, 1982).

The principal groups of ore deposits are:

1) Deposits of mafic and ultramafic magmatic association, containing Ni, Cu, PGE's, Cr, Ti and V: These include komatiite associated Ni deposits, Cu-Ni deposits in layered tholeiitic sills, Ni-Cu-PGE deposits in differentiated gabbroic stocks, chromite in layered gabbroic intrusions, and vanadiferous ilmenite in anorthositic intrusions. Most of these are magmatic deposits, although some have been affected by hydrothermal redistribution of some sulphide minerals. In addition, genetically obscure, hydrothermal-vein deposits containing copper, minor nickel and other elements occur in association with many of the major layered gabbroic intrusions. Although we will not be visiting any of these, the Shebandowan deposit, about 100 km west of Thunder bay, is the largest deposit of its type in the Archean portion

[1] Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario. K1A 0E8

[2] Resident Geologists Office, Ont. Ministry of Northern Development and Mines, 77 Granville St., Toronto Ontario. P7C 5G6

of the Canadian Shield. At Lac des Iles, about 60 km NNW of Thunder Bay, a potentially productive PGE deposit occurs in a gabbro-troctolite intrusion.

2) Stratiform and related deposits in the volcano/sedimentary terranes: These include volcanic-associated massive sulphide (VMS) deposits, Algoma-type iron formation, polymetallic vein deposits (epithermal deposits primarily), and porphyry-copper occurrences in subvolcanic intrusions. The massive sulphide deposits are products of precipitation from high-temperature hydrothermal vents on or just below the seafloor. They occur primarily in bimodal island-arc like volcanic sequences. The Algoma-type iron formations include both volcanic-hosted chert-magnetite-hematite deposits, and those in greywacke and pelite, either in sedimentary basins that may have been marginal to, but slightly younger than, the volcanic belts, or in intra-volcanic sedimentary basins. The polymetallic vein deposits formed in subaerial volcanic sequences, and probably precipitated from hydrothermal fluids of similar origin to those which made the VMS deposits. The porphyry occurrences probably formed through collapse of a VMS-type hydrothermal system into a largely crystallized subvolcanic magma chamber.

The VMS deposits visited on this trip include the highly deformed and metamorphosed GECO deposit at Manitouwadge, the moderately metamorphosed and slightly deformed Winston Lake deposit, and the greenschist-metamorphosed, slightly deformed deposits at Sturgeon lake. The Manitouwadge deposits are the unusual in the Archean in that they have extensive iron formation associated with them. At Sturgeon Lake, minor iron formation occurs at Lyon Lake. This area also demonstrates the relationship of low-grade, but laterally extensive, porphyry-copper type mineralization in a sub-volcanic intrusion.

3) Lode Gold deposits, primarily near major tectonic zones: These include quartz-carbonate vein deposits, sulphide-rich iron formation-hosted deposits, and schist-hosted deposits associated with disseminated sulphides. The gold and other vein materials were introduced late in the tectonic history of most of these areas. Fluids migrated up the major faults, from deep crustal or even upper mantle rocks, and precipitated in dilational structures. Fluid precipitation may have occurred due to structurally induced depressurization.

The Hemlo deposit, visited on this trip, although classified as a "schist-hosted deposit with disseminated sulphides", is genetically somewhat obscure (see Smyk et al., and Brown et al., this volume). Its mineral assemblage is similar to that in some epithermal systems, yet it occurs in highly tectonized rocks.

4) Vein and pegmatite deposits associated with granitoid intrusions: These include rare-metal pegmatites, and Cu-Mo veins. Most of these formed from late-stage magmatic fluids. The best examples of these are in southeastern Manitoba, but several sub-economic fields occur in northwestern Ontario.

ARCHEAN GEOLOGY OF THE WESTERN WAWA AND CENTRAL WABIGOON SUBPROVINCES

This field trip will visit parts of the Wawa and Wabigoon subprovinces. Both of these contain an older sequence of mafic volcanic rocks, that possibly formed in continental or immature back-arc rifted sequences, and a younger sequence of mafic-felsic volcanic rocks, similar to that found in island arc sequences. The older mafic-dominated sequences contain iron formations (e.g. near Atikokan, Ontario) that were deposited in a shelf-like environment, different from the deeper-water, oxide-dominant Algoma-type iron formations associated with the younger volcanic sequences. Major oxide-facies iron formation also occurs along the margins of some metagreywake dominated subprovinces, as in the Quetico (north of the Wawa and south of the Wabigoon), that are adjacent to the volcanic-dominated subprovinces. The mafic-felsic sequences contain most of the massive sulphide deposits in Superior Province, including those at Sturgeon Lake and, although less well studied, probably those at Manitouwadge. The mafic-felsic sequences typically occur in the central regions of subprovinces, and contain tholeiitic rocks near their base, bimodal (basalt-andesite and dacite-rhyolite) volcanic sequences in their central portions, and are capped by volcanic-derived sedimentary rocks. Those which contain massive sulphide deposits generally have prominent subvolcanic intrusions near their base.

A second important component of the Archean architecture are the major fault zones, commonly termed "breaks", which may bound and/or transect most of the subprovinces. These faults are typically E-W, although important NE and NW faults transect some areas. Structural studies have demonstrated that these are the product of coaxial NS shortening of the Archean crust. A complex history of deformation and metamorphism, generalized as the "Kenoran Orogeny" by Stockwell (1964), has, as its earliest manifestation, developed a series of local basins, into which were deposited coarse, immature, generally shallow-water sedimentary strata and volcanic rocks with shoshonitic compositions. These sequences are commonly called "Temiskaming" or "Seine" strata, and rest unconformably on the earlier volcanic and sedimentary rocks which comprise most of the greenstone belts.

The faults have a long history of transpressive movement, extending well into the Proterozoic. Most of the lode gold deposits in Superior Province, are associated with them. The auriferous quartz carbonate veins are in structures that can be related to movement on these faults; Sibson et al., (1989) have demonstrated that the largest deposits or districts are associated with those major fault systems that have high angle reverse motion. Most of the gold deposits associated with major disseminated sulphides are also near these major faults. This class of deposit, together with the gold-bearing sulphidic iron formations, are more controversial in origin, but their close spatial association with the major faults is too common to be ignored. The Hemlo deposit, visited on this trip, is one such disseminated deposit of uncertain origin, but closely associated with a major shear zone.

WAWA SUBPROVINCE

The portion of the Wawa Subprovince to the north of Lake Superior has not been as comprehensively studied as many other areas of Superior Province. Card (in press) indicated that greenstone belts comprise about 35 % of the subprovince, and these consist of about 60 % tholeiitic and komatiitic mafic volcanic rocks, 25 % calc-alkaline and tholeiitic felsic and intermediate , 15 - 20 % sedimentary rocks and minor alkalic and shoshonitic rocks. The range of ages present in the volcanic rocks of the Hemlo area (2770-2695 Ma; Corfu and Muir, 1989a,b) is similar to that in other volcanic belts in Superior Province. The portion of the belt at Manitouwadge was, until recently, considered to be a small outlier of volcanic rocks at the northern edge of the Wawa Subprovince, immediately adjacent to the Quetico terrane. Recent mapping by mining company geologists and Williams and Breaks, (this volume) indicate a greater extent of supracrustal rocks. High-grade metamorphism and complex deformation have destroyed many of the primary volcanological and sedimentological features, making correlation at a regional scale very difficult.

In addition to the deposits visited on this trip, the western portion of the Wawa Subprovince contains many small, past - producing gold deposits, the largest Archean Ni-Cu deposit in the Canadian Shield, INCO's Shebandowan Mine, and numerous small porphyry Mo-Cu type occurrences. The Quetico Subprovince to the north contains rare-metal pegmatites in the Georgia Lakes district.

WABIGOON SUBPROVINCE

The Wabigoon Subprovince contains about 35 % supracrustal rocks, with very similar components to those described for the Wawa Subprovince. Ages for volcanic rocks within this subprovince range from 2999 Ma for lower bimodal sequences near Lumby Lake (Davis and Jackson, 1988), through mafic/ultramafic (Davis et al., 1988) and bimodal sequences at 2770 Ma (Anglin et al., 1988), to younger calc-alkaline to tholeiitic arc-type sequences, with a range of ages of 2745 to 2703 Ma (Davis et al., 1982). Included in the latter group of rocks is the South Sturgeon Lake volcanic belt (Morton et al., and Koopman et al., this volume), host to the massive sulphide deposits visited on this trip. Much younger Temiskaming-type rocks occur near the boundary between the Quetico and Wabigoon subprovinces. Ages range from 2698 to 2688 Ma (Davis et al., 1986; 1988).

The Wabigoon Subprovince has a wide variety of mineral deposits; in addition to the massive sulphide deposits visited on this trip, large gold deposits were mined near Geraldton and Beardmore, and the Atikokan area had two large iron deposits.

GEOLOGY OF SOUTHERN PROVINCE

The Proterozoic rocks around Lake Superior contain an excellent record of Early, Middle, and Late Proterozoic sedimentary, volcanic and plutonic history.

Deformation and metamorphism has not been intense in the area north of Lake Superior. To the south of the Lake, tectonism has been more pronounced, although variable. The rocks which will be briefly examined during this trip occur in two principal tectono-stratigraphic elements, the early Proterozoic strata of the Animikie and Mille Lacs groups and the Marquette Range Supergroup, and the Middle to Late Proterozoic Midcontinent Rift Sequence. A more complete field guide describing this area is available (Franklin et al., 1982); the stops and descriptions of the geology are an abbreviated version of that guidebook.

EARLY PROTEROZOIC

Rocks of this sub-era may be divided into three principal zones. The northernmost of these, including the Gunflint and Rove formations visited on this trip, and the Mesabi and Virginia formations in Minnesota, are a transgressive sequence of shale, greywacke and abundant iron formation, probably deposited on the margin to a rift, about 2.1 to 2.2 Ga. South of Lake Superior, correlative sequences in the Marquette Range Supergroup are lithologically similar, but more deformed and metamorphosed. The second, and somewhat younger zone (1.9-1.8 Ga) contains the island arc volcanic sequences in north-central Wisconsin. The volcanic-associated massive sulphide deposits of the Rhinelander-Ladysmith area occur in this sequence. The third, and youngest element is the Central Wisconsin rhyolite-plutonic complex (ca. 1675 Ma.), composed of anorogenic, subaerially-deposited rhyolite and high-level granite bodies.

MID TO LATE PROTEROZOIC

Rocks of these suberas are associated with the mid-continent rift zone, one of the biggest intracratonic rifts in the world. This rift in plan forms an inverted "U" shape centered on Lake Superior, and extend southeast from Sault Ste. Marie, Ontario to beneath the Michigan Basin, and to the southwest to northern Kansas. The strata of the Keweenawan Supergroup which form the rift fill (Table 1) consist of three stratigraphic elements: A lower sedimentary sequence, comprising the Sibley and Puckwunge Groups, (1339 Ma., Franklin, 1976) is composed of clastic, carbonate and shale units that were all deposited in shallow water. The middle volcanic sequence that dominates the Lake Superior area (1097-1110 Ma; Davis and Sutcliffe, 1986) contains about 12,000 m of tholeiitic continental flood basalt, including subordinate sedimentary and felsic volcanic strata. An upper clastic sequence, preserved only south of Lake Superior, is composed of sandstone. Prominent intrusions are associated with the rift, including the Coldwell Alkalic Complex, visited on this trip, the Prairie Lake Carbonatite Complex, the Logan sills and ring-dyke complexes, and the Duluth Complex.

METALLOGENY OF THE PROTEROZOIC ROCKS IN THE LAKE SUPERIOR AREA

The Early Proterozoic sequences contain relatively few mineral deposit types. The southern zone contains no mineral deposits. The central zone contains several volcanic-associated massive sulphide deposits, typified by the Crandon deposit (Rowe, 1982). This deposit occurs in a sequence of felsic and intermediate epiclastic rocks, and is generally similar to other Precambrian Cu-Zn deposits (Franklin et al., 1981). The northern zone, which includes the Animikie sequences of the Thunder Bay area, contains the very large Superior-type iron-formations. In addition, uranium-bearing phosphoritic conglomerate zones occur near the unconformity between several units of the Marquette Range Supergroup and the Archean basement (Cannon and Klasner, 1976).

The Keweenawan rift has a complex metallogeny. Deposits are related to various combinations of its magmatic, stratigraphic and structural elements, as follows:

A. **Magmatic mineralization includes:**

1) Cu-Ni sulphides occur in troctolite and gabbro of the Duluth complex and Crystal Lake intrusion. These are immiscible sulphide segregations; sulphur was derived from the adjacent Early Proterozoic shale (Mainwaring and Naldrett, 1977; Ripley, 1981);

2) Copper and platinum-group elements occur in immiscible sulphide segregations of the marginal gabbro phase of the Coldwell alkalic complex (Mitchell and Platt, 1982). This mineralization will be examined on this trip (Scheinders and Smyk, this volume).

3) Nb-U occurs in betafite in the Prairie Lake carbonatite.

Table 1 Stratigraphy of the Keweenawan Supergroup Lake Superior Area.

		Southern Minnesota	N. Wisconsin-W. Michigan	Isle Royale, Michigan	Northeastern Minnesota	Thunder Bay, Ontario	Eastern End, Lake Superior
Keweenawan	Upper	Hinckley Sandstone	Chequamegon Sandstone		Hinckley Sandstone		Jacobsville Sandstone
			Devils Island Sandstone				
		Fond du Lac Formation	Orienta Sandstone		Fond du Lac Formation		
		Solar Church Formation	Freda Sandstone				
			Nonesuch Formation				
			Copper Harbor Conglomerate	Copper Harbor Conglomerate			
	Middle	Chengwatana volcanic group	Portage Lake lava series ? ? ?	Portage Lake lava series ? ?	North Shore Volcanic Group		Michipicoten Island volcanic rocks Gros Cap, Mamainse Point, Alona Bay, Cape Gargantua volcanic rocks
			South Range volcanic rocks				
	Lower	Sioux Quartzite	Bessemer Quartzite	? ?	Puckwunge Formation	Osler Group	
			Barron Quartzite	? ?		Sibley Group	
APHEBIAN and ARCHEAN		Igneous and metamorphic rocks	Igneous and metamorphic rocks	? ?	Metasedimentary rocks	Metasedimentary rocks	Metasedimentary rocks

B: **Hydrothermal-Magmatic Vein and Disseminated deposits:**

Most of these deposits occur along the eastern shore of Lake Superior. Hydrothermal-magmatic deposits with porphyry copper affinities, occur near Batchawana Bay, north of Sault Ste. Marie, Ontario (Blecha, 1974; Armbrust, 1969). The Tribag and Jogran copper deposits are breccia-pipe deposits, containing disseminated and vein-type chalcopyrite and molybdenite in both matrix and rock fragments. The Tribag deposit contains abundant scheelite and minor wolframite. The uraniferous breccia pipes associated with the Coldwell complex may also be the hydrothermal-magmatic origin (Mitchell and Platt, 1982).

C: **High to medium temperature hydrothermal veins**:

Most of these are associated with contacts of intrusions and a wide variety of country rocks, and include:

1) Pb-Zn-Ag veins in Archean volcanic and sedimentary rocks, at the outer margin of the contact aureole of the Coldwell complex. The Deadhorse Creek veins are examples of this group (Franklin et. al. 1986).

2) Pb-Zn-Ag veins occur in northwesterly faults and fractures in Archean and Early Proterozoic rocks of the east shore of Lake Superior (Pearson, 1980). These faults are probably members of the axial "strike-fault" set, generated during Keweenawan rifting. The veins are spatially associated with Keweenawan diabase, and are in the area of lamprophyre and felsite dykes,

3) Ag and Ag-sulpharsenide veins occur in the N60°E fault sets near Thunder Bay, Ontario. (Franklin et al, 1986). These faults are also axial to the rift; the veins occur at the contact between the Rove shale and Logan sills or olivine gabbro dykes. All of the occurrences in this group were probably generated from fluids that formed through heating of the basal portion of the Proterozoic and Archean basement to the Keweenawan Rift, during its period of major volcanism and intrusion. Metalliferous fluids were subsequently expelled, along rift-related faults, to the margins of the rift, because of basinal loading.

D: **Low temperature, unconformity related Pb-Zn-barite (amethyst) and uranium veins and disseminations** occur along the unconformity between the Sibley Group and various Archean basement rocks. Both vein types have derived their metals from the local basement. An example of these will be examined on this trip (see Smyk and O'Brien, this volume). The Pb-Zn-barite veins formed at reduced-sulphur-rich pinch-out traps formed where the contact of the Rossport and Pass Lake formations intersects the unconformity (Franklin and Mitchell, 1977). These veins are locally amethyst-rich. The uranium veins occur in the 330° fault set which predates the N60°E axial faults; the former set may be associated with development of the Sibley basin (Yule, 1979). Uranium occurs in fluorapatite, and wallrocks to the veins have conspicuous hematite alteration, sericitization and chloritization.

E: **Low temperature bornite-chalcopyrite-native copper veins** were mined at the Coppercorp mine, near Mamainse Point, Ontario (Heslop, 1970) These veins occur in Keweenawan basaltic flows. Similar veins occur in the Osler volcanic rocks south of Nipigon and Rossport, Ontario.

F: **Red-bed copper deposits** occur in the Nonesuch shale, within the lower part of the upper sedimentary sequence of the Keweenawan Supergroup. These deposits, typified by the White Pine mine, are confined to a very narrow stratigraphic interval within the lowest part of the Nonesuch Formation, and consist of four laterally extensive blankets of disseminated chalcocite, digenite, bornite and chalcopyrite. Each blanket is approximately 30 cm in thickness (Leone et al., 1971). Native copper deposits occur primarily in conglomeratic interbeds and amygdaloidal flow tops within the Portage Lake Lava Series (Weege and Pollock, 1971). Less commonly, copper occurs in fissure veins. Both the red-bed and native copper deposit types result from low temperature fluids that were generated either during diagenesis (Brown, 1971) or by autometamorphism of the underlying volcanic rocks. Copper precipitation occurred either during diagenesis (Brown, 1971) or syngenetically (Rohrbacher, 1968; Jost, 1968; Burnie, 1970). The deposits of this group may have a similar origin to those of group E, described above.

FIELD GUIDE TO THE MANITOUWADGE AREA

H.R. Williams[1], F.W. Breaks[1], B.R. Schnieders[2], M.C. Smyk[2], S.G. Charlton[3] and H.C. Lockwood[3]

REGIONAL GEOLOGY OF THE MANITOUWADGE AREA (H.R. Williams and F.W. Breaks)

A highly strained unit of metamorphosed volcanic and sedimentary rocks, 1 to 2 km thick, occurs along the northern boundary of the Wawa Subprovince. The strata have an easterly regional trend but with unknown stratigraphic facing (Fig. 1). To the north, metasedimentary migmatite and granulite units, originally greywacke turbidite, predominate in the Quetico subprovince. In the vicinity of Manitouwadge, predominantly mafic volcanic rocks, rare intermediate to felsic breccia and epiclastic deposits and derived sedimentary rocks, including banded magnetite and silicate iron formation, occur in a reclined, east-northeasterly plunging regional scale synform. Along strike to the east and west, layered intrusions containing anorthosite, gabbro and peridotite occur within the metavolcanic sequence.

The supracrustal rocks lie within and are intruded by regionally developed, shallow to steeply dipping, foliated to gneissic tonalite, diorite, granodiorite, and granite bodies and attendant pegmatite and aplite. Early-formed, melanocratic gneiss locally displays magmatic textures, including igneous layering, orbicular structure and magma hybridisation. Commonly, darker phases are intruded by and therefore predate lighter phases (e.g. hornblendite and meladiorite occur as inclusions within later, paler rocks). Contacts with tonalitic gneiss or mafic metavolcanic rocks are commonly highly strained, even mylonitic. Within the Black-Pic Batholith to the south of Manitouwadge, flat-lying to outward-dipping, foliated to gneissic tonalite is interleaved with supracrustal rocks. Tonalite bodies are regionally developed, domal, sheeted masses; deeper levels are strongly foliated with a sub-horizontal planar fabric that exhibits a poorly developed, north-trending rodding and mineral elongation lineation. Upper structural levels of the tonalite masses are cut by abundant granitic sheets of pegmatite and aplite, and are generally massive except where in contact with supracrustal rocks, where they attain a gneissic foliation.

Most of these deformed rocks are generally of upper amphibolite facies grade of metamorphism. Sporadically developed alteration, producing garnet, cordierite, or anthophyllite (gedrite) in mafic rocks, is most typical of the "footwall" zone to the sulphide deposits in the Manitouwadge area. Deformed massive sulphides are spatially associated with units of sulphidized banded iron formation at the contact between mafic metavolcanic rocks and a metasedimentary package.

Bedding is recognized only rarely in the metasedimentary rocks within the Manitouwadge synform. In the outlying supracrustal rocks, extreme rarity of primary igneous and sedimentary structures distinguishes Manitouwadge area rocks from those in larger greenstone belts that typically exhibit well-preserved primary structures.

Five stages of deformation are recognized: two stages (D_1 and D_2) of regionally developed isoclinal folding, regionally developed Z-style folding (D_3), and two stages (D_4 and D_5) of localized shearing. A pervasive compositional layering, migmatitic veining in tonalite bodies, metasedimentary and metavolcanic strata, and a schistosity in less metamorphosed rocks (S_2), results from a combination of a second deformation stage (D_2) and a preceding deformation

---

[1] Precambrian Geology Section, Ontario Geological Survey, Ministry of Northern Development and Mines, 77 Grenville St., Toronto, Ontario. M7A 1W4

[2] Resident Geologist's Office, Ontario Ministry of Northern Development and Mines, Box 5000, 435 S. James St., Thunder Bay, Ontario. P7C 5G6

[3] Geco Division, Noranda Minerals Inc., Box 100, Manitouwadge, Ontario. P0T 2C0

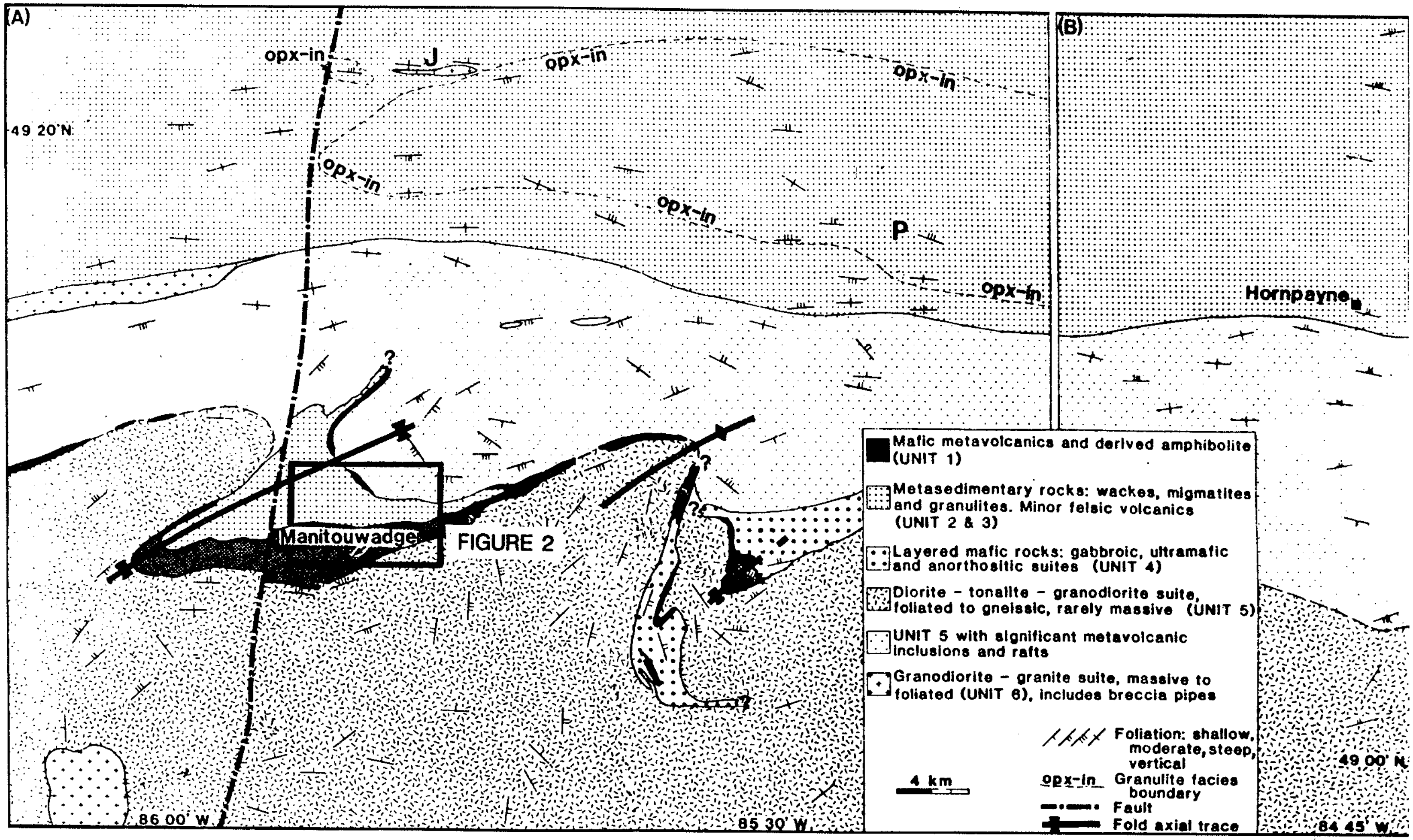


Figure 1 Regional geology of the Manitouwadge area (Williams and Breaks 1989).

stage (D_1), that both produced isoclinal folds. The strongly developed tectonic layering, obliterating primary structures, has an associated regionally developed rodding lineation and biotite elongation (L_2). Asymmetrical regional Z-style (D_3) folds deform D_2 planar and linear fabrics and migmatitic leucosomes, forming kilometre-scale, Z-style, shallowly northeast-plunging, asymmetric folds such as the Manitouwadge synform. D_4 deformation locally occurs along the southern limb of the Manitouwadge synform, centered on the ore zone where highly altered, micaceous rocks have focussed strain and further disrupted ore bodies already strained and reoriented by preceding deformation events. There, steeply west-plunging, asymmetric, Z-style D_4 folds fold the S_2 foliation and distort the L_2 rodding lineation. Increasing amounts of D_4 deformation, like that of D_3, have the effect of rotating L_2 lineations toward a more shallow east-plunging orientation.

A high state of deformation is recorded everywhere in supracrustal rocks, especially within the Manitouwadge synform, where stratigraphic facing is indeterminate and observed stratigraphic ordering and complexity is modified by ductile strain and repeated by folding and perhaps thrusting.

Sporadic Matachewan diabase dykes of northerly trend cut all rock types and the ore zone.

The main ore zone at Geco consists of a core of massive pyrite, pyrrhotite, chalcopyrite and sphalerite surrounded by disseminated pyrite, pyrrhotite and chalcopyrite in sericite schist. Subsidiary ore zones include the '4/2 copper', which lies along the northern contact of the sericite schist, and the '8/2 zinc' zone that is situated along the corresponding southern contact. The ore zones grade laterally into iron formation units. Iron formation units located within the 'Grey Gneiss Group' volcanic and sedimentary rocks are occasionally sufficiently rich in zinc to be mineable.

The orebodies trend easterly, dip steeply, or are vertical; they plunge moderately to steeply east, paralleling the L_2 lineation. The thickest portion of the ore zone is spatially associated with an easterly plunging, Z-style asymmetric fold that localized exceptionally good ore widths.

Alteration of mafic 'footwall' rocks to the north of the ore zone at GECO has produced much garnet and cordierite, along with aluminous amphiboles. 'Hanging wall' rocks are also altered, producing rocks enriched in silica, potash and alumina. These two alteration patterns are complementary, loss of alkalis in the 'footwall' being matched by the growth of muscovite in the 'hanging wall'. Most rocks, both above and below the ore zone, appear to be enriched in alumina. Alteration of mafic metavolcanic rocks also occurs west of Swill Lake, within the core and along the northwestern boundary of the Manitouwadge synform at Nama Creek.

Eastward-plunging rodding and mineral elongation lineation developed in both the 'hanging wall' and 'footwall' rocks become less steeply plunging towards the ore zone. The similarity between the rodding near the ore zone, and that observed regionally implies that the sulphide bodies in the ore zone have been subjected to intense strain, becoming elongate along steep to moderately eastward-plunging axes.

In the GECO and Willroy ore zones, abundant steeply west-plunging, asymmetric Z-style folds, when viewed on horizontal surfaces, have the appearance of kinks. These and associated extensional shears and crenulations, rotated boudinage and fish structures attest to the development of a dextral simple shear regime. The relative timing of the Z-style folds was during or after the formation of muscovite, after the production of much quartz veining and after the D_2 deformation that produced the ore rodding. East of the mine at Mose Lake, abundance of Z-folds increases to the north, towards the ore zone; in addition, the steep, easterly plunging L_2 lineation in mafic rocks clearly becomes less steep. The L_2 rodding here, and at equivalent exposures close to the ore zone, displays a variation in plunge, ascribed in part to the effects of D_3 and D_4 deformation.

Foliated to gneissic tonalite sheets occur subconcordant with the stratigraphy and are interpreted to be synchronous with the D_2 deformation. Northeast-trending pegmatitic granitoid sheets cut the ore zone at high angles and represent the latest products of felsic igneous activity.

LOCAL GEOLOGY OF THE MANITOUWADGE CAMP
(B.R. Schnieders, M.C. Smyk, S.G. Charlton)

Field stops which exemplify mineralization, alteration, structural and lithologic characteristics in the vicinity of past and present producing mines in the Manitouwadge camp are described below and shown on figure 2.

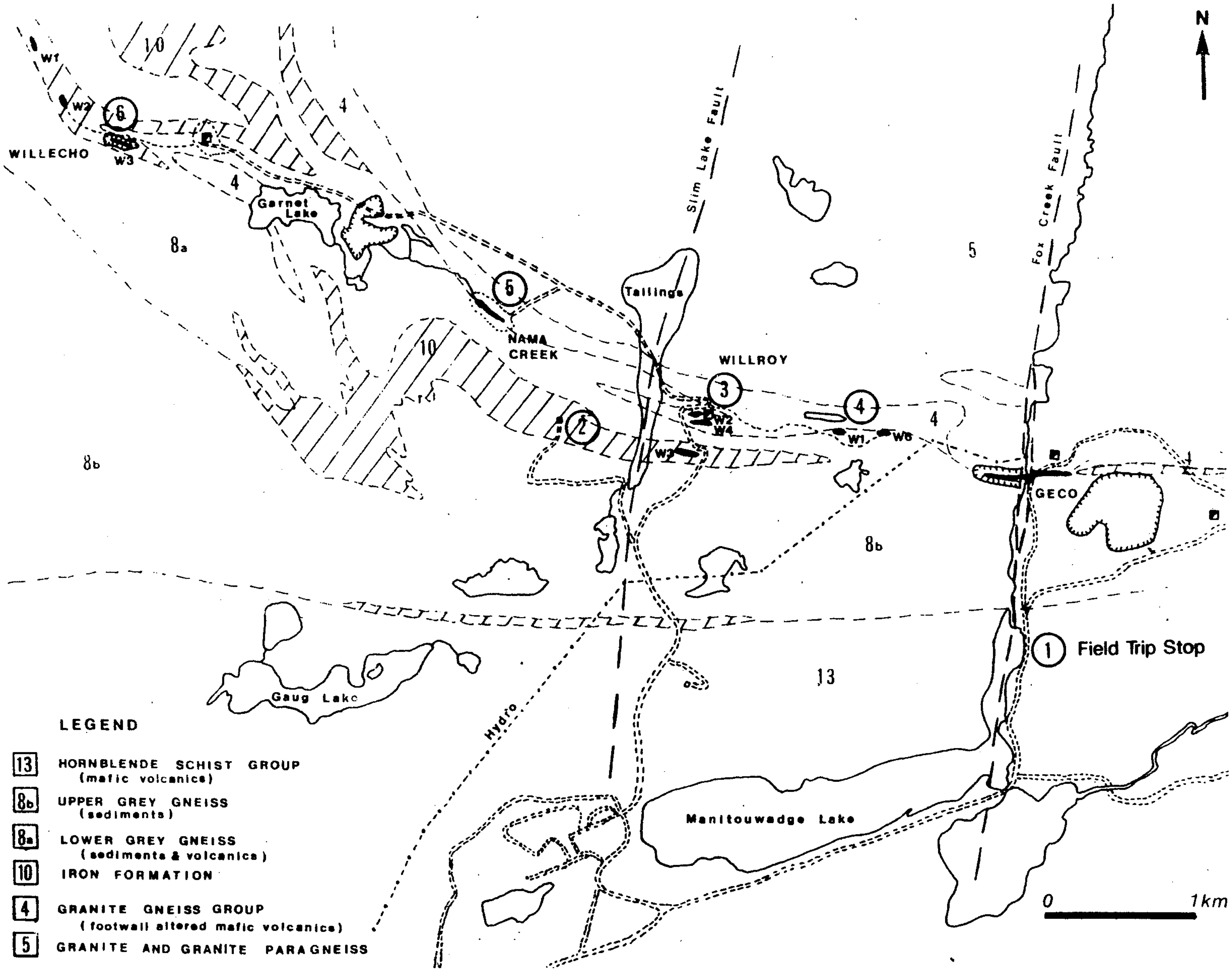


Figure 2 Local geology, south limb, Manitouwadge synform.

TABLE 1: Selected Whole-Rock Analyses, Manitouwadge Synform

Sample	1	2	3	4	5	6	7
SiO_2	50.6	64.8	59.10	51.25	74.09	76.64	84.53
Al_2O_3	14.61	12.82	9.84	13.08	12.03	11.70	7.58
Fe_{total}	10.57	8.76	11.43	20.56	5.04	2.03	1.77
CaO	8.75	3.17	10.68	1.28	0.21	0.36	0.19
MgO	5.74	2.09	4.48	6.83	3.27	0.44	0.37
Na_2O	2.29	2.29	1.81	0.15	0.21	3.12	0.32
K_2O	0.63	2.20	1.74	0.26	2.05	4.45	1.74
TiO_2	1.23	0.21	0.32	2.92	0.29	0.11	0.09
MnO	0.23	0.66	0.53	0.26	0.06	0.03	0.01
P_2O_5	0.27	0.05	0.06	0.58	0.06	0.00	0.01

Sample Descriptions:

1 hornblende schist, average analyses (STOP 1)
2 quartz-feldspar-biotite gneiss, average analyses (STOP 1)
3 mafic breccia, single analysis (STOP 3)
4 biotite-anthophyllite-cordierite gneiss, average analyses (STOP 3)
5 biotite-sillimanite gneiss, average analyses (STOP 3)
6 'biotitic quartzite', average analyses (STOP 4)
7 sericite schist, average analyses (STOP 4)

Rock types found within the Manitouwadge syncline and mine areas have been locally named (Pye, 1957; Milne, 1974; Freisen et al., 1982). Several of these names can be somewhat misleading in describing local rocks and their altered equivalents outside the mine area. To the north of the ore, the 'Granite Gneiss Group' is composed of mafic volcanic and minor sedimentary rocks, and mafic, inclusion-rich tonalite to granodiorite. The ore zone is contained within the 'Sericite Schist', in part a deformation-alteration- induced feature located along what once might have been a felsic volcanic unit. The 'Grey Gneiss Group' is composed of both metasedimentary and minor felsic volcanic rocks. The 'Hornblende Schist Group' comprises highly deformed mafic, pillowed metavolcanic rocks.

STOP 1: GATEHOUSE ROAD SECTION

This road section consists of schists (Hornblende Schist Group (13)) with rusty altered zones in the south, and sporadic tonalite sheets becoming more common northwards. At the north end metasedimentary rocks (Grey Gneiss Group (8)) dominate. The contact is gradational and marked by numerous tonalitic intrusions.

In the south section exposure of steeply dipping schist (Sample 1, Table 1) contains rare elongate pillow selvages along which alteration to biotitic, garnetiferous and anthophyllite-bearing assemblages, with local staurolite, has taken place. Pillows are elongate parallel with steeply plunging mineral and stretching lineation. Intensity of deformation at this location is typical of much of the Manitouwadge region, and primary structures are rarely preserved. Tonalite sheets and quartz veins are boudinaged, extension direction parallels that displayed by mineral lineation. The schists are laminated, perhaps due to shear.

Further north along the section there is a thin gossan which represents a continuous and extensive iron formation which occurs along or near the mafic metavolcanic-metasedimentary (13-8) contact. Further west this unit represents a banded quartz-magnetite iron formation commonly referred to as the South Iron Formation by GECO geologists.

Foliated metasedimentary rocks (Lower Grey Gneiss Group (8b)) are highly deformed at the north end of the section as the road bends around to the east. These rocks commonly described as quartz-feldspar-biotite gneiss (Sample 2, Table 1) show little, if any, evidence of primary structures such as bedding or grading. Across the road, the northeast end of Manitouwadge Lake and Fox Creek mark the north-trending Fox Creek Fault.

STOP 2: MICROWAVE TOWER IRON FORMATION

At the summit of this hill, the east-trending ore zone along the south limb of the Manitouwadge synform may be picked out by observing the various open cuts and headframes. These outcrops demonstrate the variability in composition and appearance of iron formation in the region along strike from the ore horizon. Some are thinly laminated magnetite-quartzite rocks, others are sulphide-rich assemblages. The light coloured layers are generally wider than the dark layers and are made up of recrystallized quartz. In the centers of the wider, light coloured layers the quartz is milky, while in the thin layers it is smoky due to tiny inclusions of mafic minerals. The dark layers are made up of quartz and magnetite commonly with varying amounts of hornblende, actinolite, cummingtonite, biotite, epidote, diopside, carbonate and garnet. These iron formations locally contain pyrite and pyrrhotite and sometimes extend laterally into massive sulphide horizons (e.g. Willroy #3 Zn Deposit). The iron formation has been locally intruded by tonalite dikes. Isoclinal folds in laminated iron formation may plunge steeply to the east, but are variable in form and attitude. The iron formation is hosted within metavolcanic and metasedimentary rocks (Upper Grey Gneiss Group (8a-8b)). On the roadside leading up to this location, massive to foliated felsic volcanic rocks are exposed.

STOP 3: WILLROY MILL SECTION

Just north of the old mill site are glacially polished exposures of mafic breccia, consisting of pale, angular fragments set in a chloritic matrix (Sample 3, Table 1). This brecciation may represent a primary rock such as a pillow or flow-top breccia, but a tectonic origin has not been discounted. Deformation intensity is surprisingly low here, but fragments are elongated down an easterly plunging lineation.

Mineral assemblages include diopside, garnet and epidote, indicating recrystallization of calcic bulk rock compositions. These assemblages will contrast with those seen in later stops, and this alteration is not typical of the"footwall-type" hydrothermal alteration in the mining camp.

North towards the railway cut, numerous loose boulders of ore remain in the surface rubble. In the cut, an excellent exposure of typical "footwall" alteration is present. These garnetiferous and amphibole-rich rocks represent hydrothermally altered and deformed mafic metavolcanic and minor metasedimentary rocks (Granite Gneiss Group (4)) which comprise the stratigraphic footwall (structural hangingwall) to the Willroy #2, #4 and #5 ore zones. The rocks (Samples 4 and 5, Table 1) are enriched in MgO, Cu and Zn and depleted in Na_2O and CaO. The enrichment of magnesia and alumina at the expense of alkalis produces assemblages that include anthophyllite (gedrite), biotite, garnet, sillimanite and cordierite with local sericite. The "footwall" rocks are foliated, producing slabby outcrops, with surfaces that exhibit strongly developed L_2 rodding and mineral lineation. Ore zones in the camp plunge parallel to this 40° to 50° easterly plunging lineation.

STOP 4: WILLROY #1 ORE ZONE SECTION

This section is described from south to north, going from hangingwall to footwall across the ore zone. The first stop is in mildly foliated aplitic felsic rocks cut by a discordant sheet of similar quartzo-feldspathic composition. This location has been historically identified as a laminated biotitic quartzite (Sample 6, Table 1) of the Grey Gneiss Group (8). Other interpretations are suggestive of an altered and deformed felsic metavolcanic rock or an igneous intrusive rock. This rock directly overlies the ore horizons at both GECO and Willroy. Minor garnet is present in this unit.

These rocks are inhomogeneously deformed. The outline of the discordant, 20 cm wide intrusion can be traced across the foliation in the host until it reaches a rusty zone, at which point, the intrusion thins dramatically and veers off to the east. The outcrop has been intruded by both diorite and felsic dikes.

To the north, this leucocratic unit develops an intense schistosity and is compositionally a rusty quartz-sericite schist now part of the Sericite Schist Group (2), containing lensoid quartz masses. The rock (Sample 7, Table 1) characteristically exhibits a platy parting of muscovite separating quartz-feldspar lithons. Steep, westerly plunging Z-folds parallel a weak to strongly developed crinkle lineation that deforms the pervasive L_2 mineral lineation.

Up the road and further north into the section the schist changes to one dominated by quartz, sericite and elongate clots of sillimanite. Further down the road you can observe on the distant ridge the open cut between #1 and #4 shafts at the GECO Mine. You are standing on the extension of the mineralized horizon, which has been mined beneath your feet.

Further north, the sericite schist becomes even more siliceous and is exposed on the eastern end of the fenced area. The Willroy #1 ore zone was a chalcopyrite-rich (2-3%) disseminated stringer zone. The rock is Na_2O-depleted and relatively K_2O-enriched. In the open stope a moderately southwesterly plunging crinkle lineation can be seen to overprint the regionally developed L_2 lineation. This crinkle is associated with minor Z-style folds that are only found immediately adjacent to the ore zone.

The entire, well-exposed transition from relatively massive (volcanic?) rock to quartz-muscovite schist is considered to represent a gradient in strain and alteration.

Further north, a major change in rock type occurs just to the north of the ore zone. A sharp contact with altered footwall metavolcanic rocks (Granite Gneiss Group (4)) occurs and these mafic granoblastic rocks are coarse-grained, poorly foliated, steeply dipping and composed of garnet, cordierite, anthophyllite (gedrite), staurolite and biotite. Porphyroblastic garnet and cordierite, contain well-developed internal planar fabrics that have subsequently been overprinted by recrystallization in the matrix. As a result of the recrystallization, "footwall" rocks do not at first sight appear to be as deformed, but quartz veins in this material are highly elongate and boudinaged and a strongly developed mineral lineation plunges east.

STOP 5: BIG NAMA ORE ZONE

At this flooded open cut, the Big Nama ore zone was situated at the same stratigraphic horizon (8/4) as the Willroy #1 ore zone. The sulphide ore zone and adjacent sericite schist has been intruded by one of a suite of northeast-trending pegmatite sheets. These pegmatite sheets are interpreted to be responsible for coarsening the pyrite-, pyrrhotite-, sphalerite- and galena- bearing ore by recrystallisation. The mineralized sheet itself is lenticular in nature, likely the product of strain. On the north side of the open cut typical garnet-anthophyllite footwall alteration of mafic metavolcanic rocks is preserved. Note that the planar fabric in these rocks is considerably less steeply dipping than at the preceding stops.

STOP 6: WILLECHO #3 ORE ZONE SECTION

This open pit is the surface expression of the shallowly north-dipping Willecho #3 ore zone. Several mineralized zones varying in width from less than 1 metre to greater than 3 metres have a variable, lenticular to sheet-like form elongate down the lineation, which is here plunging shallowly to the east. The ore zones, hosted by iron formation and sillimanite- and garnet-bearing schists, are exposed on the pit wall. The coarse-grained, often zinc-rich ore zones contain pyrite, pyrrhotite and sphalerite and are perhaps concordant with enclosing mafic "footwall" rocks. Evidence of strong deformation, even in this area close to the hinge zone of the Manitouwadge synform, is given by the presence of mylonitised granitic sheets paralleling the general foliation a few hundred metres to the west of this pit.

GEOLOGY AND EXCURSION GUIDE TO THE GECO MINE

(H.C. Lockwood)

INTRODUCTION

This guide briefly describes the geology and regional setting of the Geco Mine. The mine is operated by Noranda Minerals Inc. - Geco Division; it is located at Manitouwadge, Ontario. General descriptions of the geological features likely to be observed during the underground tour of the mine by participants in the 1990 IAGOD field trip to the area are included.

Readers are referred to the definitive descriptive report on the Geco mine and surrounding area ("The Geology of the Geco Base Metal Deposit"; R.G. Friesen, G.A. Pierce, and R.M. Weeks, 1982) for more comprehensive details of the mine and regional geology.

HISTORY OF THE GECO DEPOSIT

The Geco deposit is in the Manitouwadge mining camp, which is situated in the Thunder Bay Mining District. The mine is located 200 miles east of Thunder Bay, Ontario (Fig. 1).

J. E. Thompson, a geologist with the Ontario Department of Mines, located gossan zones on the northeastern arm of Manitouwadge Lake during a regional mapping program in 1931. In 1953, three prospectors from Geraldton, Ontario staked the nucleus of what was to become the Geco property. Following a massive staking rush to the area, which also resulted in the discovery of the Willroy, Nama Creek, and Willecho deposits, the Geco mine was brought into production in 1957 with an initial mining rate of 3300 tons per day and an ore reserve of 14,000,000 tons.

Geco has been in continuous production since 1957, and has produced over 47,000,000 tons of ore grading 1.9% copper, 3.8% zinc, and 1.7 ounces per ton silver. Minor amounts of gold, cadmium, and bismuth are currently recovered from the ore; the orebody contains an average of 0.30% lead, which has been recovered in the past, but it has not been economically feasible to recover lead since 1988 and the lead circuit in the mill has been removed.

The mine currently operates at 3700 tons per calendar day with a staff of 480 people. Most of the production is from extraction of the large pillars left by primary mining of the deposit. The property is serviced by two shafts and 19 levels; the deepest mining level is 3850 feet below surface. The principal mining method used is blasthole stoping with immediate backfill (which is required to support the relatively weak walls). Most ore is removed from stopes by multiple transverse slusher scrams feeding into a central orepass system. A quarry directly above the orebody provides backfill, which is fed directly to individual stopes by a complex system of raises. Some trackless methods are also in use.

The concentrator produces copper and zinc concentrates, which are shipped to smelters at Noranda and Valleyfield, Quebec.

As of January 1, 1990, the Mineral Inventory at Geco stands at 11,100,000 tons grading 1.62% copper, 2.94% zinc, and 1.14 ounces per ton silver.

Production from the Willroy, Willecho, and Nama Creek deposits totalled 7,900,000 tons grading 0.94% copper, 4.87% zinc, and 1.79 ounces per ton silver. Production from these deposits ceased in March, 1977. Noranda Minerals Inc. - Geco Division now holds the rights to all three of these properties. The general location of these deposits relative to Geco is given in figure 2.

REGIONAL GEOLOGY

The Manitouwadge mining camp is located in Superior Province, near the northern edge of the Abitibi - Wawa metavolcanic belt (Fig. 1), near the interface between this belt and the Quetico metasedimentary belt. The interface between these two Archean belts is transitional and their age relationship is controversial (see Williams and Breaks, this volume). Regional metamorphism has produced gneisses and schists of the almandine - amphibolite facies. Recent detailed mapping of the Manitouwadge area has subdivided this sequence into four conformable east - west trending units. From north to south these are: the Quetico paragneisses; a granitized zone of interbedded felsic paragneisses and mafic orthogneisses; the Manitouwadge Mine Series; and, to the south, a granitized zone of felsic orthogneisses and paragneisses (Fig. 2).

This strata in these four units are folded, developing the easterly plunging Manitouwadge Synform and other less well defined parallel structures as shown in figure 2. The Manitouwadge Synform contains all the known orebodies in the area. Regional mapping indicates that the Manitouwadge Synform is an overturned portion of the Mine Series Stratigraphy; this is supported by metal zonation within the Geco deposit and by the presence of an extensive zone of hydrothermal alteration lying to the north of the Geco deposit.

Within the Manitouwadge Synform, the Mine Series rocks can be subdivided into four groups (Fig. 3). From north to south, these are: the Granite Gneiss Group; the ore - hosting Sericite Schist Group (too

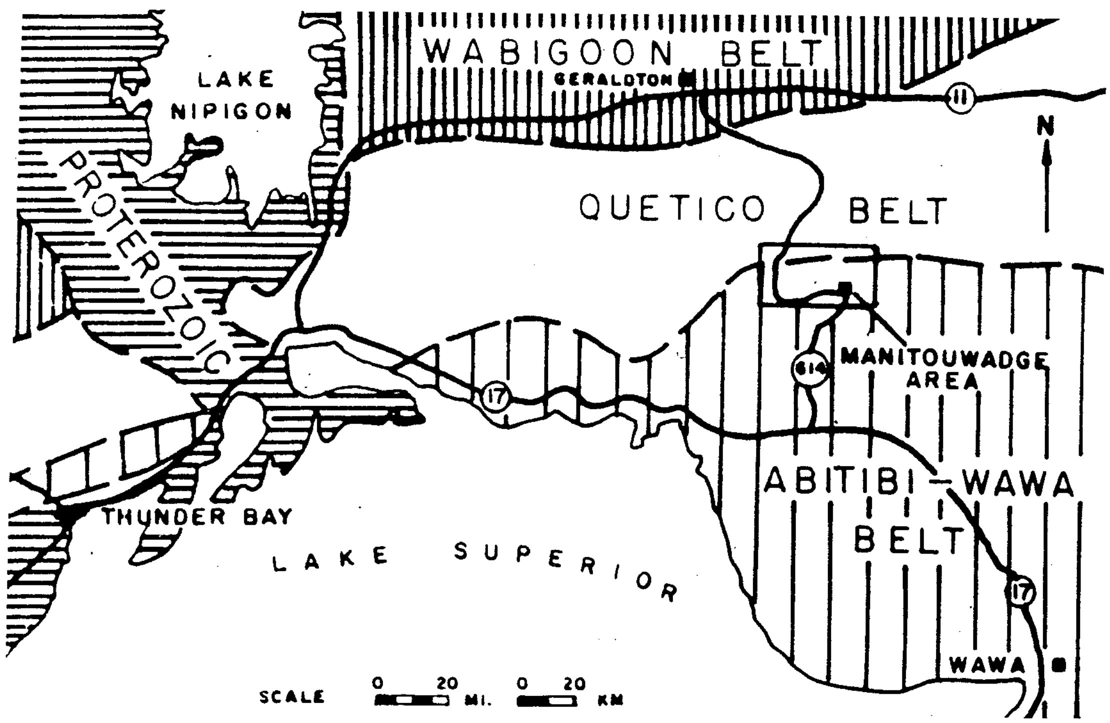


Figure 1 Location Map of the Manitouwadge Area.

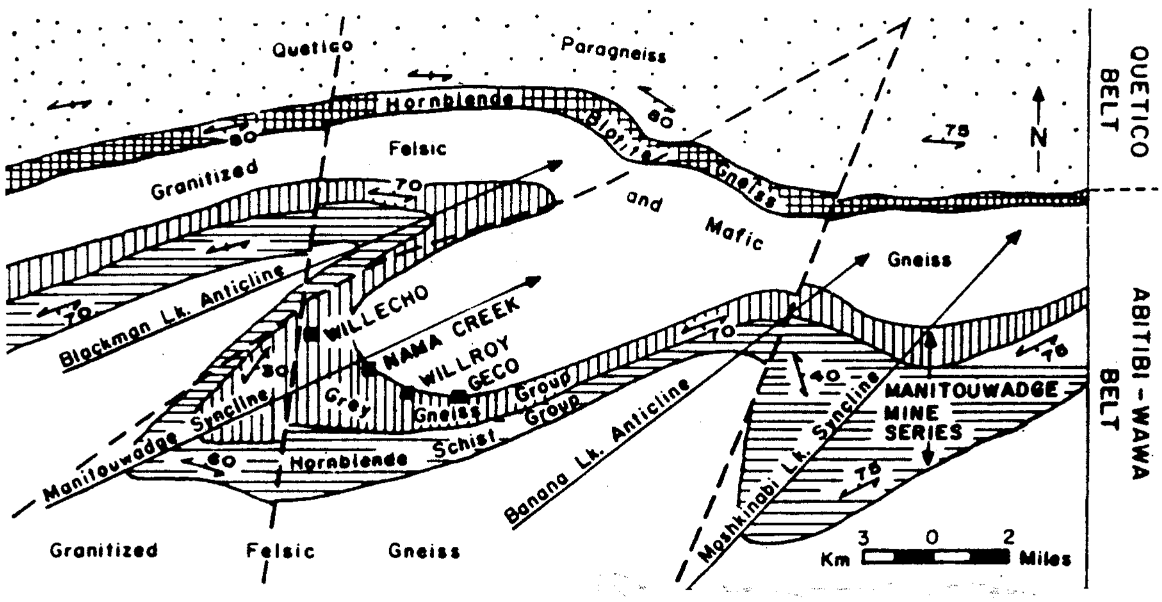


Figure 2 Regional Geology, Manitouwadge Area Geco mine.

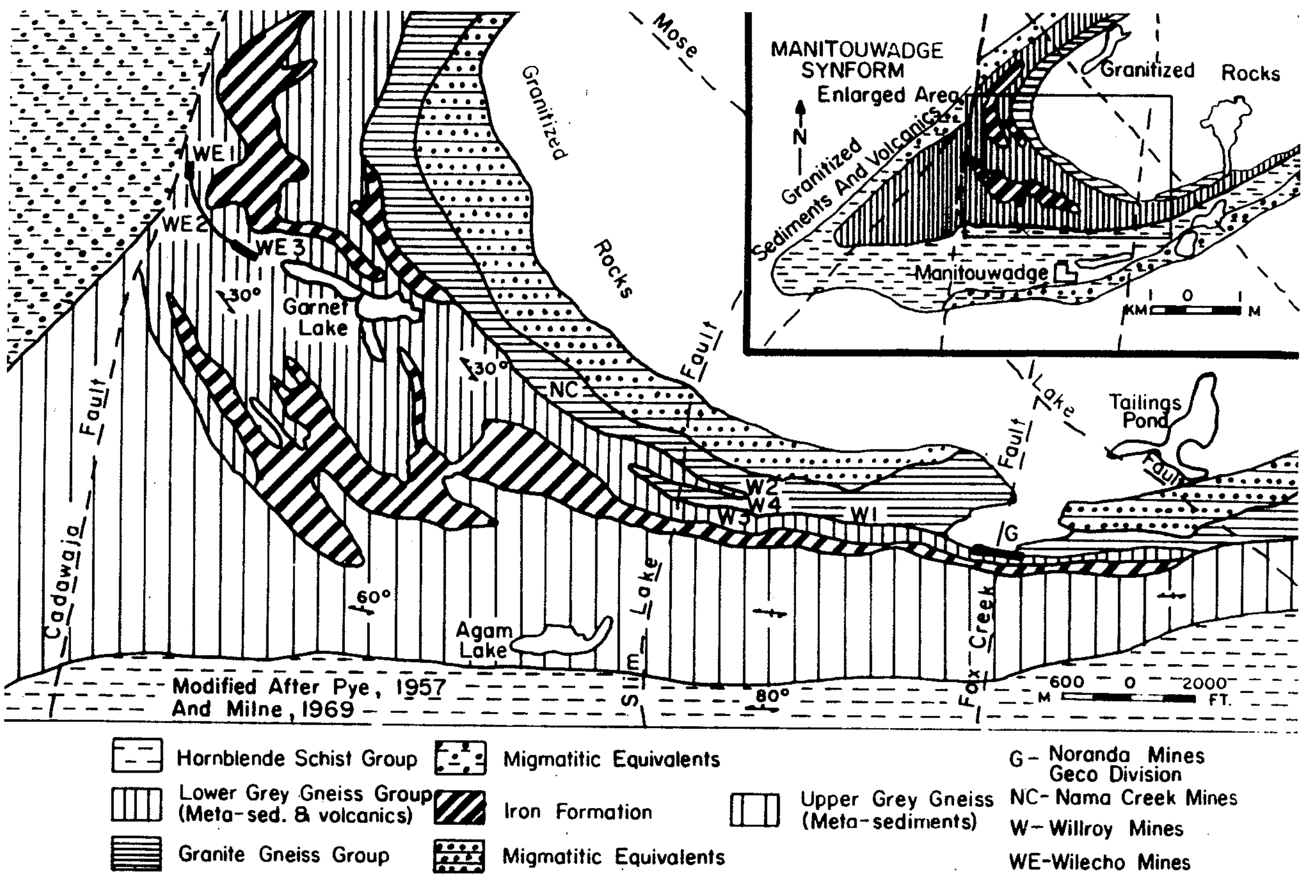


Figure 3 Geology of the Manitouwadge Synform between Willecho and Geco.

narrow to be shown at the scale of Fig. 3); the Grey Gneiss Group; and the Hornblende Schist Group.

Known regional intrusive rocks include (from oldest to youngest): concordant metagabbro bodies, found mainly in the Hornblende Schist Group; pegmatite and granite dykes related to phases of granitization; and a series of Keweenawan diabase dykes, which are structurally controlled by the major fault systems which trend north - south, northeast, and northwest and appear as strong regional lineaments showing only limited displacements.

The Geco orebody is located on the south limb of the Manitouwadge Synform.

MINE GEOLOGY

All of the producing ore zones at Geco are found in the Sericite Schist Group of rocks, which are bounded on the north by the Granite Gneiss Group rocks and on the south by the Grey Gneiss Group rocks. The ore zones trend easterly and dip vertically. Figure 4 shows the general disposition of the 3 major ore zones at Geco and their position within the Mine

Series stratigraphy.

The main orebody consists of a core of massive pyrite, pyrrhotite, chalcopyrite, and sphalerite surrounded by disseminated pyrite, pyrrhotite, and chalcopyrite in sericite schist. It plunges to the east at approximately 35 degrees, and is controlled on its lower and western limits by a similarly plunging "Z" shaped drag fold (locally known as the Geco Drag Fold). The main ore zone is thickest near the drag fold, and rapidly thins eastward and upward.

Other ore zones at Geco include the 4/2 copper zone, which lies along portions of the Sericite Schist - Granite Gneiss contact and the 8/2 zinc zone, which lies along portions of the Sericite Schist - Grey Gneiss contact. Descriptions of these zones are included in later discussions on stratigraphy. As well, iron formations in the Grey gneiss group are occasionally zinc bearing; one such zone - the ZIF zone - appears to be partially mineable.

The Geco Drag Fold, while previously thought to be a major structure complementary to the Manitouwadge Synform, is now interpreted to be associated with only the massive sulphide portion of the main orebody, in that it developed during regional deformation as a compensation feature around the excessive volume of sulphide material. Of course, it is still considered to be a major structural feature from an economic point of view because its development played an important part in the localization of exceptionally good ore widths. It is typically asymmetrical and undulates down - plunge at an average angle of 35 degrees to the east.

Several other features in and around the orezone are believed to have developed contemporaneously with the drag fold. These are: minor "Z" and "S" - shaped drag folds; boudinage, best displayed by the massive ore; and tight crenulations in the sericite schist unit.

Pegmatite dykes are common and strike north - east through the entire mine stratigraphy; within the sericite schist, these have been deflected by bedding plane foliation, resulting in noticeable thinning. A set of small - scale west plunging "Z" - shaped drag folds coexist with these dykes and their interference with the overall dominant east plunge in the mine area is thought to be the cause of undulation in the plunge of the Geco Drag Fold. Only remnant pegmatite material is found within the massive sulphide zone.

Quartz diorite bodies are found throughout the mine stratigraphy and are considered to represent an older set of intrusives. They are narrow, continuous, dyke - like rock units which are subparallel to bedding plane foliation. Only remnant quartz diorite material is found within the massive sulphide zone.

Diabase dykes cut all the mine rocks.

In cross section, the thickest part of the main ore zone is found at the lower end near the drag fold. Boudinage of the massive sulphides is evident as the sulphides thin upwards. Of particular note is the distribution of disseminated sulphides; these are best developed at the bottom and on the south side of the massive zone.

Intense deformation and metamorphism has severely disrupted and complicated original stratigraphy in the vicinity of the Geco Drag Fold. The disseminated and massive portions of the main orebody show evidence of being folded by the drag fold. In the sericite schist, the folding is typically isoclinal and complex. The disseminated copper zone is considered to closely follow original bedding. Quartz diorite dykes, which crosscut this primary feature are tightly folded and often transposed into bedding plane foliation. In places, younger pegmatite dykes further complicate the structure with their interfering west plunging folds.

The entire sequence is believed to represent separate cycles of mineralization in a predominantly volcano - sedimentary depositional environment, which has subsequently undergone intense regional metamorphism and deformation. The Sericite Schist Group is considered to be an ore - related alteration halo overlying an altered mafic basal formation - the Granite Gneiss Group - which could represent an altered sea floor assemblage of basaltic composition. The Sericite Schist Group is contained within a package of relatively "unaltered" rocks that make up the Grey Gneiss Group. The lower portion of the Grey Gneiss Group would be time - equivalent to the ore stratigraphy. A schematic stratigraphic column illustrating this interpretation is shown in figure 5.

ORE STRATIGRAPHY

Figure 4 shows the mineral assemblages found

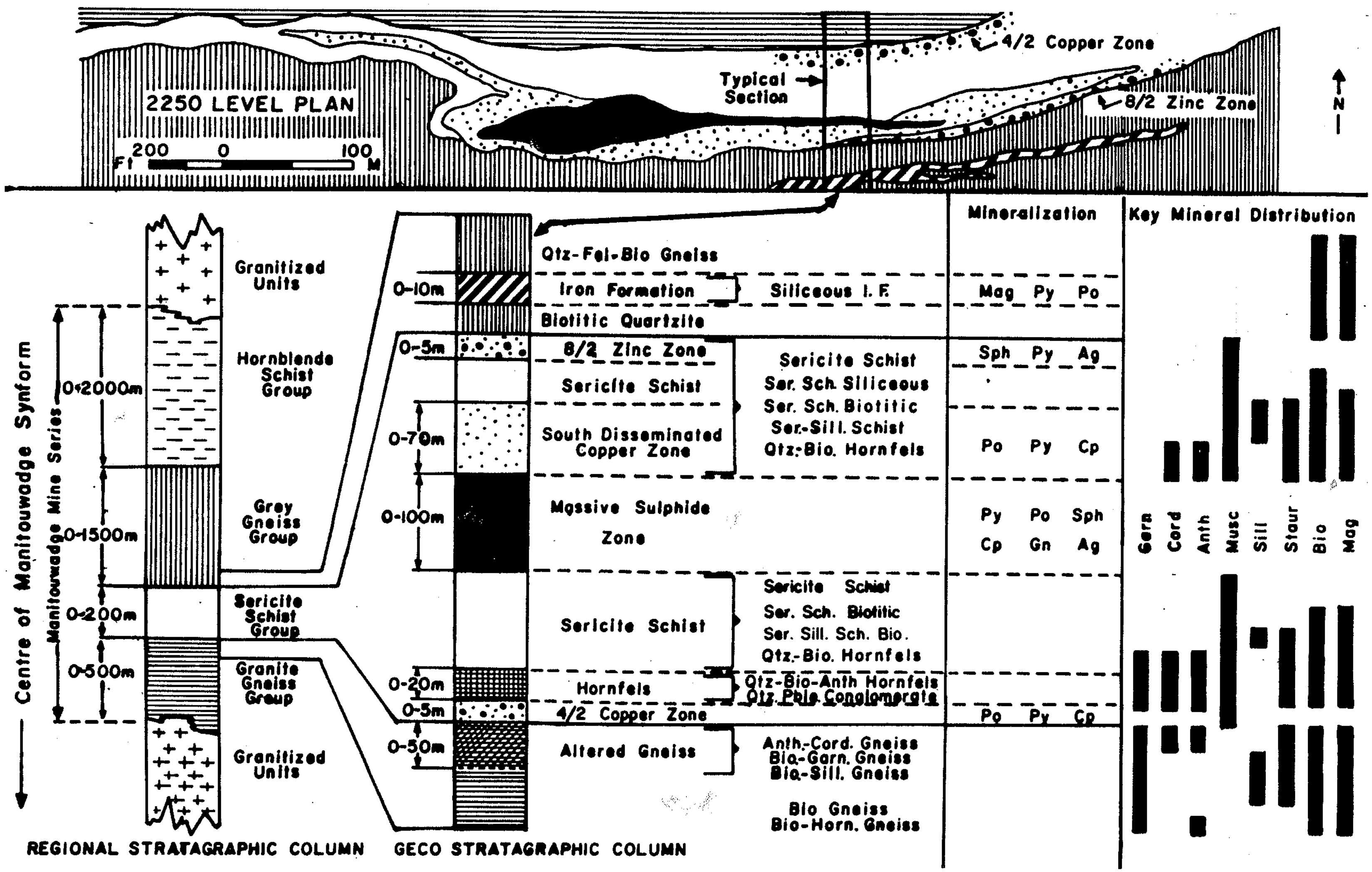


Figure 4 Schematic Stratigraphic column Illustrating Generalized Relationships of Ore Zones in the Geco Mine Area.

GECO
SCHEMATIC STRATIGRAPHIC COLUMN

(not to scale)

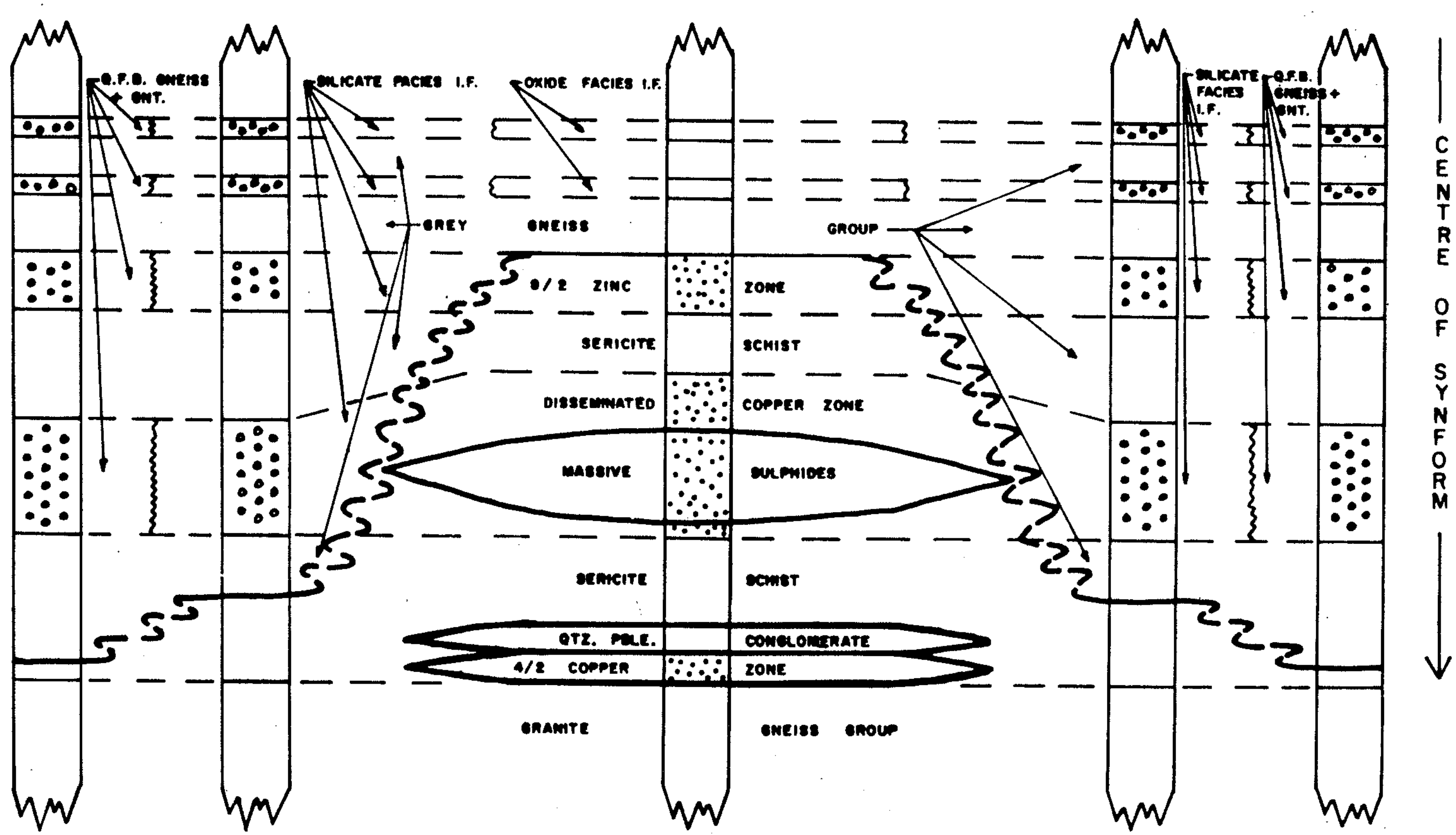


Figure 5 Geco Schematic Stratigraphic Column.

in the ore stratigraphy.

The Granite Gneiss Group is a diverse group of medium to coarse - grained mafic to intermediate rocks containing biotite - hornblende gneiss, biotite - sillimanite gneiss, and biotite - anthophyllite (Gedrite)-cordierite gneiss, all of which can be found to be locally garnetiferous. The top of this last unit generally marks the Sericite Schist - Granite Gneiss Group contact and is locally conformably overlain by the 4/2 Copper Zone. This zone is essentially a fine to medium grained disseminated copper zone containing chalcopyrite, pyrrhotite, pyrite, and wall rock fragments. Sphalerite and the zinc spinel gahnite are common along its eastern and western extremities. The zone is locally sheared and folded, displays pinch and swell along its length, and varies in width from a few inches to fifteen feet.

In many places, the 4/2 Copper Zone is overlain by a 9 foot thick band of weakly mineralized quartz pebble conglomerate. The quartz pebbles are discus shaped, up to 10 inches in diameter, and are contained in an anthophyllite - rich matrix (Bakker, 1979). The current belief is that this pebble conglomerate may represent a tectonically disrupted, poorly developed iron formation. Below the 4/2 Copper Zone, the Sericite Schist Group consists of an interlayered series of sillimanitic, biotitic, and siliceous sericite schists with the more biotitic and sillimanitic members found towards the quartz - pebble conglomerate. Above the 4/2 Copper Zone, the more siliceous units containing up to 70% quartz and 30% muscovite are more abundant. These units often contain erratic amounts of pyrrhotite, pyrite, and chalcopyrite and achieve their highest copper grade near the overlying main orebody, where they referred to as part of the main ore zone in mine terminology.

The main orebody is composed of a massive pyrite - pyrrhotite - sphalerite - chalcopyrite zone overlain by disseminated sulphides. This massive zone consists of coarse crystalline sulphides and varies in width from several inches to over 150 feet. It also contains recoverable amounts of lead, gold, silver, cadmium, and bismuth as well as minor cassiterite.

The massive zone has a complex mineral paragenesis. The sulphides show evidence of having been remobilized and recrystallized during metamorphism. They are generally conformable to the surrounding sericite schist. Remobilization is indicated by local small discordancies and by numerous remnants of wallrock and intrusives within the sulphides. Figure 6 shows the metal distribution within the massive sulphide portion of the main orebody.

The overlying disseminated copper zone is essentially siliceous sericite schist containing pyrite, pyrrhotite, and chalcopyrite, which locally forms near massive concentrations. It grades laterally into silicate - oxide facies iron formation outside the sericitic alteration halo, but immediately adjacent to the iron formation, becomes noticeably zinc - rich and copper - poor.

Upwards in the stratigraphy, the main orebody grades into poorly mineralized sericite schist which is typically light coloured and which contains less biotite and sillimanite than the units below the main ore zone. These schists grade upwards into the overlying 8/2 Zinc Zone, which is essentially sericite schist containing disseminated sphalerite, pyrite, and minor amounts of galena. It varies in thickness from 3 to 15 feet and marks the upper limit of the Sericite Schist Group. It also grades laterally outside the alteration halo into silicate - oxide facies iron formation.

Throughout the Sericite Schist Group, erratic lenses and pods of medium to coarse grained biotite - anthophyllite - cordierite hornfels can be found. The geometry and distribution of these units is not fully understood.

The lower Grey Gneiss Group, as exposed in the mine, is characterized by grey to buff coloured, well banded quartzites and grey, fine to coarse grained foliated quartz - feldspar - biotite gneisses.

Interbedded with these quartzites and gneisses are the silicate facies iron formations on the flanks of the ore stratigraphy. As previously mentioned, these probably represent lateral equivalents of ore horizons. They are characterized by chlorite, garnet, hornblende, pyrite, pyrrhotite, and magnetite. Oxide facies iron formations mainly overlie the ore horizons in the mine area. Their very low alumina content indicates that they are probably a chemical sediment, likely representing a late stage hydrothermal event similar to that which deposited the ore zones below, but generally with insufficient copper and zinc to make them economic. Laterally, they also grade into silicate facies iron formation. Although both types of iron formation are generally devoid of copper and zinc, ore grade widths of sphalerite - bearing silicate iron formation (ZIF zone)

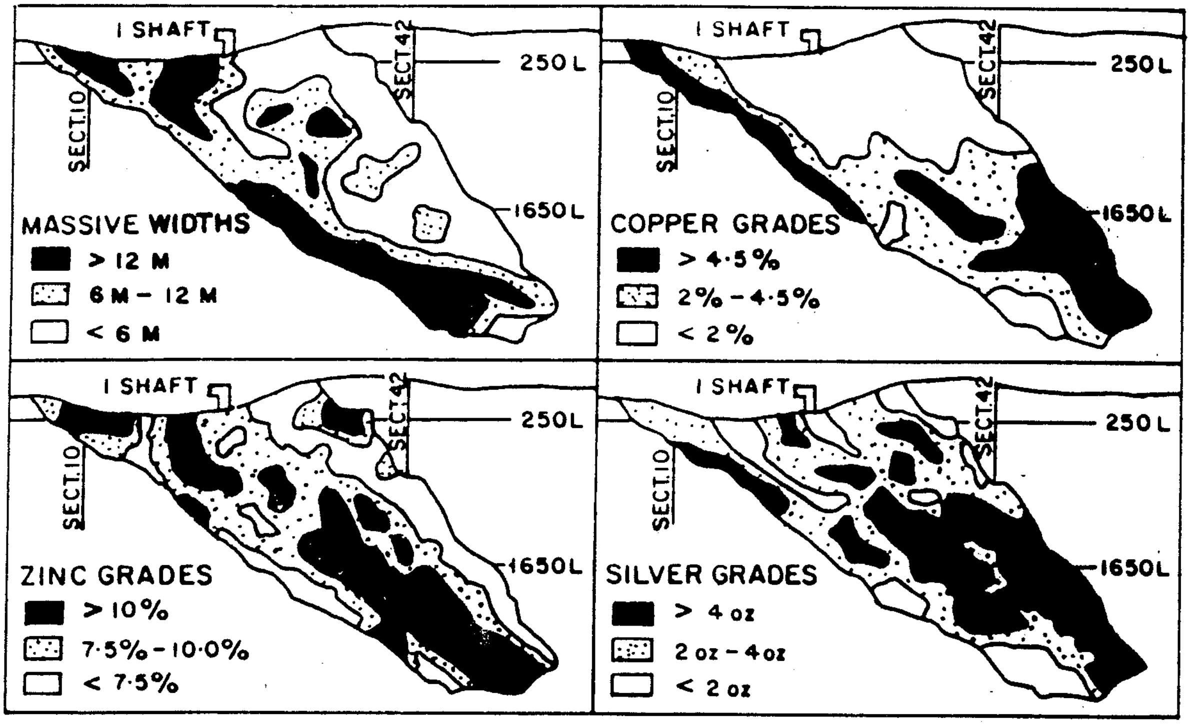


Figure 6 Widths and Ore Distribution of the Massive Ore Zone.

have been encountered in the lowest levels of the mine.

Outwards from the mine area, both types of iron formation commonly grade over a short distance into garnetiferous quartz - feldspar - biotite gneisses.

The upper members of the Grey Gneiss Group and the Hornblende Schist Group are not discussed. Friesen et al. (1982) give a full discussion of these units.

Figure 7 shows the major element distribution across the mine strata. The main features are: the mafic nature of the Granite Gneiss Group; the felsic nature of the Sericite Schist and Grey Gneiss Groups; the depletion of calcium and sodium across the ore and altered basal horizons; and the subsequent increase in magnesium across these same two horizons. Figure 8 shows the copper and zinc distribution across the same section and highlights the anomalous nature of the Sericite Schist, Granite Gneiss Group, and the iron formations. These are all features typical of many volcanogenic stratiform deposits, and they demonstrate Geco's similarities to this type of deposit.

CONCLUSIONS

The Geco deposit consists of several ore zones. Although these are highly variable in volume and in geometry, they are considered to be separate, conformable lithologies arranged in a stacked order within a sericite schist alteration halo, from whence they grade laterally into iron formations. The concept currently favoured is that separate ore zones represent individual localized cycles or facies of ore deposition during periods of widespread hydrothermal activity. This ore stratigraphy has been developed on an extensively altered basal formation at the upper portion of the Granite Gneiss Group, and is characterized by an abundance of anthophyllite - cordierite units, variations of which can also be found throughout the Sericite Schist Group.

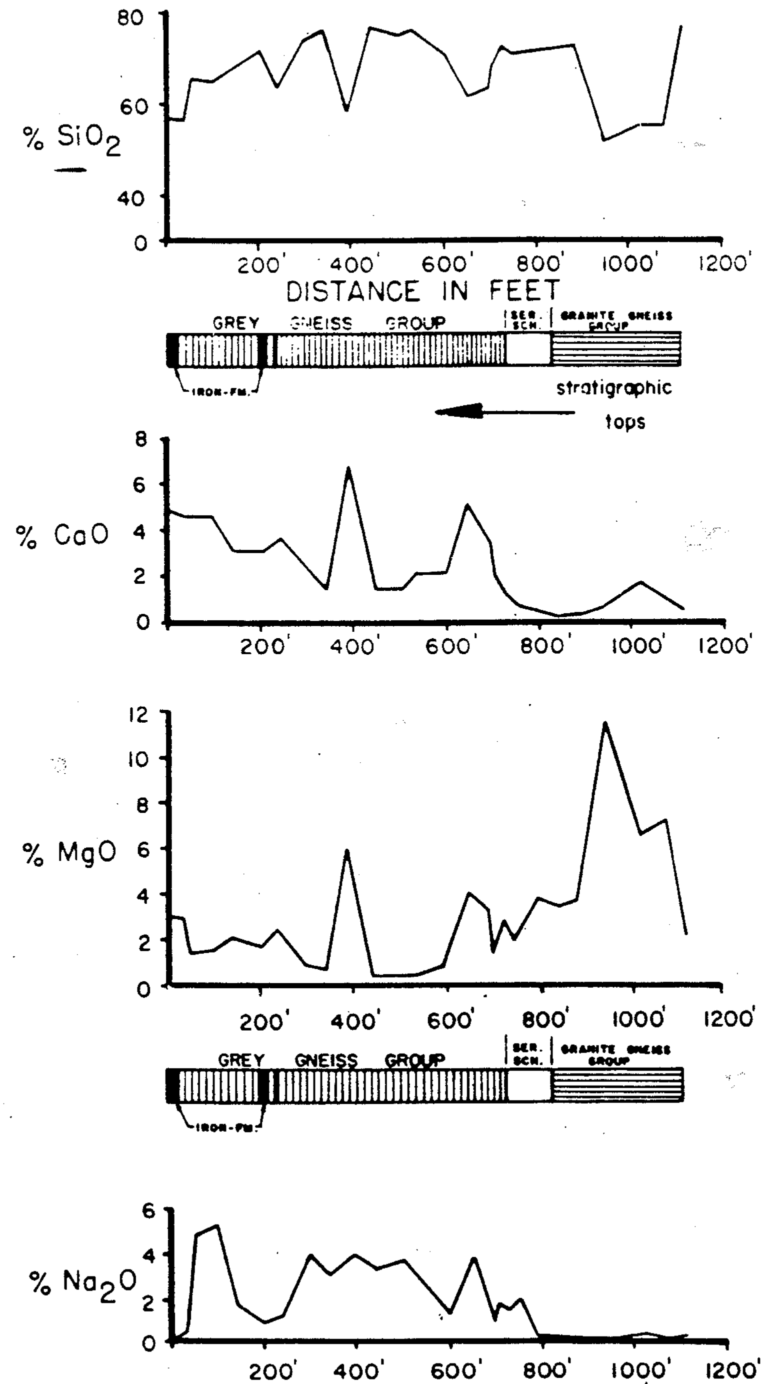


Figure 7 Major Element Distribution Across Geco Stratigraphy.

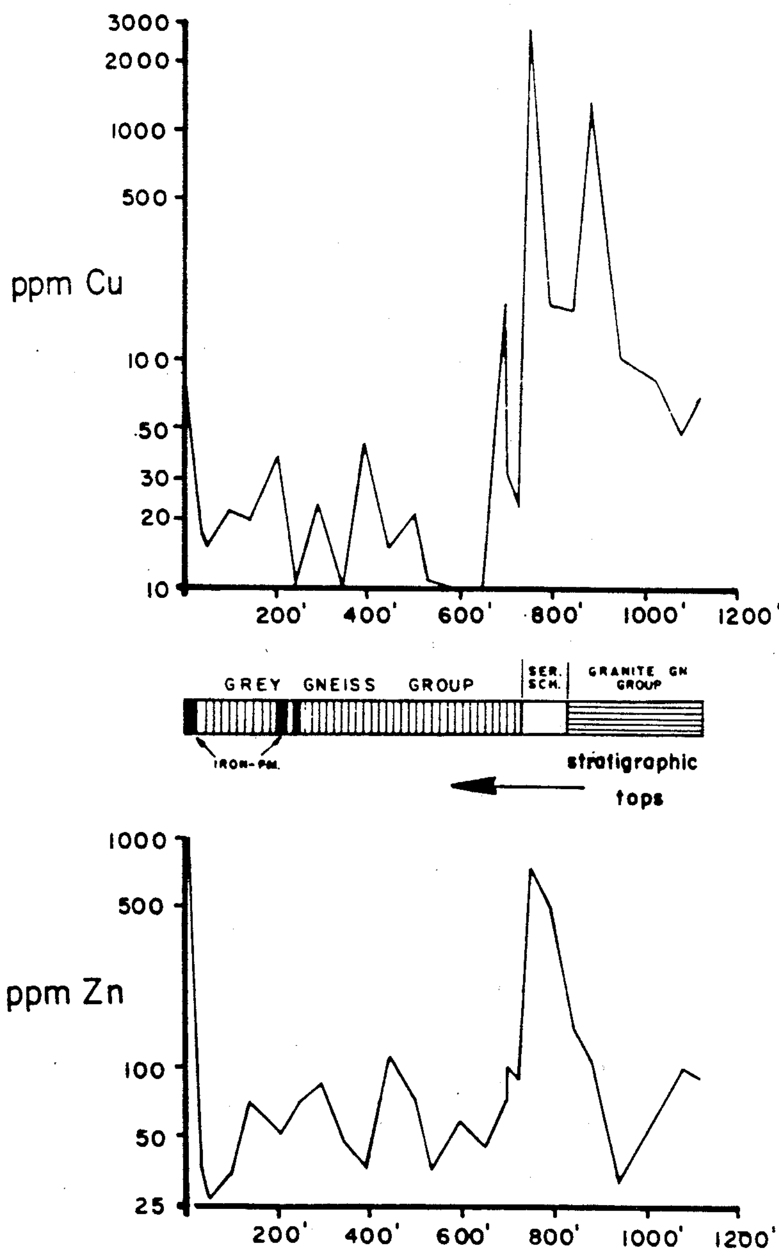


Figure 8 Trace Element Distribution Across Geco Stratigraphy.

NOTES ON MINE TOUR

Because of the advanced state of mining at Geco, there are relatively few exposures of massive ore remaining in the mine. Therefore, specific locations for the tour will be decided upon in August, 1990 in order to give participants the best possible exposure to the ore and wallrock stratigraphy depending on mine development.

Figure 9 is a typical plan (2450 level at Section 54E in this case) showing the altered footwall Granite Gneiss group rocks, the 4/2 copper zone, the massive sulphides and surrounding disseminated ore and barren sericite schist, the 8/2 zinc zone, and the overlying Grey Gneiss Group, which contains numerous sulphide - bearing silicate and oxide facies iron formations.

This represents a complete section across the Mine Series. It may not be possible to view this section in any one place, therefore, some considerable travel around the mine may be required for participants to view the entire section. This will be the aim of the tour.

Due to the relative lack of good exposures underground, participants on the tour will be required to walk several miles and to do some climbing of ladders. Coveralls, boots, and safety gear will be provided at the mine. The tour group should be at the mine by 7:50 in the morning to go underground by 8:20. The tour will take approximately 3 hours. A number of displays and samples in the Geology office will compliment the underground tour, and the Geco Geology staff will be on hand for questions and comments.

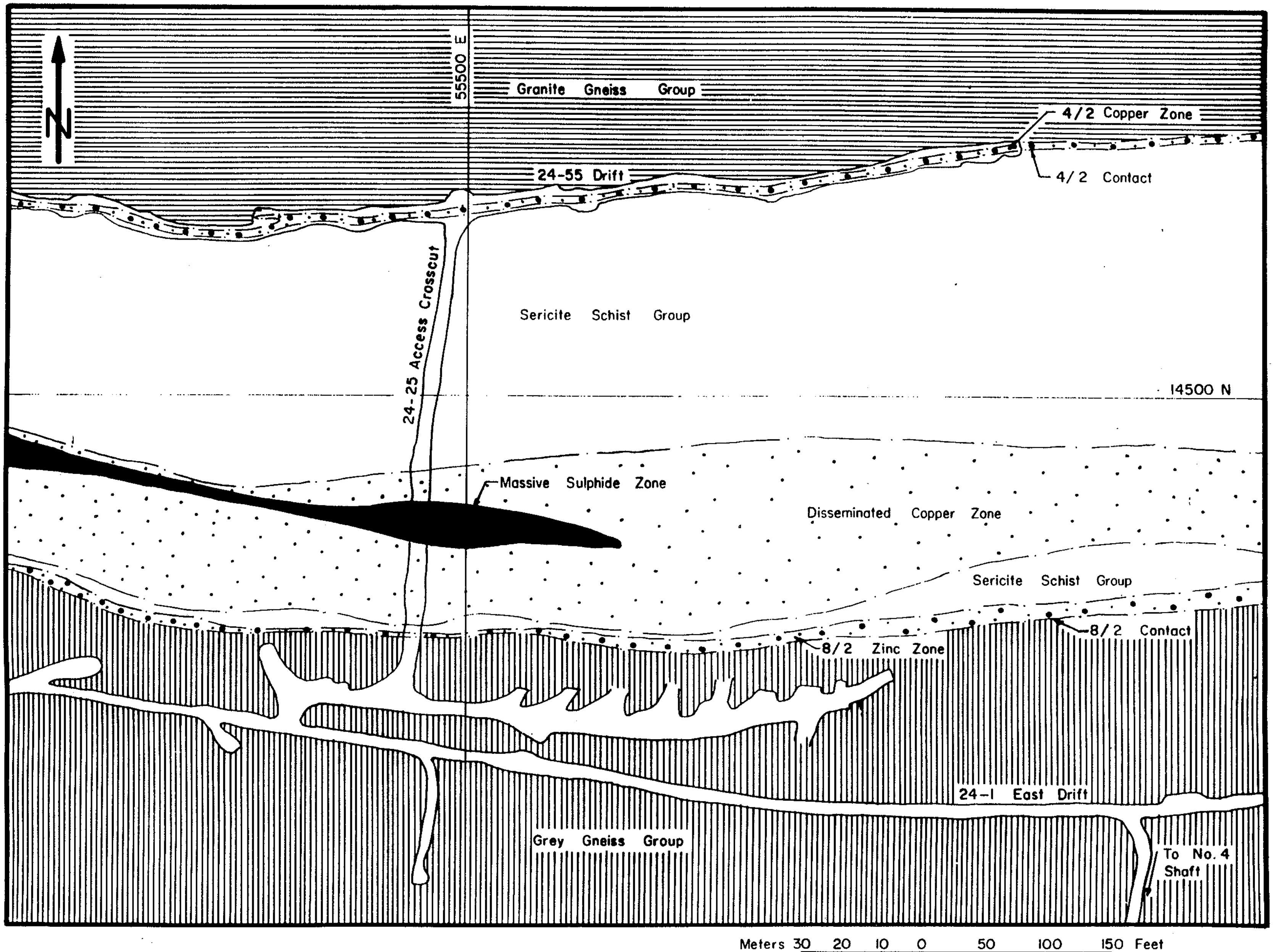


Figure 9 Simplified Geological Plan of 2450 level.

FIELD GUIDE TO THE HEMLO AREA

M.C. Smyk[1], B.R. Schnieders[1] and T.L. Muir[2]

INTRODUCTION

The Hemlo gold deposit is located approximately 35 km east of the Town of Marathon adjacent to and beneath the Trans-Canada Highway (Highway 17). This deposit is currently being mined by three companies: Hemlo Gold Mines Inc. (Golden Giant Mine), Teck-Corona Operating Corporation (David Bell Mine), and Williams Operating Corporation (Williams Mine). These three mines collectively produced 1.18 million ounces of gold in 1989 (Fig. 1 and 2), which accounted for over one-half of Ontario's gold production and approximately one-quarter of Canada's gold production. Ore reserve and grade estimates are presented below.

ORE (Fig. 3)	**RESERVES** (tonnes)	**GRADE** (g/t Au)
David Bell Mine[1]	7 323 000	14.06
Quarter Claim[2]	2 056 800	11.38
Golden Giant Mine[3]	18 026 487	10.85
Williams Mine[4]	29 455 000	6.53

1 P. Bankes, Chief Geologist, David Bell Mine, personal communication, 1989
2 Canadian Mines Handbook, 1989-1990
3 B. Kusins, Chief Geologist, Golden Giant Mine, personal communication, 1989
4 undiluted reserves; A. Guthrie, Chief Geologist, Williams Mine, personal communication, 1990

This section of the field trip guide describes the Hemlo deposit within the contexts of local and regional geology. A brief account of the exploration and development history will be presented. The references contain additional information regarding the various aspects of the deposit. Other field guides have been prepared for the Hemlo area by Patterson (1964) and Harris (1986, 1987).

EXPLORATION HISTORY

Exploration in the Hemlo area dates back to 1869 when gold was discovered by Moses Pee-Kong-Gay near the present town of Heron Bay. Shafts were sunk on auriferous quartz veins and a small amount of ore was shipped.

In the 1920's, J. LeCours, a station agent with the Canadian Pacific Railway at Hemlo station sank test pits on a mineralized shear zone 6 km southwest of the present deposit. J.E. Thomson mapped the area in 1930 and 1931 for the Ontario Department of Mines and recommended several areas, including the area northeast of the Hemlo station and another around Manitouwadge Lake (Thomson, 1932).

In the 1940's, Zeb and Simon Moses noticed "shiny minerals" in the rocks on the north side of Moose Lake while checking Zeb's father's (Peter Moses) trapline. They told Peter Moses, who prospected and collected samples from the area in 1944 (Peter Moses,

[1] Resident Geologist's Office, Ministry of Northern Development and Mines, Box 5000, 435 S. James St., Thunder Bay, Ontario. P7C 5G6

[2] Precambrian Geology Section, Ontario Geological Survey, Ministry of Northern Development and Mines, 77 Grenville Street, Toronto, Ontario. M7A 1W4

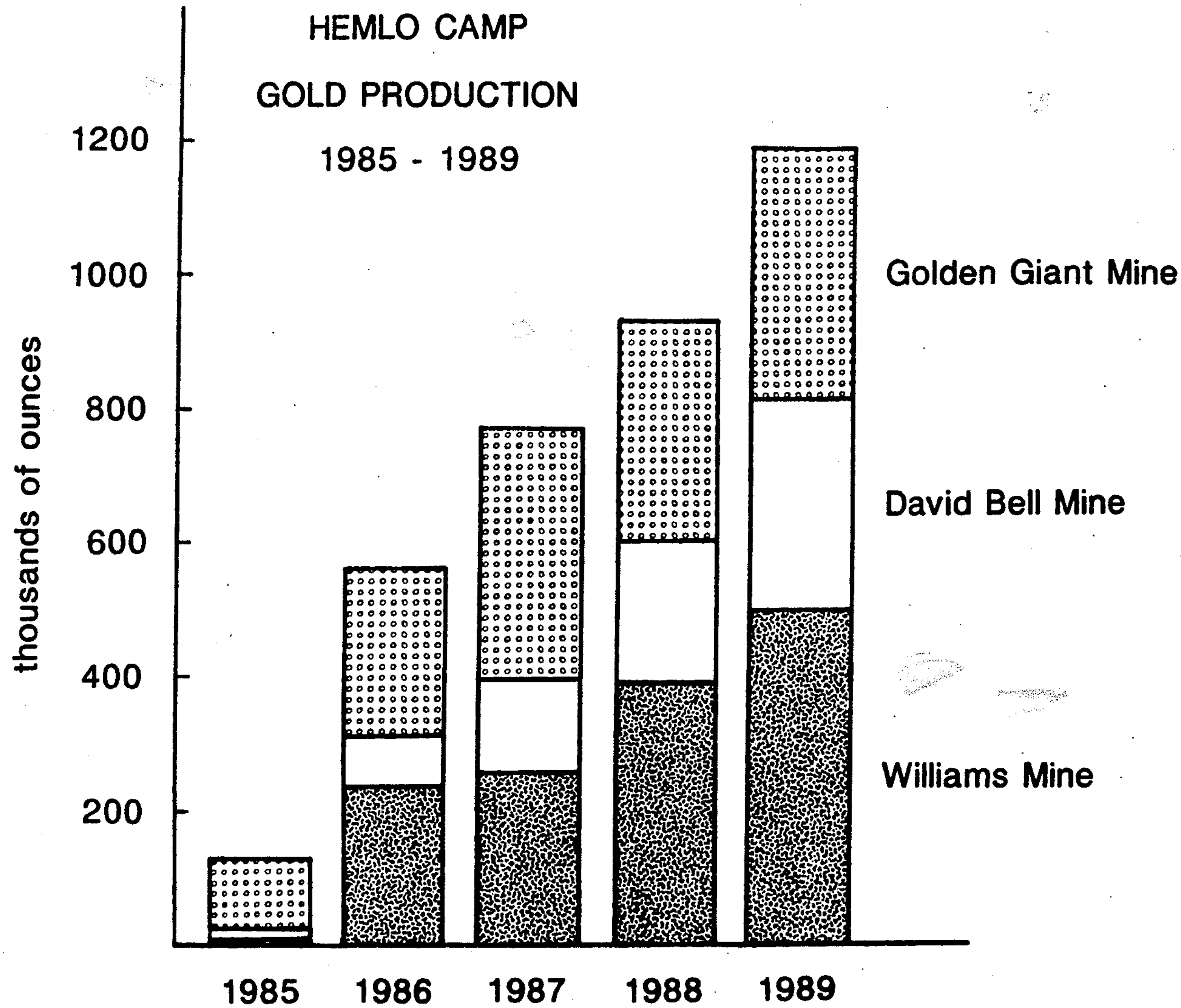


Figure 1 Hemlo camp gold production from 1985-1989.

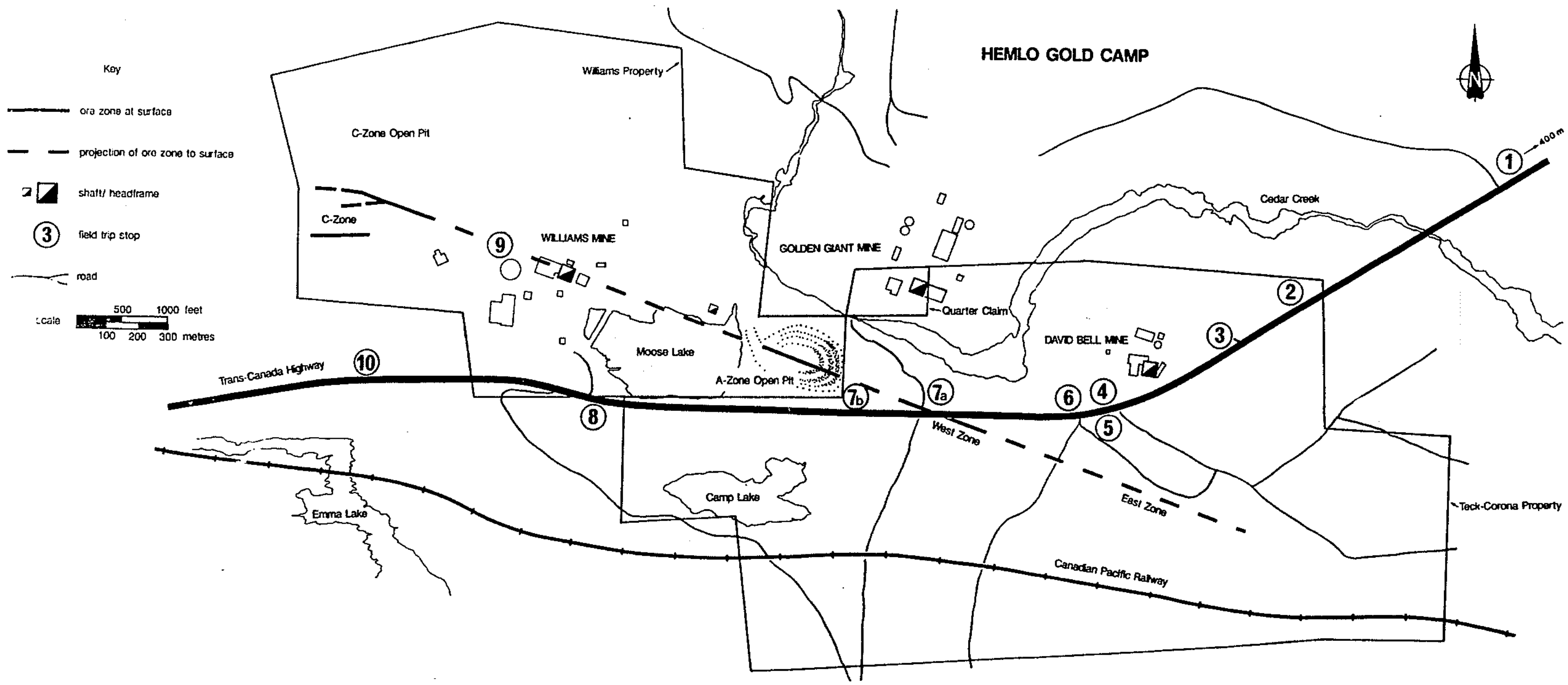


Figure 2 Mining property and ore zone locations in the Hemlo gold camp.

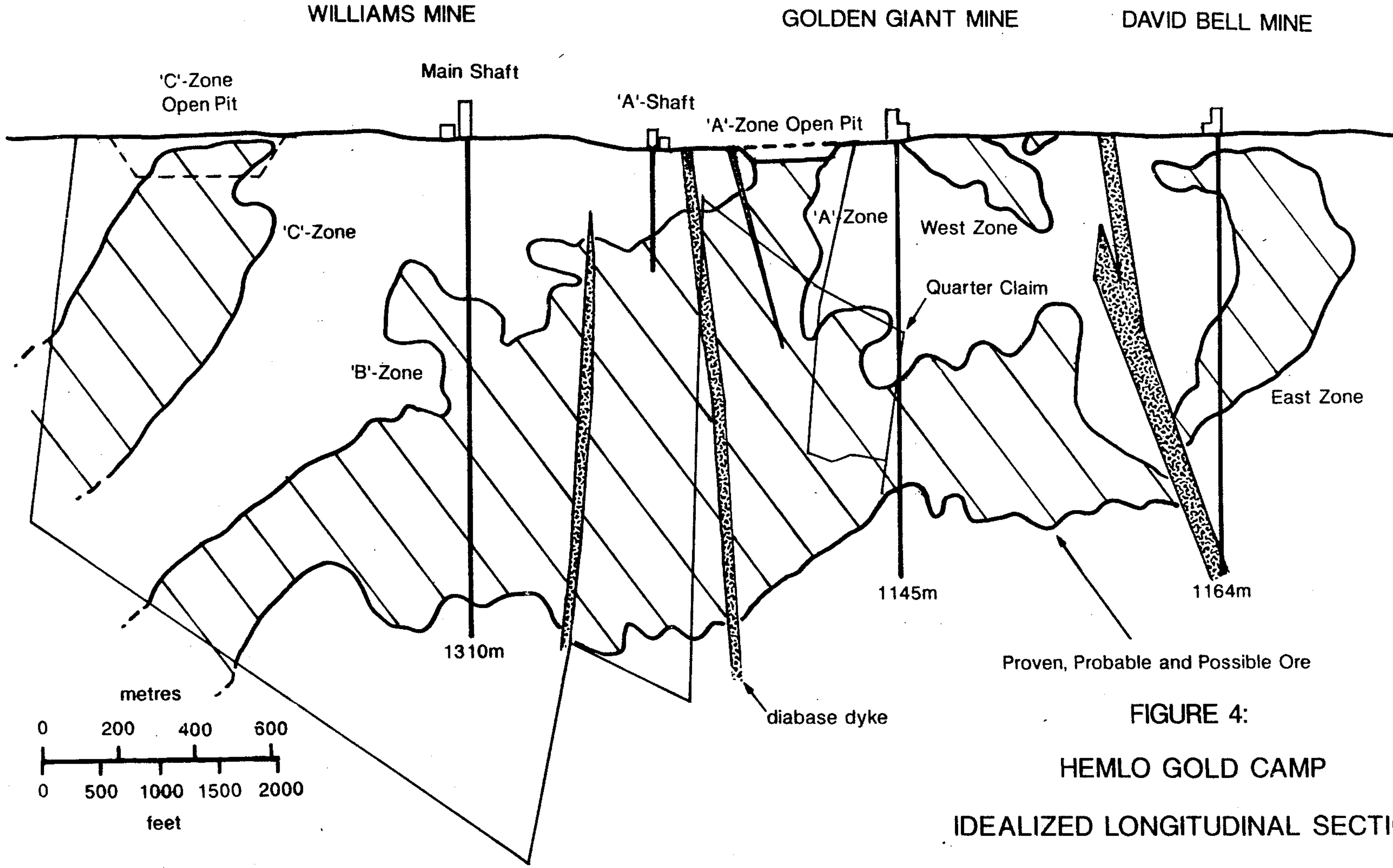


FIGURE 4:

HEMLO GOLD CAMP

IDEALIZED LONGITUDINAL SECTION

(modified after Canadian Mining Journal, February, 1990)

Figure 3 Idealized longitudinal section of the Hemlo gold camp (modified after Canadian Mining Journal, February, 1990).

prospector, Marathon, personal communication, 1990). Peter Moses apparently brought the showing to the attention of Harry Ollmann of Heron Bay, who in turn interested Dr. Jack K. Williams of Maryland. Eleven claims were staked in 1945: five were recorded in Ollmann's name in August, 1945, and six were recorded in William's name in September, 1945. The eleven claims they staked in 1945 became known as the Ollmann-Williams property. Gold values were encountered in a large shear zone and during the next year stripping, trenching and diamond drilling were conducted. The X-Ray diamond drilling program was unsuccessful mainly due to the inefficient operation of the machine and likely the poor core recovery (Page, 1948). Another prospector, Walter Baker, panned gold in Cedar Creek during the 1940's.

Additional claims were staked in 1946 adjacent to the Ollmann-Williams property by a group including consulting geologist, Trevor Page, Moses Fisher (J.E. Thomson's guide), and J.K. Williams. In 1947 the Lake Superior Mining Corporation was formed and carried out mapping, trenching and X-Ray diamond drilling on both properties under the direction of Page (Page, 1948). Although a mineralized zone consisting of 76 653 tons grading 0.3 ounce Au per ton was outlined within what was termed the Lake Superior Shear Zone, work was suspended.

In 1951 Teck-Hughes Gold Mining Ltd. optioned the Lake Superior property. Six thousand feet of diamond drilling increased the deposit size to 89 000 tons at a grade of 0.27 ounce Au per ton. The prevailing US$35 per ounce gold price precluded further work. Teck-Hughes undertook further drilling in 1957 and 1959. The Lake Superior Mining Company was dissolved in 1965.

M.W. Bartley and T.W. Page wrote a geological report on the Hemlo area (Bartley and Page, 1957) for the Canadian Pacific Railway and stated: "The section from Hemlo to Struthers has received considerable prospecting attention to date, and its potential depends mainly on further exploration...".

The Lake Superior property had been staked intermittently by prospectors until 1973 when Ardel Explorations Ltd. staked it and drilled three holes. The deposit tonnage was increased to 150 000 tons grading 0.21 ounce Au per ton. Ardel dropped the claims and was later followed by Cypress Resources Ltd. Claims were staked by R.G. Newman in 1976 west of the Williams property and investigated by Copper Lake Explorations Ltd. in 1977. Soil and rock geochemical sampling succeeded in identifying a zone of anomalous gold values (<10 000 ppb) roughly coincident with the contact between metasedimentary and quartz-feldspar-porphyritic rocks (Resident Geologist's Files, Ministry of Northern Development and Mines, Thunder Bay). T.L. Muir of the Ontario Geological Survey mapped the Heron Bay and hemlo areas in 1977 and 1978, respectively.

Recent developments in the Hemlo camp have been summarized by Bitler, (1988) and Lefolii, (1987), among others. The presented history draws upon many published and unpublished sources in the resident geologists files, Ontario Ministry of Northern Development and Mines, Thunder Bay, Ontario.

Beginning in December, 1979, prospectors Donald McKinnon and John Larche staked the claims surrounding the 11 patented Williams claims, previously known as the Ollmann-Williams property. The two men agreed to pool their claims in a partnership that included Toronto financiers Claude Bonhomme and Rocco Schiralli. In September, 1980, Corona Resources Ltd. a Vancouver-based junior exploration company, optioned the claim group. Corona began a $600 000 drilling program in January, 1981, under the supervision of consulting geologist, David Bell. In March, R. Hughes and F. Lang optioned 156 of the partnership claims to the east and west of the Williams and Corona properties. These claims were put into the holding of the their companies, Golden Sceptre Resources Ltd. and Goliath Gold Mines Ltd., who subsequently relinquished controlling interest to Noranda Exploration Company Ltd.

In May, 1981, representatives of LAC Minerals Ltd. visited Corona's drill site and exchanged information pursuant to a possible joint-venture agreement. Diamond drilling had by then delineated a deposit of 340 000 tonnes grading 6 g/t Au. While Corona negotiated with Williams' widow in Maryland for the Williams property, diamond drilling was stepped back to the east of the outlined deposit. Drill hole 76 intersected a 10.5-foot section grading 0.209 ounce Au per ton at a depth of 336.5 feet. This new, separate zone was the main Hemlo orebody. Corona shares, buoyed up by the new discovery, soared from $2 to $34 by year-end. The Hemlo gold rush, ultimately involving 180 companies, ensued.

Both Corona and LAC had been actively

negotiating for the Williams property. In July, 1981, Mrs. Williams accepted LAC's offer. Corona, citing a breach of a fiduciary agreement, sued LAC for ownership of the Williams claims. In August, 1982, LAC announced the discovery of the deposit on their property, which would become known as the Page-Williams Mine. Drilling by Goliath Gold Mines intersected the northward-dipping extension of the ore zone. The Goliath deposit became the joint-ventured Golden Giant Mine. By the end of 1985, all three mines had commenced gold production.

In March, 1986, after several months of testimony, the Supreme Court of Ontario awarded the Page-Williams Mine to International Corona Resources Ltd. LAC appealed the decision to the Ontario Court of Appeal but continued to operate the mine under conditions imposed by the court. The Ontario Court of Appeal upheld the earlier decision in October, 1987. The Supreme Court of Canada later granted LAC the right to appeal the provincial court ruling. In August, 1989, the Supreme Court of Canada awarded the Page-Williams Mine, Canada's largest gold producer, to Corona. The mine name was shortened to the "Williams Mine".

REGIONAL GEOLOGY

The geology of the Hemlo area (Fig. 4) has been mapped by Thomson (1931, 1932), Bartley and Page (1957), Milne (1967, 1968), Muir (1982a,b), Siragusa (1985) and Siragusa and Chivers (1985). This regional mapping shows that the Hemlo deposit occurs in an east-trending sequence of Archean metavolcanic and metasedimentary rocks, sandwiched between major granitoid complexes. This sequence has been termed the Hemlo-Heron Bay greenstone belt of the Abitibi-Wawa subprovince.

Muir (1982a,b) tentatively subdivided the greenstone belt sequence into two groups: (1) the southern Playter Harbour Group comprising mainly tholeiitic mafic flows with subordinate, intercalated, intermediate to felsic tuff and siltstone; and (2) the northern Heron Bay Group comprising mainly dacitic and rhyolitic calc-alkalic pyroclastic rocks, tholeiitic basalt and minor calc-alkalic basalt. The progressive decrease in grain and fragment size, together with the general increase in volcaniclastic and epiclastic sedimentary units towards the east, was interpreted in terms of a volcanic complex, centered in the Heron Bay area, shedding distal material into a basin to the east (Muir, 1982b, 1983).

To the south of the belt lies the Pukaskwa Gneissic Complex which consists mostly of weakly foliated to gneissic phases of tonalite and granodiorite with pegmatite and aplite dykes. A marginal zone, up to 1 km thick, exhibits a weak mylonitic fabric generally oriented parallel to the volcano-sedimentary contact. Within the greenstone belt lie two major granodioritic plutons: the Heron Bay Pluton to the west, and the Cedar Lake Pluton, with the smaller, satellite Cedar Creek Stock, to the east. The the north of the belt lies a granitoid gneiss complex which is separated from the belt by the crescent-shaped granodioritic to quartz monzonitic Gowan Lake Pluton.

The metamorphic grade ranges locally from greenschist to amphibolite facies. The variations reflect the proximity to granitoid batholiths and/or the superposition of multiple episodes of metamorphism and hydrothermal alteration.

In the vicinity of the Hemlo deposit, Muir and Elliott (1987) and Muir et al., (1988) identified at least four generations of structures produced by at least two deformation events: (1) an early phase resulting in small-scale F1 folds and possible low-angle normal or thrust faults; (2) a major, regional, second phase producing small- to large-scale, tight to isoclinal, generally northwest-plunging F2 folds and a penetrative axial planar schistosity and differentiated layering (S2), possibly associated with sinistral shearing and mylonitization; (3) dextral shear which locally resulted in s-c-c' mylonites and small- to medium-scale, generally northeast-plunging F3 folds with axial planar schistosity and crenulation cleavage (S3); and (4) small-scale F4 kink folds and brittle faults.

The interrelationships of structural elements and intrusive bodies in the vicinity of, and within, the ore zone at Hemlo suggest that intrusions such as the Cedar Lake Pluton and associated(?) dykes were likely emplaced after F1, during the late stages of, or after, F2, but before F3. The relationships between structural and metamorphic elements within the deposit remain somewhat controversial because of the complex,polymetamorphic/metasomaticcharacterofthe ore zone.

All of the aforementioned Archean rocks are intruded by much younger Proterozoic intrusive rocks including diabase dykes, and later lamprophyre and

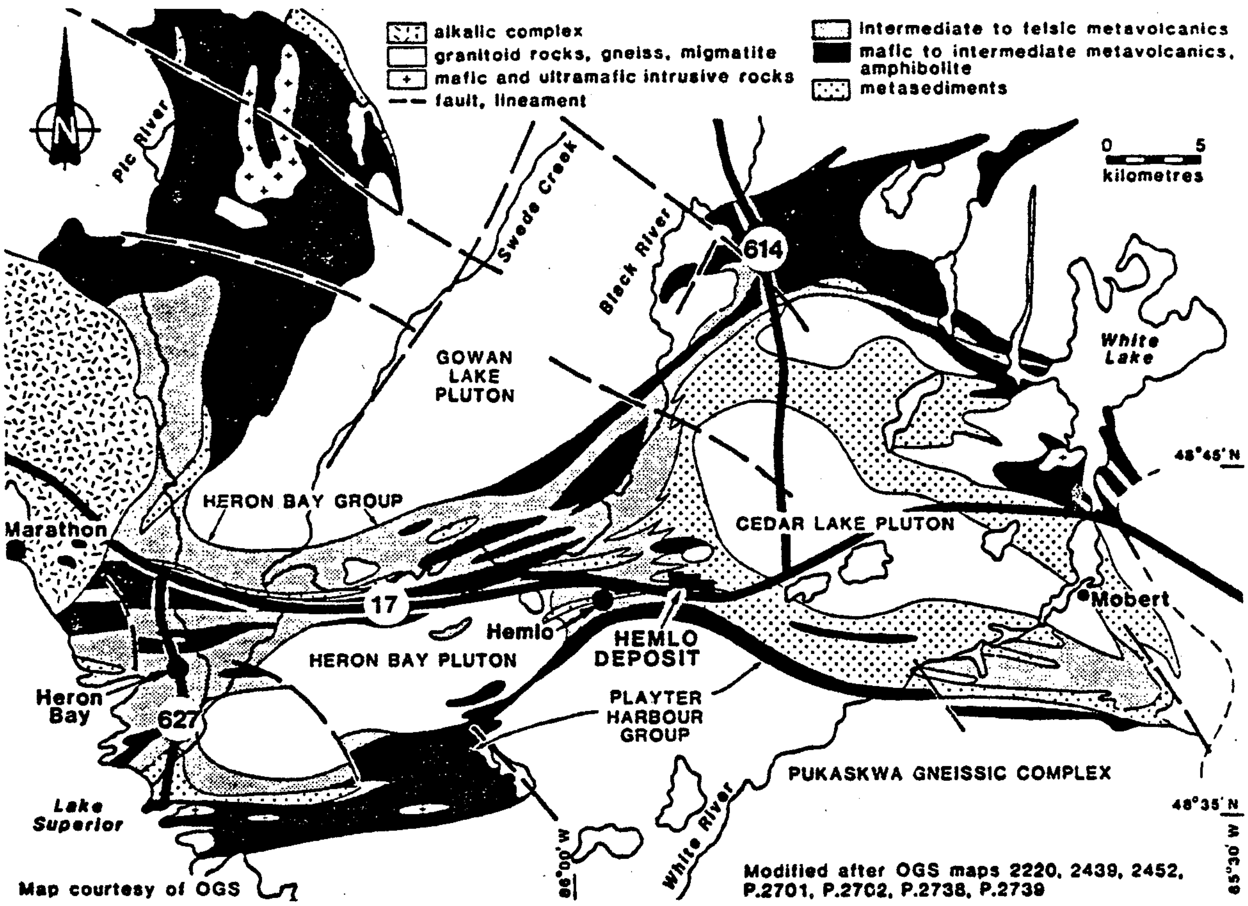


Figure 4 Regional geology of the Hemlo greenstone belt

alkalic dykes that are likely temporally associated with the Coldwell Alkalic Complex, situated west of Marathon. the Coldwell Complex has been dated at 1108+/-1 Ma for early gabbroic and syenitic phases, and ≈1099 Ma for late granitoid phases, by Heaman and Michado (1987).

HEMLO DEPOSIT

The Hemlo deposit is situated on the bifurcated southern segment of the Hemlo greenstone belt, within highly deformed, schistose, felsic metavolcanic and metasedimentary rocks, possibly of the Heron Bay Group. The rocks locally strike at 110° to 115° and dip between 60° and 70° to the northeast.

Page (1948) first recognized the Hemlo Fault as an important structural feature and identified the close co-planar relationship the fault had with the lake Superior Shear Zone. Page (1948) also recognized that all the gold discoveries at that time were hosted by the lake Superior Shear Zone, and stated that the zone had been traced on surface for over seven miles.

Hugon (1984) demonstrated that the Hemlo deposit is contained within a major ductile, dextral, shear zone, situated at the boundary between felsic metavolcanic rocks and metasedimentary rocks.

Underground development has since confirmed the existence of parallel mineralized zones in both the metavolcanic and metasedimentary rocks, as well as mineralized zones which transect the metavolcanic-metasedimentary contact. Hugon (1966) stated that the Hemlo gold deposit occupies the most intensely deformed, central portion of a large-scale, wide zone of ductile, oblique thrusting. Hugon (1986) further suggested that the mineralizing event was syntectonic and occurred after peak metamorphic conditions were established.

The Hemlo deposit is interpreted by the authors as being hosted within a 110° striking flexure of generally east-striking, transposed stratigraphy. Local geology and field trip stops are shown in figures 5a and 5b.

FIELD STOP DESCRIPTIONS

STOP 1: CEDAR LAKE PLUTON

The Cedar Lake Pluton (CLP) and the nearby Cedar Creek Stock (CCS) are weakly foliated, locally massive, hornblende-biotite granodiorite and microcline-megacrystic, hornblende-biotite granodiorite. The margins of the intrusions are characterized by fine-grained, biotite granodiorite with numerous aplite and pegmatite dykes and a marked absence of microcline megacrysts. Zircon U-Pb geochronology has produced dates of 2688 Ma, 2687 Ma, 2687 Ma, and 2684 Ma for the internal phase of the CLP, the marginal phase of the CLP, a plagiclase-phyric dyke located 50 to 100m west of the CLP, and the CCS, respectively (Corfu and Muir, 1989).

Stop 1 is situated on the southwestern margin of the CLP where it is in contact with metasedimentary rocks. Altered and deformed mafic xenoliths are common within the microcline-megacrystic phase. Molybdenite- and pyrite-bearing quartz veins occur locally but contain no gold. Joint surfaces in the metasedimentary rocks and the granodiorite are locally coated with fluorite, epidote, and blue chlorite.

STOP 2: METAWACKE

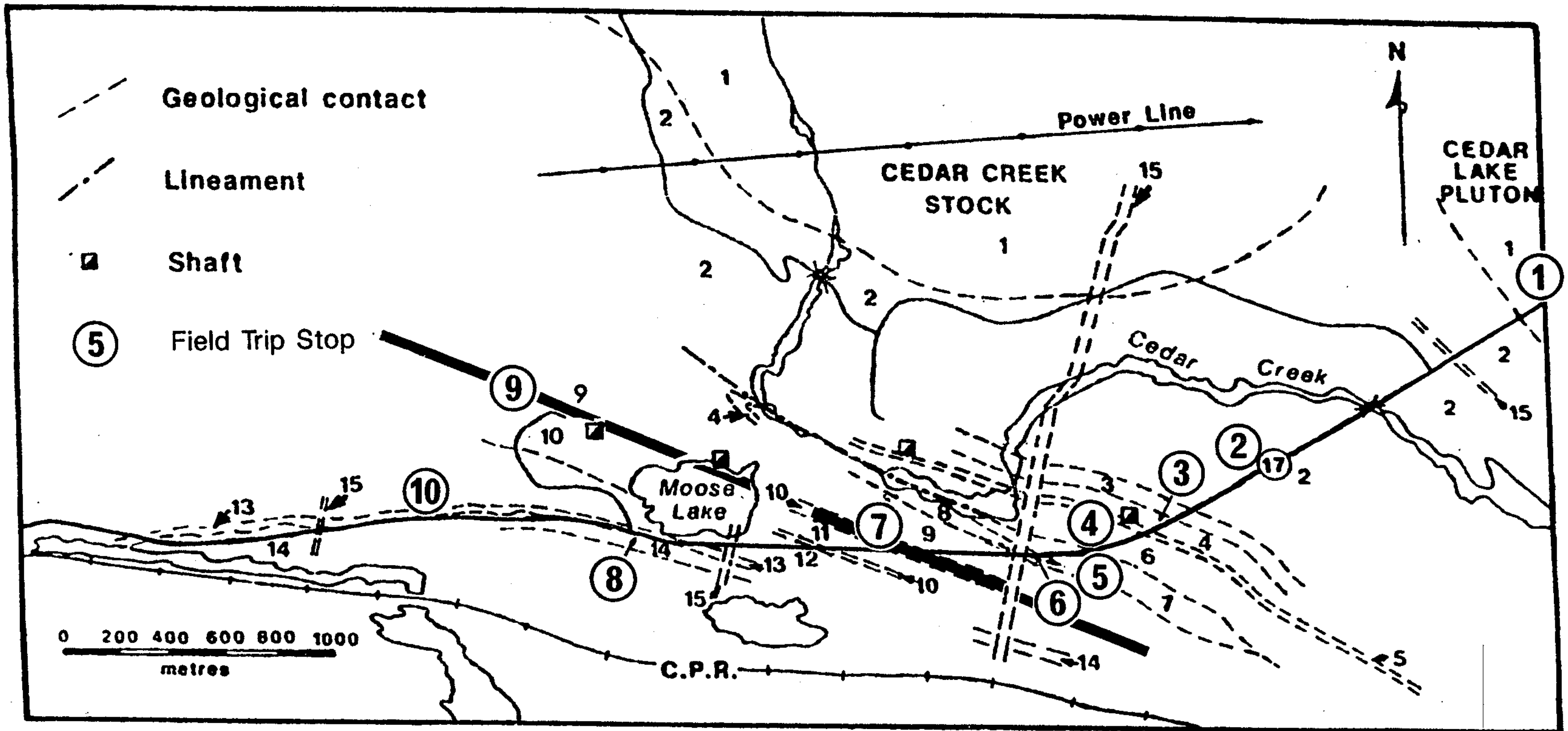
This unit comprises a number of 'packages' of metasedimentary rocks ranging in thickness from 2 to 20 m. The packages are characterized by differences in the thickness of lithologic layering, mineralogical composition, and grain size. These fine- to medium-grained, dark grey rocks consist of various proportions of feldspar, quartz, biotite, amphibole, and locally chlorite, sericite and garnet. Some layers are rich in sand-sized feldspar and biotite, while others, locally referred to as 'calc-silicate layers', have abundant Ca-rich amphibole.

Rocks in the vicinity of Stop 2 may display grading, repeated compositional variations from quartzo-feldspathic to feldspar + biotite + amphibole, and asymmetrical distribution of laminae within composite layers. These features suggest the presence of incomplete Bouma cycles and further suggest that the sediments were perhaps deposited as turbidites. Most primary sedimentary features have been destroyed by deformation. An east-closing F_2 fold is traceable across the highway at this stop. A preferred dimensional orientation developed in biotite defines an S_2 foliation, axial planar to the F_2 fold and locally oblique to $S_1(||S_0)$ compositional layering. The incipient transposition of minerals along S_2, has produced biotite-rich, flame-like structures or a 'pseudo-flame' cleavage.

STOP 3: INTERBEDDED METACONGLOMERATE AND METAWACKE

This unit is approximately 65m thick and can be traced along strike for 2.5 km. It consists of individual matrix-supported, polymictic conglomerate units interbedded with medium- to coarse-grained metawacke. Clasts range from unsorted to moderately sorted and are of granule- to cobble-size. Two lithic types predominate: feldspar porphyry, and lesser biotite- or biotite-chlorite-schist. Less commonly, quartz-phyric clasts and very fine-grained feldspathic lenses (clasts?) are evident. The metawacke tends to be dark grey and coarse-grained with large feldspar grains.

The biotite-rich matrix of both the metaconglomerate and metawacke has commonly developed a strong schistosity. At this stop, the felsic and mafic clasts are flattened and locally Z-folded. This small-scale Z-fold is ascribed to the dextral shear event (F_3)



1 Granodiorite
2 Metawacke
3 Multicolour-layered metamudstones
4 Interbedded metaconglomerate and metawacke
5 Felsic metavolcaniclastic tuff
6 Multilayered metamudstones and derived schists
7 Felsic metavolcanic and metavolcaniclastic deposits
8 Metamudstone (Garnet-Staurolite-bearing)
9 Feldspathic metasediments
10 Quartz-feldspar Porphyritic Complex
11 Layered metawacke and magnetite ironstone
12 Feldspathic metasediments
13 Feldspathic mylonite
14 Gneissic amphibolite
15 Diabase

Figure 5a Lithologic map units, Hemlo area (after Muir and Elliott 1987).

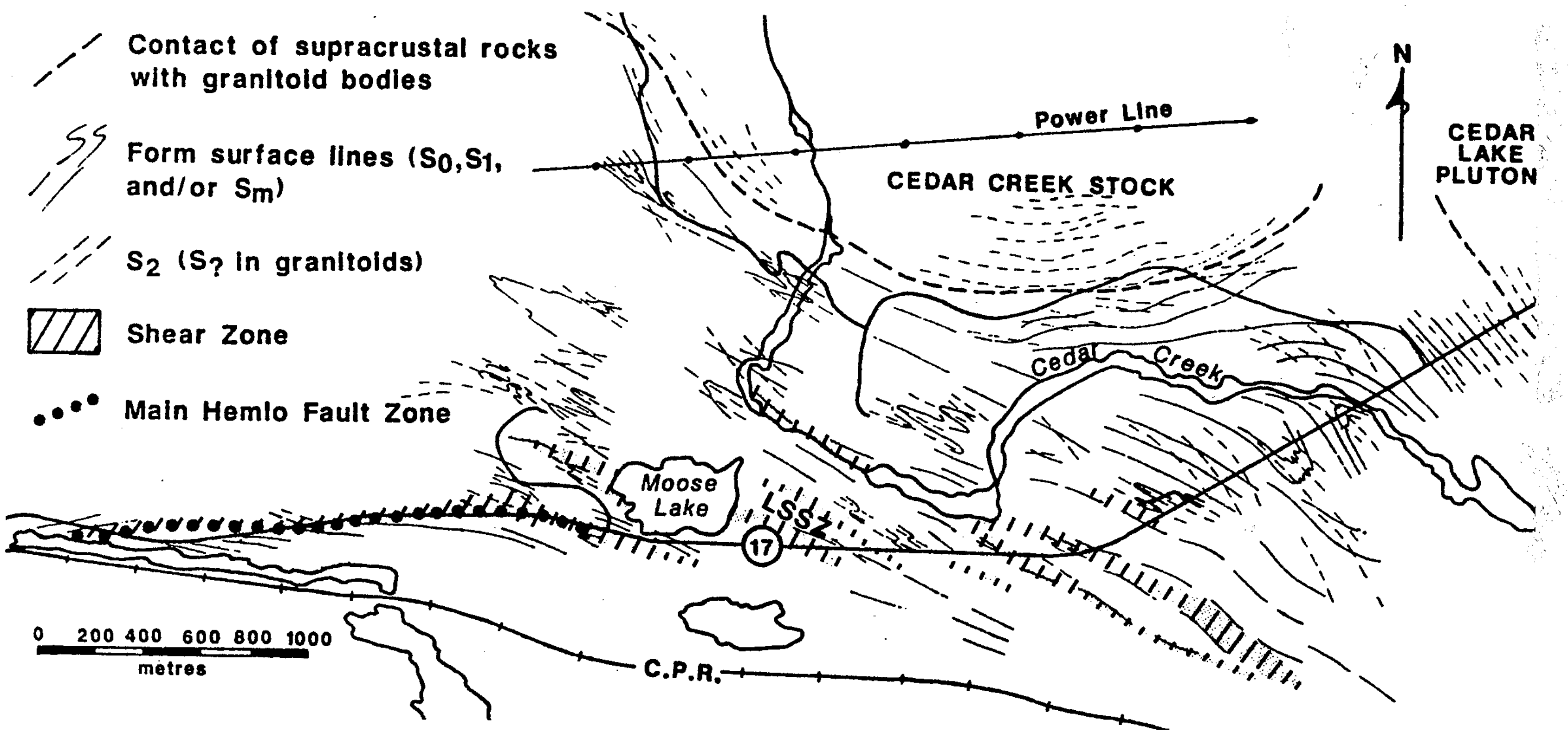


Figure 5b Form-surface map, Hemlo area (after Muir and Elliott 1987).

STOP 4: MULTI-LAYERED METAMUDSTONE AND DERIVED SCHIST

This unit, as exposed along the highway, consists, in its structurally upper part, of layered (2 to 5cm thick) metasedimentary rocks. These layers consist predominantly of feldspar-biotite, biotite-feldspar, feldspar, or amphibole. Quartz is present in varying amounts in the first three types, and garnet is locally present in the biotite-bearing layers. Proceeding structurally lower in the section (to the west-southwest), the rocks become coarser-grained and schistose, and contain additional minerals such as muscovite, and porphyroblastic anthophyllite, sillimanite, staurolite, and possibly cordierite (as completely retrograded lenses up to 3 cm long). The layering is recognizable even where it becomes more diffuse. The layers rich in Ca-amphibole display boudinage and have reaction rinds up to 2 cm thick along their margins. In the core of this section of schists, refolded folds, now isoclinal or intrafolial, are present along with s, c, and c' fabrics, indicating a dextral sense of shear. It is interpreted that local hydrothermal alteration of sedimentary rocks produced compositions in layers which were conducive to the development of the above porphyroblasts. These rocks initially may have been pelitic.

Further west, the grain size decreases, the porphyroblasts consist of garnet and staurolite, the amphibole-rich layers become sparse and the shear fabric is locally less pervasive. It is here that the presence of pyrite produces a rusty weathered outcrop surface. This rusty zone is locally referred to as the Upper Barren Sulphide or "Sucker" Zone, the latter because of its similarity to the main ore zone and the lack of economic ore grade analyses. It commonly contains sericite and lesser amounts of green mica and rare, erratic gold values up to 0.5 ounce Au per ton. Most samples, however, return <0.01 ounce Au per ton.

A narrow Proterozoic diabase dyke intrudes these highly deformed and altered metasedimentary rocks at the western end of the stripped outcrop area.

STOP 5: FELSIC METAVOLCANIC AND METAVOLCANICLASTIC ROCKS

This unit comprises a variety of rock types which have resulted from both primary and tectonic processes. In the northeast part, this unit appears to be a thickly layered (3 to 10 cm), tuffaceous rock with coarse quartz crystals and widely spaced schistosity planes. As we proceed southwest, quartz-feldspar-phyric pyroclastic rocks display crude layering (>0.5 m), best defined by differences in the abundance and distribution of lapilli-sized fragments. An anastomosing, pervasive schistosity may overprint or emulate a fragmental rock texture. Numerous lenses with moderately to highly flattened quartz eyes are present but are of undeterminate origin. Pseudotachylite-filled fractures occur in the southwesternmost outcrops.

STOP 6: GARNET-STAUROLITE-BEARING METAMUDSTONES

This rock consists of 2 to 25 mm thick, light to dark grey layers which are locally attenuated. Amphibole-rich ('calc-silicate') layers display boudinage. Garnet and staurolite porphyroblasts, <1 cm in diameter, display clockwise rotation in the horizontal surface due to dextral shear. Small fibrolitic sillimanite lenses with splaying terminations are present.

Immediately to the west of this outcrop a north-striking diabase dyke separates the Corona West Ore Zone from the East (Main) Ore Zone.

STOP 7: QUARTZ-FELDSPAR-PORPHYRITIC COMPLEX // WEST ORE ZONE

Where exposed near the highway, this composite unit consists of variably altered, strongly deformed, felsic to intermediate, quartz-feldspar-phyric rocks with 1 to 3 m thick layers. Within these layers, minor to abundant lenses having felsic or intermediate compositions display a variety of grain size, texture and mineral abundances. Northwest of Moose Lake this unit consists of quartz-feldspar-phyric varieties of heterolithic pyroclastic rocks, sericite schist, and massive, locally microcline-altered rocks.

Small to extensive patches of K-feldspathization and pyritization are locally developed. Schistosity is penetrative and defined by closely spaced planes of sericite. At the southwest end of the stripped outcrop, a rusty sericite schist containing fine-grained molybdenite, pyrite and green mica represents the surface expression of the West Ore Zone. Historically this outcrop was identified as part the Lake Superior Shear Zone. This zone extends from surface to a depth of approximately 250 m and is exposed on surface along a strike length of approximately 400 m. It contains 340 000 tonnes of 'ore-grade' rock containing an average of 6 g/t Au (Burk et al., 1986). On the south side of the highway along strike from the West Ore Zone, is one of the original Hemlo highway outcrops, a 1 to 2 m^2 exposure of pyritic sericite schist.

STOP 7B: CORONA TRENCH AND 'A-ZONE' OPEN PIT (LAKE SUPERIOR SHEAR ZONE)

The 'A-Zone' open pit can be viewed on the way to the Corona overburden trench. Pit development began in October, 1985 to mine the William's 'A-Zone' orebody. This orebody was approximately 30 m wide on surface and bifurcated to the east and west. The final truckload of ore left the pit in July, 1986 and the ore stockpile was exhausted in September, 1986. 3.8 million tonnes of rock, including approximately 800 000 tonnes of ore at an average grade of 5.13 g/t, were extracted (A. Guthrie, Chief Geologist, Williams Mine, personal communication, 1990). The pit is 275 x 170 m at the top, 150 x 25 m at the bottom and 70 m deep.

The Corona trench was excavated to expose the surface expression of two narrow mineralized zones which extend between the West Ore Zone to the east and the 'A-Zone' portion of the main Hemlo deposit to the west. The two mineralized zones are approximately 15 m apart in the trench and are hosted by pyrite and green mica-rich sericite schist, within less altered quartz-feldspar-phyric rocks. Gold content ranges up to 3 g/t Au. Overgrown trenches from the late 1940's (Lake Superior Mining Corporation) are still visible between the Corona trench and the open pit. The Lake Superior field camp site is located directly south of the highway and is recognized by a diamond drill core dump pile and old building foundations.

STOP 8: GNEISSIC AMPHIBOLITE

The unit at this locality is an extension of a pillowed mafic metavolcanic unit. Although the unit can be traced for several kilometers to the east and west, most of the primary features have likely been obscured or destroyed by deformation. The rocks range from locally schistose, fine-grained amphibolite with quartz-feldspar-epidote veinlets, to quartz-feldspar-epidote-bearing lensoid amphibolite. In the latter variety, gneissosity is largely defined by modal differences in amphibole, feldspar, and locally biotite or epidote.

STOP 9: "HERITAGE OUTCROPS", WILLIAMS MINE

(N.B. hard hats and safety glasses are mandatory on the mine property).

The "Heritage outcrops" are preserved by the Williams Mine because they demonstrate some important aspects of the structure and alteration associated with the deposit. They are situated at the contact between the (footwall) quartz-feldspar-porphyritic complex and (hanging wall) feldspathic metasedimentary rocks. The up-dip projection of the main Hemlo deposit occurs at this contact on the Williams Mine property. The surface projection of the ore zone, approximately 800 m down-dip from here, is manifested as a schistose, rusty zone that has been feldspathized and pyritized. Molybdenite occurs in minor amounts locally. Gold values are erratic and may rarely reach 4 to 6 g/t Au (A. Guthrie, Chief Geologist, Williams Mine, personal communication, 1990).

The metasedimentary and metaporphyritic rocks are intruded by a variety of narrow, mafic, intermediate and felsic dykes. Quartz veins and pods, containing variable amounts of tourmaline, muscovite, chlorite, kyanite and pyrite, are common.

The character of pervasive alteration and deformation is readily apparent. The style of very tight to isoclinal folding that prevails near the mines is exemplified here. A pervasive, spaced, axial planar, pressure solution(?) cleavage (S_2), locally termed "striped cleavage", is associated with these F_2 folds. Units have been disrupted and transposed by shear zones and faults on all scales.

STOP 10: HEMLO FAULT ZONE

The Hemlo Fault was first identified and described in 1948 by T. Page of the Lake Superior Mining Corporation. Page considered the fault to be the most important structural feature because of an assumed relationship to the Lake Superior Shear Zone with which the gold occurrences were associated. The Hemlo Fault was noted as having exceptional complexity of rock types, deformation, and mineralization compared to other local structures. The fault has been traced for several kilometers, but is best exposed west of the mines where the highway right-of-way follows the topographic lineament that defines the fault on surface (Stop 10). The fault zone ranges in thickness up to several meters and is the locus for mafic to felsic dyke emplacement, strong shearing, and tourmalinization. There are similarities and differences between lithologic units on opposite sides of the fault. The suggestion that the fault may be a major structural break is perhaps borne out by equivocal indications of facing reversals across the fault and significant differences in U-Pb ages for metavolcanic rocks at Hemlo (2772+/-2 Ma) and Heron Bay (2695+/-2 Ma) (Corfu and Muir, 1989a), although the Hemlo Fault has not been traced to Heron Bay.

Mafic dykes with picritic basalt composition are sheared and altered to actinolite-, chlorite-, and talc-rich schists. Sinistral movement is indicated by the prevalence, near and within the Hemlo Fault Zone, and S-shaped F_2 folds and possibly of undeformed remnants of the S_2 cleavage oriented clockwise to mylonitic layering. The truncation of layering in the gneissic amphibolite south of the highway (Stop 8) against feldspathic, mylonitic layering, north of the highway is ascribed to early sinistral motion on the hemlo Fault. However, zones of pervasive dextral shear exist in the fault zone, and are characterized by s-c-c' fabrics in schistose layers, F_3 Z-folds and layer-parallel slickensides. Chlorite-actinolite-talc phyllonites at Stop 10 have been the locus of much of this dextral shear. Feldspathic veins with coarse-grained, black tourmaline occur locally.

GEOLOGY OF THE GOLDEN GIANT MINE

Peter Brown, Albert Chong, Bob Kusins and Ken McNena
Hemlo Gold Mines Inc.
Golden Giant Mine

INTRODUCTION

The Hemlo gold deposit is located along Highway 17, 35 km. east of Marathon, Ontario (Fig. 1). The Golden Giant Mine, operated by Hemlo Gold Mines Inc., encompasses the central down-dip portion of the Hemlo deposit, between the Williams Mine (operated by the Williams Operating Corporation) to the west, and the David Bell Mine (operated by the Teck-Corona Operating Corporation) to the east.

The Hemlo gold deposit has been explored by

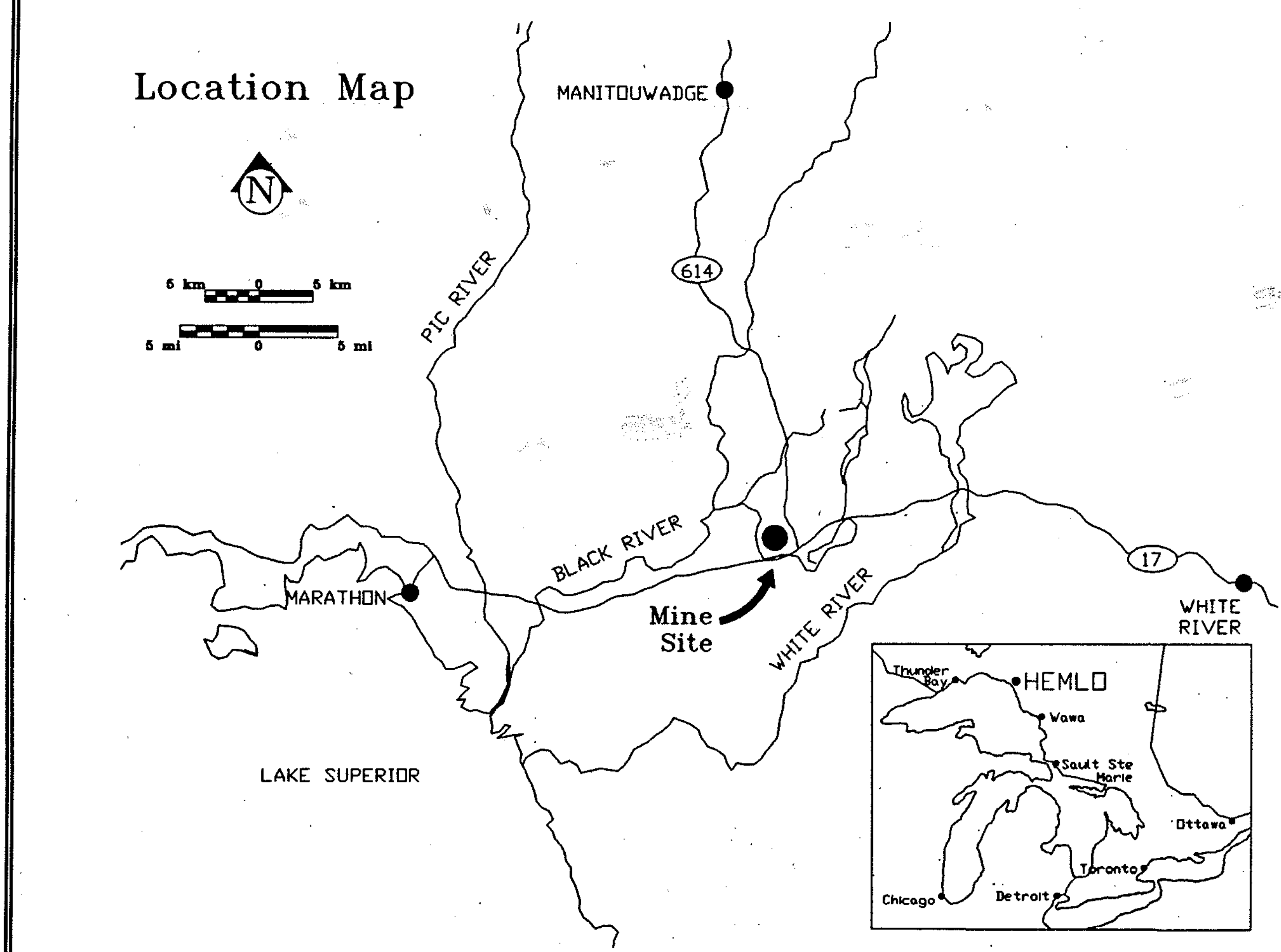


Figure 1 Location map.

various individuals and companies as far back as the 1920's. In 1931, Department of Mines geologist J.E. Thomson recommended the area for exploration. Major discoveries were made by International Corona Resources, Lac Minerals, and Goliath Gold Mines in 1981 and 1982, leading to the eventual delineation of the Hemlo gold deposit.

In November 1982, Noranda Exploration Co. Ltd. optioned the Goliath Gold property and agreed to continue exploration and bring the deposit into production within two years. The first gold bar was produced in April 1985. In 1987, corporate restructuring resulted in the formation of Hemlo Gold Mines Inc. (Noranda Inc. holding 50.6% of shares), with the Golden Giant Mine as its major asset.

Production to December 31, 1989 totals 1,436,163 oz. of gold from 3,892,218 tonnes mined. Current ore reserves stand at 16,900,000 tonnes grading 10.65 grams per tonne. The mine employs approximately 320 people from the surrounding towns of Manitouwadge, Marathon, and White River.

The Golden Giant Mine uses open stope blast hole mining methods with delayed backfilling in the majority of the deposit. This method takes advantage of the deposit's geometry and uniform mineralization. In narrower zones (3-8 m width) a modified Avoca cut and fill mining method is used. Five stoping areas have been established and all are ongoing at various stages of extraction. To date, the mine has operated in the 2800 - 3000 tonne per day (tpd) range, but with recent commissioning of the rail haulage system and main crusher, 3000 - 3200 tpd is easily being attained. Ore is delivered to surface via a six compartment, 1145 m. shaft in two 16 tonne bottom-dump skips. It is then conveyed to the mill where it undergoes crushing to 80 % passing minus 200 mesh. After crushing, the resulting slurry is thickened to 55 % solids before delivery to a leach circuit and in turn a carbon-in-pulp (CIP) circuit. The gold is stripped from solution in the CIP circuit and is then electrowinned and refined, producing a dore bar of average 93 % gold and 3.5 % silver. The mill contains a molybdenum flotation circuit, but to date, impurities are the main problem in recovering this metal. Metallurgical work is ongoing in an attempt to resolve this problem and bring the circuit on-line.

REGIONAL GEOLOGY

The Hemlo gold deposit is located within a sequence of Archean metasedimentary and volcanic rocks which comprise the Hemlo-Heron Bay greenstone belt. The greenstone belt varies from 8 to 20 km. long, and is part of the east trending Schreiber-White River section of the Shebandowan-Wawa Subprovince within the Superior Province of the Canadian Shield (Muir, 1982a, 1982b). Rocks of the greenstone belt are divided into the southerly Playter Harbour sequence, dominated by high iron tholeiitic basalts with interbedded thin tuffaceous and clastic metasedimentary rocks, and the stratigraphically overlying Heron Bay sequence, a series of calc-alkaline intermediate to felsic metavolcanic rocks which grade eastward into a clastic- dominated metasedimentary sequence (Muir, 1982a, 1982b). Late Archean granitoid plutons have intruded the supracrustal metamorphic rocks, and include the Gowan Lake quartz monzonite to the north, the Heron Bay granodiorite to the southwest, and the Cedar Lake granodiorite and its satellite Cedar Creek stock to the north of the Hemlo deposit. The late Archean Pukaskwa granodioritic gneissic complex forms the southern limit of the greenstone belt. The belt is truncated to the west by the Proterozoic Port Coldwell alkalic complex.

A broad, east trending, doubly- plunging synform has been formed in the rocks of the Hemlo-Heron Bay greenstone belt, the axis of which is in part coincident with a major north-east trending fault zone (the Heron Bay-Black River fault) and associated subsidiary splay faults (Brown et al, 1985). The Hemlo, or Lake Superior shear zone is a major regional deformation zone located on the south limb of the Hemlo synform and is spatially associated with the Hemlo deposit. It comprises a complex series of brittle and ductile faults occurring as an east-west trending structure between the Cedar Lake pluton/Cedar Creek stock to the north and the Pukaskwa gneissic complex to the south. It may be related to the Heron Bay-Black River fault.

The Hemlo gold deposit is located near and along contacts between rocks of metasedimentary and felsic metavolcanic - volcaniclastic origins, in the lower part of the Heron Bay sequence along the south limb of the Hemlo synform. The rocks in the mine area strike at an azimuth of 110 to 115 degrees and dip 60 to 70 degrees northeast. The deposit location roughly coincides with the facies transition between metasedimentary rocks to the east and metavolcanic / volcaniclastic rocks to the west (Muir, 1982a, 1982b). Metamorphic grade of the rocks in the Hemlo deposit area is middle amphibolite facies as indicated by the presence of kyanite, garnet, and staurolite in metapelitic rocks. This is higher than the indicated metamorphic grade of the Heron Bay sequence in general (upper greenschist to lower amphibolite facies)(Muir, 1982a, 1982b), and is probably due to a contact metamorphic upgrading effect resulting from the intrusion of numerous granitoid plutons in the vicinity of the deposit.

MINE SITE GEOLOGY

ROCK TYPES

The stratigraphic sequence in the Hemlo deposit area consists of metamorphosed clastic, volcaniclastic, volcanic, and possible epizonal intrusive rocks. In some portions of the stratigraphy these lithologies are traceable for several kilometers. Shearing along lithological contacts as well as alteration and mineralization overprinting make protolith identification for many of these lithologies very difficult.

Detailed descriptions of all the rock types associated with the Golden Giant deposit can be found in Kuhns (1988). Short descriptions for 5 of the principal metamorphic units and intrusive rocks are given below. These are the units that would be encountered in a typical underground visit at the Golden Giant Mine.

UNIT 2 - Originally defined as "footwall metasediments", these rocks were distinguished from the hangingwall metasedimentary rocks on the basis of stratigraphic location. Mineralogical and geochemical studies (Kuhns, 1988) have indicated that the two are basically identical. See Unit 7 for a detailed description.

UNIT 3 - The "Footwall Schist" of the Hemlo deposit consists of white to yellowish-grey, strongly foliated, thinly layered quartz-muscovite-feldspar schist. Average composition is 40% quartz, 33% muscovite, 14% microcline, and 7% plagioclase. Unit 3 also contains tourmaline, pyrite, and lesser biotite, green mica, garnet, barite, anhydrite, fluorite and rarely, scapolite. Varieties include Subunit 3a, which contains lenticular quartz eyes and minor plagioclase eyes, Subunit 3b which contains lenticular plagioclase eyes, and Subunit 3c which is similar to Subunit 3a but contains minor biotite. The Main Ore Zone and Lower Ore Zone generally coincide with the upper and lower margins of Unit 3 respectively.

UNIT 4 - Unit 4 represents a complicated heterogeneous lithology. It is a dark grey to black, medium grained, strongly foliated and compositionally banded quartz-mica-feldspar-amphibole schist. Average composition of the dominant Subunit 4a is 40% quartz, 24% biotite, 8% plagioclase, 7% actinolitic hornblende, 3% tremolitic hornblende, and 7% pyrite. This subunit contains lenticular fragments of feldspar porphyry, microcrystalline grey quartz granofels, and biotite schist. Subunit 4b is a sericite-green mica altered version of Subunit 4a, while Subunit 4c is a non-fragmented quartz-mica-feldspar-amphibole schist which is dark grey to black, fine grained, equigranular, granoblastic textured, and strongly laminated and foliated. Minor calcite and rutile, and trace amounts of garnet, scapolite, barite and rarely, andalusite occur in Unit 4. Unit 4 locally contains economic grades of gold mineralization and is in fact locally mined, but the majority of this unit contains little or no gold mineralization. The distribution of Unit 4, which shows an overall antipathetic relationship with the bulk of gold mineralization, pinches out rapidly with depth and to the east and west of the Hemlo deposit (Walford et al., 1986; Burk et al., 1986). Generally Unit 4 overlies Unit 3, but the two units can be locally interbanded.

UNIT 5 - Unit 5 lithologies represent mineralized rocks and have been designated a separate lithologic group because of extreme difficulty in determining the pre-mineralization protolith. Based on mineralogical associations, the wide variety of ore types within the Main Ore zone can be placed into two main subtypes: 5a-feldspathic and 5b- sericitic. Subunit 5a consists of bluish-grey (due to molybdenite), fine grained, granoblastic mosaic-textured quartz-microcline granofels. It is visually subdivided into high or low pyrite-barite types corresponding chemically to low or high aluminum content, as pyrite-barite content increases at the expense of microcline. High aluminum samples contain an average of 55-60% microcline, 30-35% quartz, 6% pyrite, and

3% muscovite, while low aluminum samples contain on average 45% quartz, 30% barite, 15% microcline, and 10% pyrite. This subunit is commonly fragmented, with molybdenite and native gold-rich granofels fragments supported in a Au-Mo poor granoblastic matrix of barite and/or pyrite. Accessory minerals include albite plagioclase, biotite, garnet, calcite, hornblende, tourmaline, zircon, sphene, magnetite, and anhydrite. Barian microcline is the second most abundant barium-bearing species after barite, containing up to 9.5 wt. % BaO in the ore zone (Harris, 1989). Pyrite occurs as very fine euhedral to anhedral grains in the bluish-grey ore, and as coarser grains in the matrix between ore fragments. There is no direct correlation between gold grade and pyrite content. Banding, defined by increasing mica content, becomes more prevalent as 5a grades into other subunits. Subunit 5b consists of a white to yellowish-grey (muscovite-rich) to pale green (green mica-rich), fine to medium grained granoblastic textured, foliated quartz-muscovite-microcline schist. Average composition is 35-40% muscovite, 35-40% quartz, 15% microcline, and 8% pyrite with minor accessory barite, biotite, and tourmaline, and trace amounts of calcite, zircon, sphene, apatite, and magnetite. Molybdenite and native gold are characteristically foliated and intergrown with the muscovite and green mica. This subunit typically exhibits flaser fabric, and compositional banding defined by alternating 5b, 5a, barite-rich, or pyrite-rich layers.

UNIT 7 - This unit represents clastic metasedimentary rocks, and is typified by a diverse group of foliated and compositionally banded, medium to fine grained rocks. These rocks form the hangingwall and footwall sequences to the mineralized zone/footwall schist package of rocks, both sequences being essentially identical. Unit 7 is divided into 14 subunits, all of which occur interbedded throughout the Hemlo stratigraphy. The two most common subunits intersected by the access drifts of the Golden Giant Mine are 7d and 7e. Subunit 7d consists of the dominant brown to black, laminated and banded, fine grained, granoblastic textured, foliated quartz-biotite-feldspar schist found throughout the Hemlo stratigraphy, interbanded with medium to dark green amphibole-feldspar-biotite granofels. The schist contains on average 47% quartz, 32% biotite, 10% plagioclase, and 8% microcline. The amphiboles in the granofels are dominantly hornblende and actinolite, and they occur with minor diopside. Minor phlogopite occurs with biotite, and the feldspars are potassium-feldspar and minor plagioclase. The granofels has an average composition of 40% hornblende, 20% quartz, 10% microcline, 9% biotite, 8% actinolitic hornblende, 8% plagioclase, and 5% tremolitic hornblende. This subunit is commonly referred to as the calc-silicate unit. Subunit 7e is a pyritic quartz-muscovite-feldspar schist spatially associated with the main ore zone. It is fine grained, granoblastic textured, light brownish-grey to grey, laminated and foliated, and exhibits gradational contacts with other subunits. This subunit is slightly anomalous in gold and contains rare grains of molybdenite, green mica, and arsenopyrite. It's average composition is 46% quartz, 26% muscovite, 13% pyrite, 10% microcline, and 4% plagioclase.

INTRUSIVE ROCKS - Intrusive rocks found in the Hemlo stratigraphy are divided into four units.

Feldspar porphyry quartz monzonite and monzodiorite sills, referred to as Unit 9, occur as thin to thick (0.5-30m) nearly concordant intrusions within and below the Main Ore Zone, and less commonly in the hangingwall matasediments. The sills occur individually or in swarms containing up to 20 sills. Thicker sills (5-30 m) generally exhibit minor to moderate sericitization of the feldspars, and in places are cut by quartz-orthoclase-fluorite veins.

Basaltic to dacitic sills, known as Unit 10, occur throughout the mine area. They are thin (5-50 cm) and seem to cut all mineralization and alteration associated with the deposit. These "mafic sills" as they are commonly referred to, are foliated and exhibit small scale folding and boudinage. Average composition is 35-40% biotite, 35-40% hornblende, 10% plagioclase, 0-10% microcline, and 0-10% quartz with minor zoisite, pyrite, magnetite, calcite, and sphene.

Diabase dykes represent Proterozoic mafic intrusions which cut the strata at high angles. These dykes, referred to as Unit 12, are composed of pyroxene and plagioclase, and typically exhibit gabbroic textures internally, which grade outward to chilled margins. Four main diabase dykes cut the Main Ore Zone, ranging in size from 1-30 m in width.

The fourth main intrusive unit is the Cedar Creek stock, a small oval granitoid intrusion located approximately 800-850 m north of the Main Ore Zone. This intrusion is 2.5 by 1.5 km in surface extent, and its southern margin is subconcordant with the upper metasedimentary schists. The stock consists mainly of medium grained, hypidiomorphic granular rock, which has an average composition of 25-35% quartz, 15-35% plagioclase, 10-25% microcline, 10-20% hornblende, and 1-5% biotite, with minor magnetite, sphene, and apatite.

TOUR STOP DESCRIPTIONS

1. 4750 MAIN DRIFT

The planned first stop on the underground tour is on the 4750 Level, just outside of the 4750 Shaft Station (Fig. 2). At this location, the main access to the Golden Giant orebody moves from the hangingwall to the footwall, and thus is one of the last opportunities to examine the hangingwall metasediments. From the corner where the drift turns south (Fig. 3), a complete stratigraphic section can be examined covering approximately 50 m into the hangingwall, down through a non-economic portion of the mineralized horizon (off the east end of the orebody), and into the footwall schist.

The hangingwall metasediments are composed mostly of Unit 7d "calc-silicates" with interbanded metapelites in the form of subunit 7a, which is garnetiferous and subunit 7b, which is characterized by garnet-staurolite $^{+}/_{-}$ kyanite-sillimanite. Moving south or downward through the stratigraphy, subunit 7e is encountered just above the mineralized horizon. This quartz-muscovite-feldspar schist is almost always found in this stratigraphic location, and is thought to represent the upper portion of an asymmetrical sericitic alteration envelope surrounding the mineralized zone. It is weakly mineralized, most obviously by pyrite, but studies show that it is also enriched in most of the ore-related minor elements (Ba, Ag, Hg, As, Sb). This unit also shows a very strong, pervasive planar foliation, the same as seen in the ore zone, indicating the serictic alteration was developed prior to or at the same time as metamorphism and tectonism.

The mineralized horizon at this location is almost completely composed of quartz-mica-feldspar-amphibole schist, Unit 4. This particular location shows extensive interbanding of subunit 4a with its green mica-sericite altered version, subunit 4b. Note the fragmental nature of subunit 4a and the variety of lithologies present in the fragments. Also visible is a 2.0 - 2.5 m-wide porphyry sill, Unit 9a. Total width of the mineralized horizon at this location is 18-19 m, slightly less than average for the deposit. Chip sample grades (Fig. 4) show anomalous gold values in this area, which is typical of Unit 4. These values are not good enough to meet minimum mining requirements currently set at 2.8 grams/tonne over 3.0 m.

Moving southward again into the footwall schist, subunit 3a is encountered. This quartz-muscovite-feldspar schist is the most common variety of Unit 3, and is characterized by lenticular quartz eyes, lesser plagioclase eyes, and commonly contains tourmaline crystals. This unit is rhyolitic in chemical composition (Kuhns et al, 1986) and is interpreted as being an altered porphyry intrusion. It is interbanded with subunit 3b, which is microcline-rich and contains lenticular plagioclase eyes. This buff to pink coloured rock is marked by potassium enrichment and a decrease in muscovite along with more massive (less foliated) texture. Unit 9 porphyry intrusions and Unit 10 "mafic sills" extensively cut the footwall schist throughout the extent of the Hemlo deposit. Here, subunit 9a feldspar-porphyry-quartz- monzonite sills, and subunit 10a fine grained dacitic sills can be examined.

2. 4700 SUBLEVEL - 0W, 2W CROSSCUTS

The main stop on the underground tour is

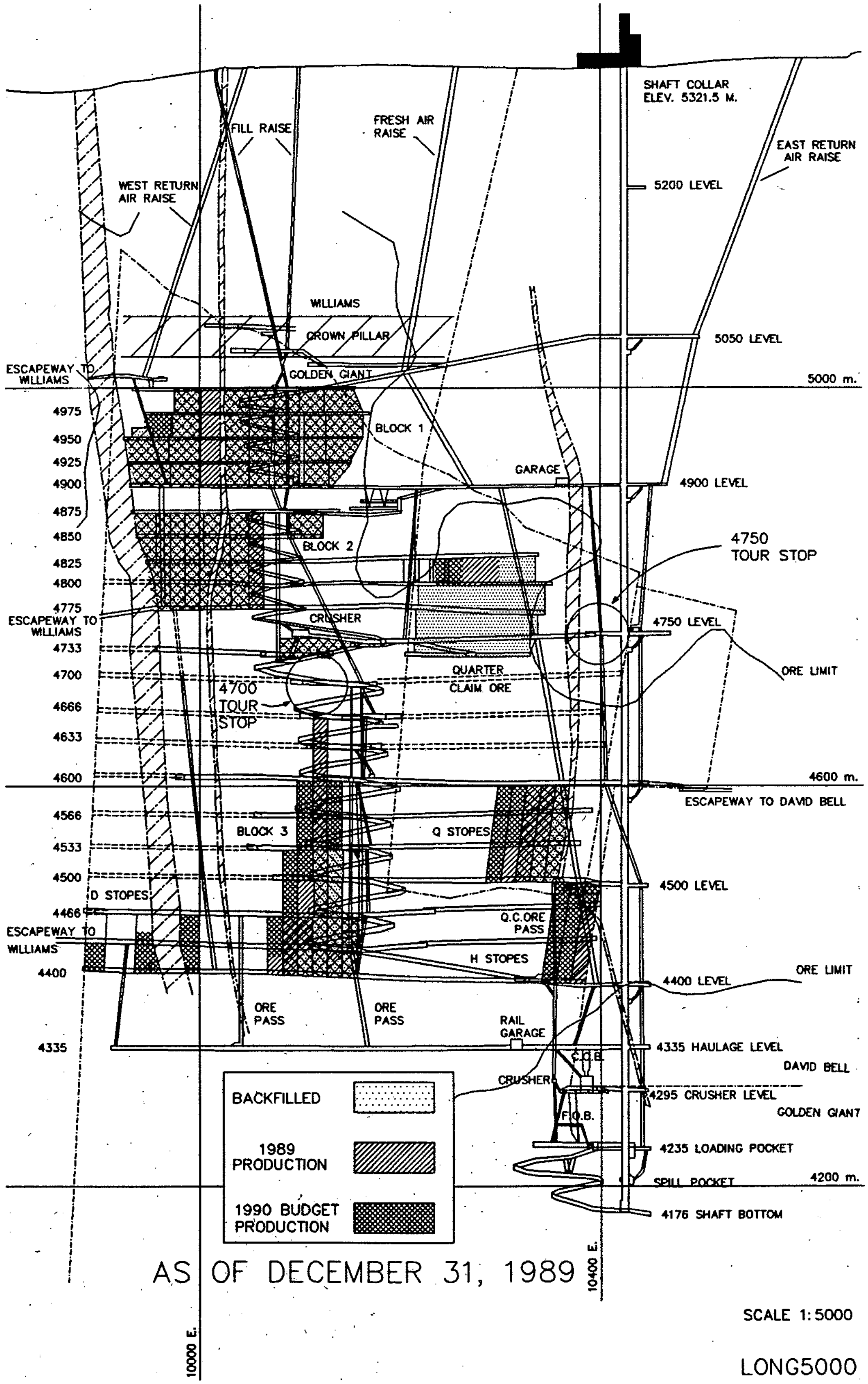


Figure 2 longitudinal section.

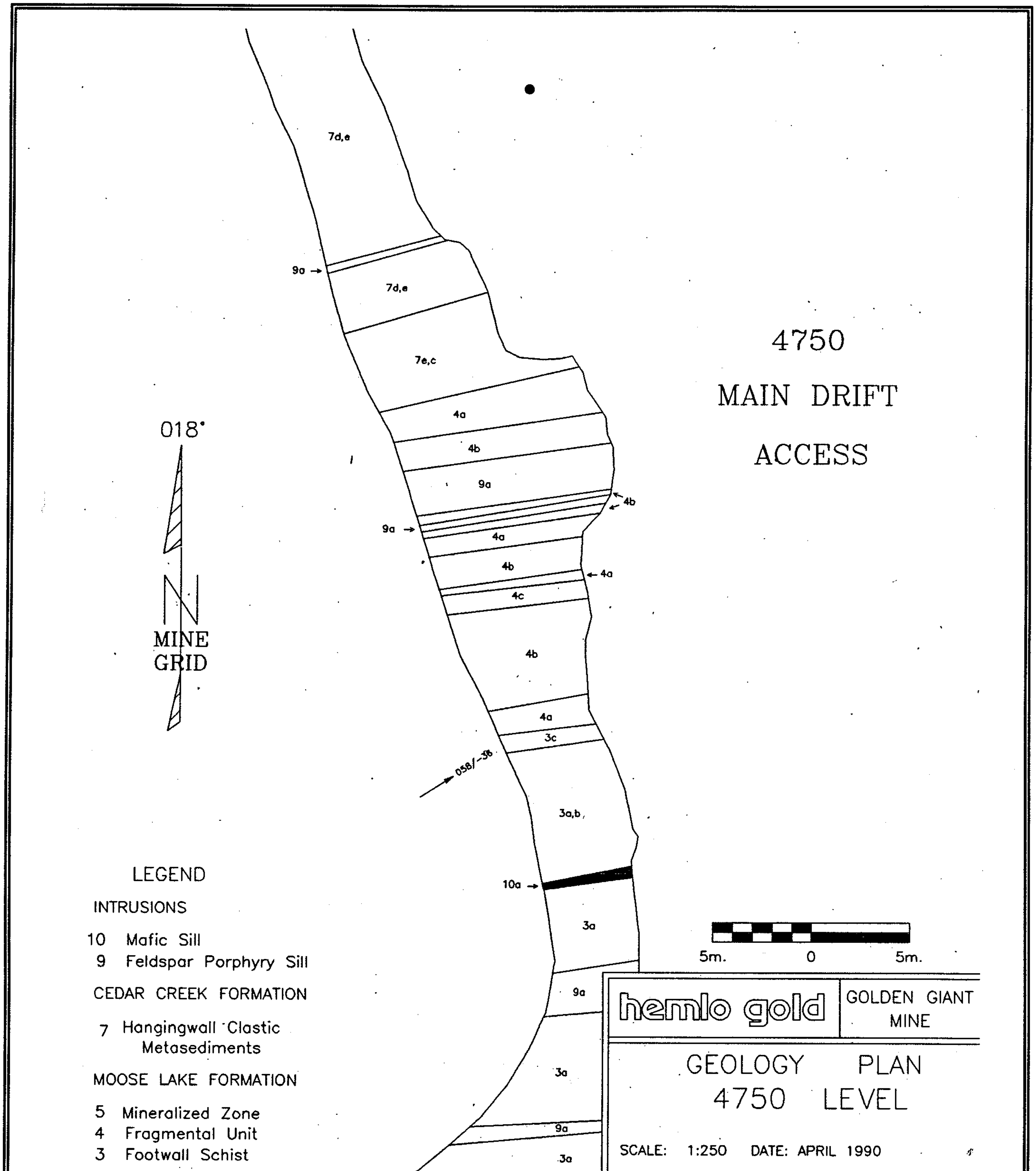


Figure 3 Geology Plan

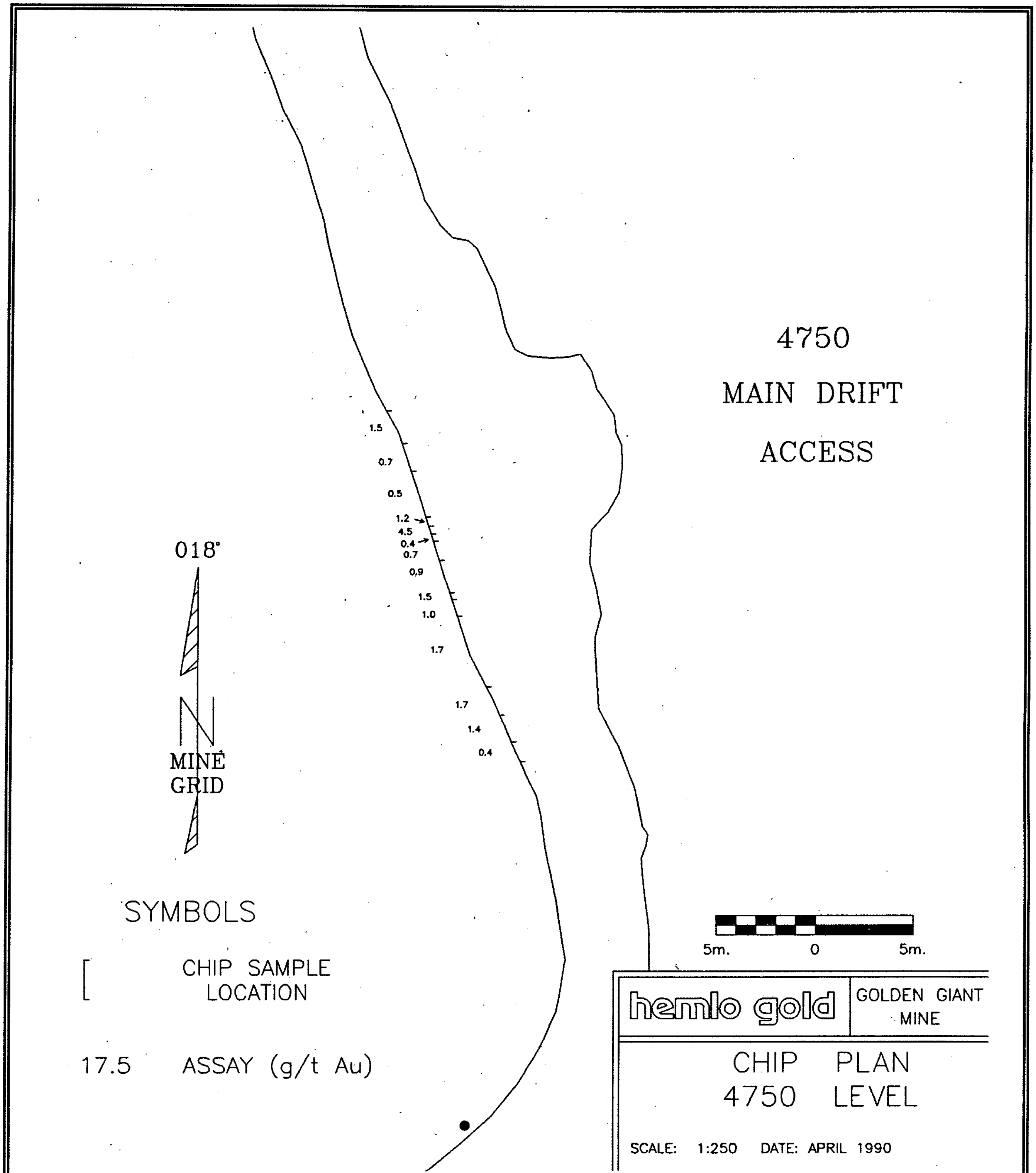


Figure 4 Chip plan

planned to be the 0 West and 2 West Crosscuts on 4700 Sublevel (Fig. 5). This location is approximately half-way down dip on the Hemlo Gold portion of the orebody, just east of the upper portion of the Quarter Claim (Fig. 2). The ore at this location has split into two lenses, hangingwall and footwall, separated by a band of subeconomic Unit 4 quartz-mica-feldspar-amphibole schist. The hangingwall metasediments are poorly exposed, as cutting into them with drift walls causes excessive dilution. The footwall rocks can be examined in the 4700 Footwall Access Drift and the southern portions of the crosscuts.

The ore exposed in the cross-cuts and sill drifts is typical of the Hemlo deposit, with occurrences of feldspathic (5a) ore, sericitic (5b) ore, and baritic (5d) ore. The footwall lens is dominated by baritic ore 5d, which is typical for most of the Hemlo deposit. It is cut by numerous "mafic sills" (subunit 10a) and contains some quartz-rich material (subunit 5d,q). Visible gold within a quartz pod is marked on the west wall of the 2W cross-cut in approximately the middle of the ore zone. The band of Unit 4 quartz-mica-feldspar-amphibole schist separating the two ore lenses contains a very good occurance of "mafic fragmental" subunit 4a. This is best viewed in the north part of the 2W cross-cut. At the contact between Unit 4 and the footwall ore lens, an unusual occurance of footwall schist can be seen. Also at this location is a good exposure of green mica schist. The hangingwall lens of the ore shows a distinctive lack of barite, which is typical, but is rather sericitic (5b) as opposed to being more typically feldspathic (5a).

Overall, the ore zone at this location shows moderate to strong tectonic disruption as indicated by the folding visible in the footwall ore lens, boudinage of mafic sills (subunit 10a), and offsets along crosscutting fractures. The degree of tectonic disruption is higher here as compared to the rest of the deposit, due in part to the lenticular nature of the ore. Zones which are composed solely of Unit 5 generally show less tectonic disruption. Grade distribution within the zone at this location is typical of the deposit. The footwall lens shows lower than average grades (Fig. 6) as compared to the hangingwall lens, due to increased amounts of barite and pyrite which tend to dilute the gold content.

OBSERVATIONS AND DISCUSSION

To date, the following observations have been made concerning the principal features of the Golden Giant deposit:

1. The Main Ore Zone is stratiform, but not stratabound. Non-economic mineralization generally follows the same form.

2. Ore is hosted within a variety of lithological units. Protoliths are very difficult to determine as a result of the high degree of metamorphism and deformation, but are thought to be represented by mud and siltstones, felsic and mafic volcanic and volcaniclastic rocks. There is a lack of any recognizable iron formation.

3. The dominant style of mineralization in the Golden Giant deposit is disseminated gold and molybdenite within rocks containing microcline-quartz, muscovite-quartz, biotite-microcline-quartz, and rarely calc-silicates. A secondary, and minor style of mineralization is as deformed quartz pods and veinlets.

4. Gold grades within the Main Ore Zone are associated with molybdenum content and generally have lower values in the presence of coarse pyrite and/or an increase in barite content.

5. Ore-related elements found in association with the Au and Mo mineralization include Hg, Ag, Ba, As, Sb, V, Zn, and locally minor W, Te, and Tl. Studies indicate strong spatial correlation between Au, Ag, and Mo as the principal disseminated ore type, and Au, Hg, and Sb as a minor, quartz pod (remobilized) ore type (Kuhns, 1988).

6. Ore is hosted within amphibolite facies metamorphic rocks, and mineralization preceded peak metamorphism. This is indicated by the presence of kyanite and sillimanite in the hangingwall metasediments adjacent to the ore zone and gold grains in contact with prograde kyanite (Kuhns, 1988).

7. Alteration consists of an interior potassic (microcline/biotite) zone with localized silicification and pyritization, a surrounding sericitic/phyllitic (muscovite +/- pyrite) zone, and an outer, discontinuous aluminosilicate (kyanite) "halo". Secondary alteration consists of calc-silicate (actinolite/tremolite) zones in

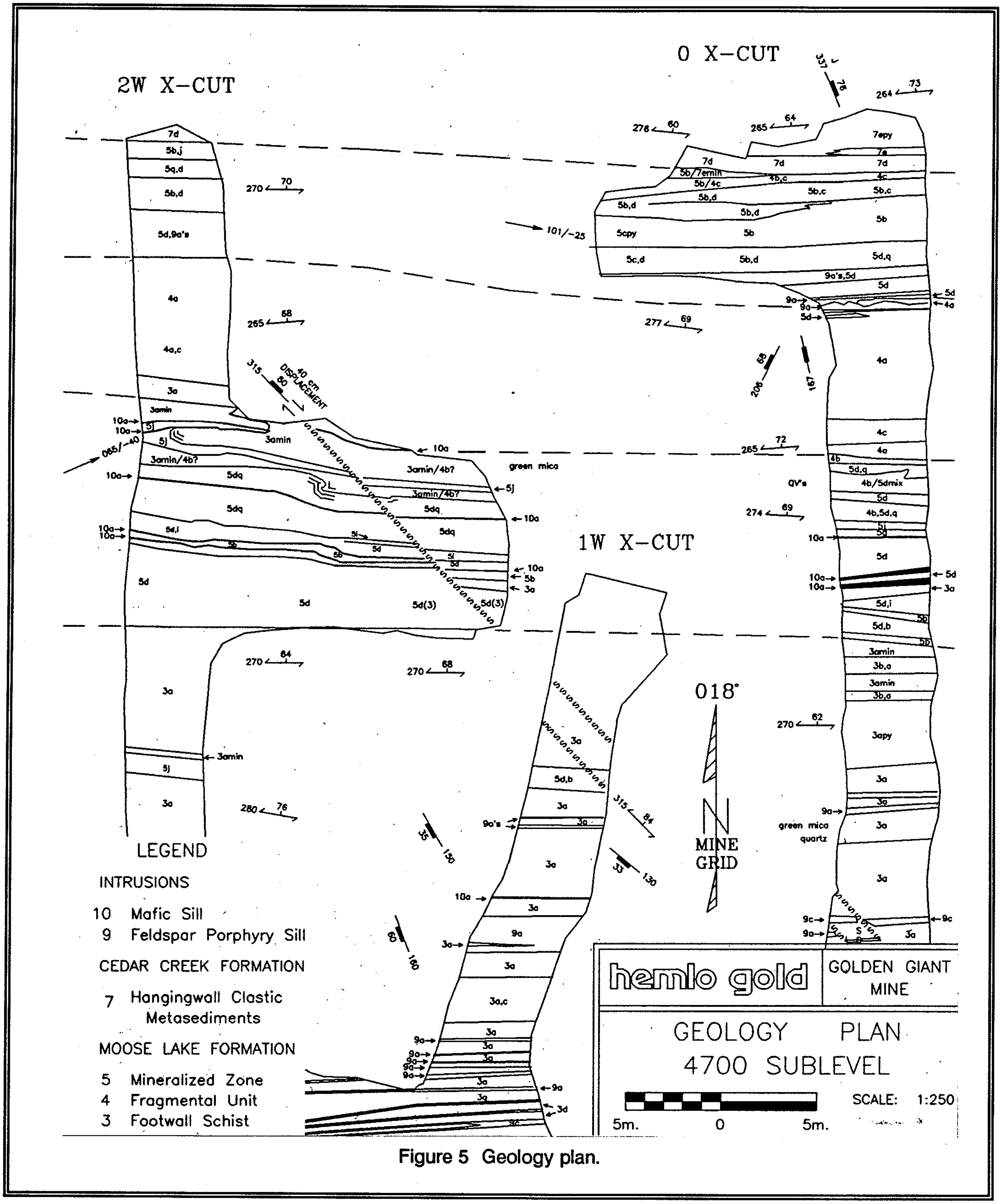


Figure 5 Geology plan.

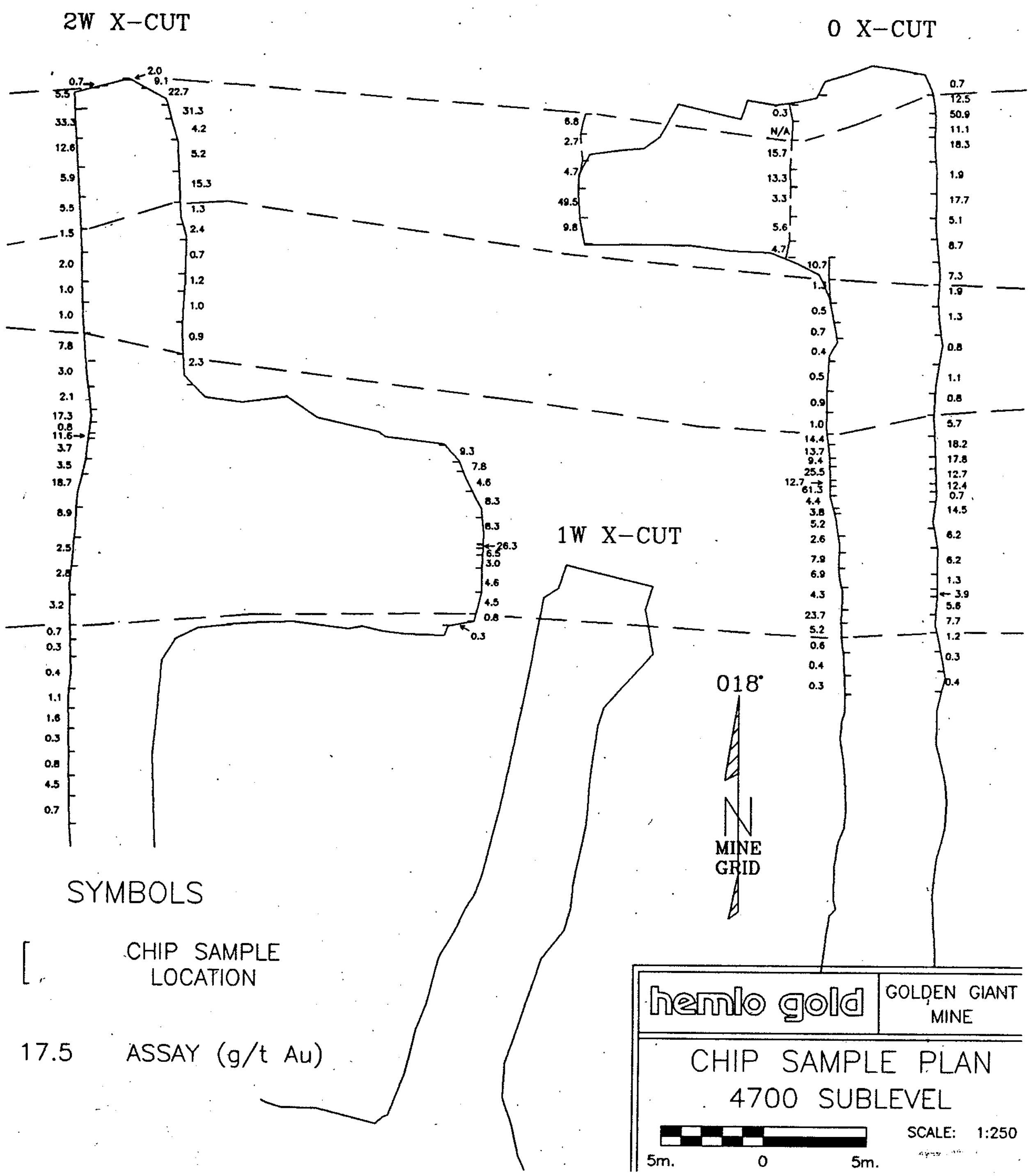


Figure 6 Chip sample plan.

and around the mineralized rocks; a weak, widely distributed, fracture controlled sericitic alteration (bleaching); a secondary aluminosilicate (fibrolite sillimanite) zone coincident with the kyanite zone; and local carbonate alteration associated with parts of the barren hangingwall rocks.

8. Mineralized zones in the Golden Giant deposit are spatially associated with a highly deformed and mineralized quartz eye porphyry (footwall schist), and numerous post-mineralization feldspar porphyry sills and dykes related to the Cedar Creek stock and Cedar Lake pluton.

9. At least three deformational events have been recognized in the Hemlo area (Kuhns, 1988):

   i) Pre-peak metamorphic isoclinal folding and faulting. This first event is recognized by the presence of isoclinal folds through which a penetrative metamorphic fabric has developed.

   ii) Syn-peak metamorphic isoclinal folding and post-peak metamorphic ductile-brittle shearing and associated drag folding. The second folding event is indicated by the presence of isoclinal folding of the metamorphic fabric and refolding of F_1 generation structures. The ore zones and non-mineralized country rocks are strongly foliated and exhibit dextral mylonitic and cataclastic textures attributed to post-peak metamorphic ductile-brittle shearing.

   iii) Late brittle faulting. Brittle deformation is indicated by multiple well developed angular fault breccias and clayey to rock flour- rich gouge zones which are developed subparallel to the regional metamorphic fabric.

10. The presence of a major fault structure (Lake Superior Shear Zone) containing mylonitic and cataclastic textures, and which is also the focus of early porphyry (footwall schist) and later feldspar porphyry dykes, suggests that this structure has been reactivated many times.

SUMMARY

The Hemlo deposit has undergone a complex geological history which makes genetic modelling of the orebody rather difficult. Some of the key points that have been recognized are that the ore zones have had rigid structural controls which are evident by the well defined contacts between ore and host rocks. Also, the mineralization shows an antipathetic relationship with the Unit 4 quartz-mica-feldspar-amphibole schist, which may have provided a suitable site for focussing mineralization. In addition, the porphyry intrusions often follow the same trends as the gold grade, indicating a strong spatial assocation and possible source for the mineralizing event.

Hemlo Gold Mines Inc. is committed to an ongoing detailed examination of the Hemlo deposit and surrounding area. This, in the form of exploration drilling, structural and geochemical studies - both in house and academic, three dimensional computer-aided modelling, and sharing of information with neighbouring operations will help towards a better understanding of this world class gold deposit.

PROTEROZOIC AND ARCHEAN GEOLOGY: HEMLO TO WINSTON LAKE

B.R. Schnieders and M.C. Smyk
Resident Geologist's Office,
Ministry of Northern Development and Mines,
Box 5000, 435 S. James St.,
Thunder Bay, Ontario.
P7C 5G6

The Trans-Canada Highway between Hemlo and the Winston Lake turnoff transects a variety of Archean to Late Proterozoic rocks of the Terrace Bay greenstone belt and the Coldwell Alkalic Complex, respectively (Fig. 1). The following field stops serve to illustrate some of the best developed and preserved characteristics of these diverse rocks.

STOP 1: LAYERED GABBRO OF THE COLDWELL ALKALIC COMPLEX

This stop is located approximately 1.8 km east of the Marathon turnoff on the Trans-Canada Highway. Parking is available on a large extended gravel shoulder/ rest area on the south side of the highway. An excellent exposure of the well-layered Eastern or border gabbro of the Proterozoic Coldwell Alkalic Complex (Fig. 2) extends along the length of the gravel shoulder. This early gabbroic phase of the Complex and related syenites have been dated by Heaman an Michado (1987) at 1108$\pm$1 Ma. The gabbro at this stop is intruded by dykes of medium-grained to pegmatitic ferroaugite syenite, and has been described by Mitchell and Platt (1982):

"The arcuate mass of basic rocks which define the eastern and northern margin of the complex is commonly referred to as the border gabbro to distinguish it from the alkaline gabbro of Center 2. This border gabbro is considered to belong to Center 1 activity as it is intruded in many places by ferroaugite syenite. The petrological relationship between the two magmas is, however, unclear. Ferroaugite syenite is unlikely to be a direct differentiate of the gabbro because of the greater volume of the former and the lack of mineralogical gradations between the two rock types. The zone of gabbro defines a prominent magnetic low... . This is considered by Lilley (1964) to be due to reversed magnetization of the gabbros. The gabbros are composed of (Fo_{67-43}) augite, plagioclase (An_{60-35}) and minor orthopyroxene (En_{55-66}) (Lum, 1973). The orthopyroxene may be a product of assimilation of Archean metasedimentary rocks, a xenocryst derived from the pyroxene hornfels thermal aureole, or a relict high-pressure phase.

The gabbro has been extensively prospected for its copper potential as accumulations of pyrrhotite and chalcopyrite, with minor pentlandite, cubanite, pyrite, bornite, arsenopyrite and mackinawite (Watkinson et al., 1973; Lum 1973) are common.

The excursion stop is close to the contact between the gabbro and the ferroaugite syenite. Many pegmatites of ferroaugite syenite cut the gabbro at this locality. The gabbro contains variable amounts of Archean xenoliths. Here the gabbro shows all transitions from massive homogeneous gabbro to rocks with well-developed igneous layering. The layers are not traceable over long distances and do not serve to outline the structure of the gabbro intrusion."

Copper and minor nickel and platinum group element mineralization is associated with the gabbroic rocks of the Coldwell Complex. On the eastern margin of the Complex, several textural and compositional varieties of gabbro occur in a crescentic mass. The Marathon deposit, approximately 10 km north of this stop, is under current joint ownership of Fleck Resources Ltd. and Euralba Mining Ltd., and has drill-indicated reserves of 47 million tons of possible ore grading 0.42% Cu, 0.02 ounce per ton Pt, 0.054 ounce per ton Pd, and trace Ni, Co, Rh, Au and Ag. It is described by Dahl et al., (1986; 1987 a, b), and Watkinson et al., (1986).

FIGURE 1:

General geology, Schreiber to Coldwell

② Field Trip Stop

FAULTS
KEWEENAWAN
KEWEENAWAN DIABASE DIKES
KEWEENAWAN ALKALINE COMPLEX
ARCHEAN
GRANITOIDS
METASEDIMENTARY and METAVOLCANIC ROCKS

N

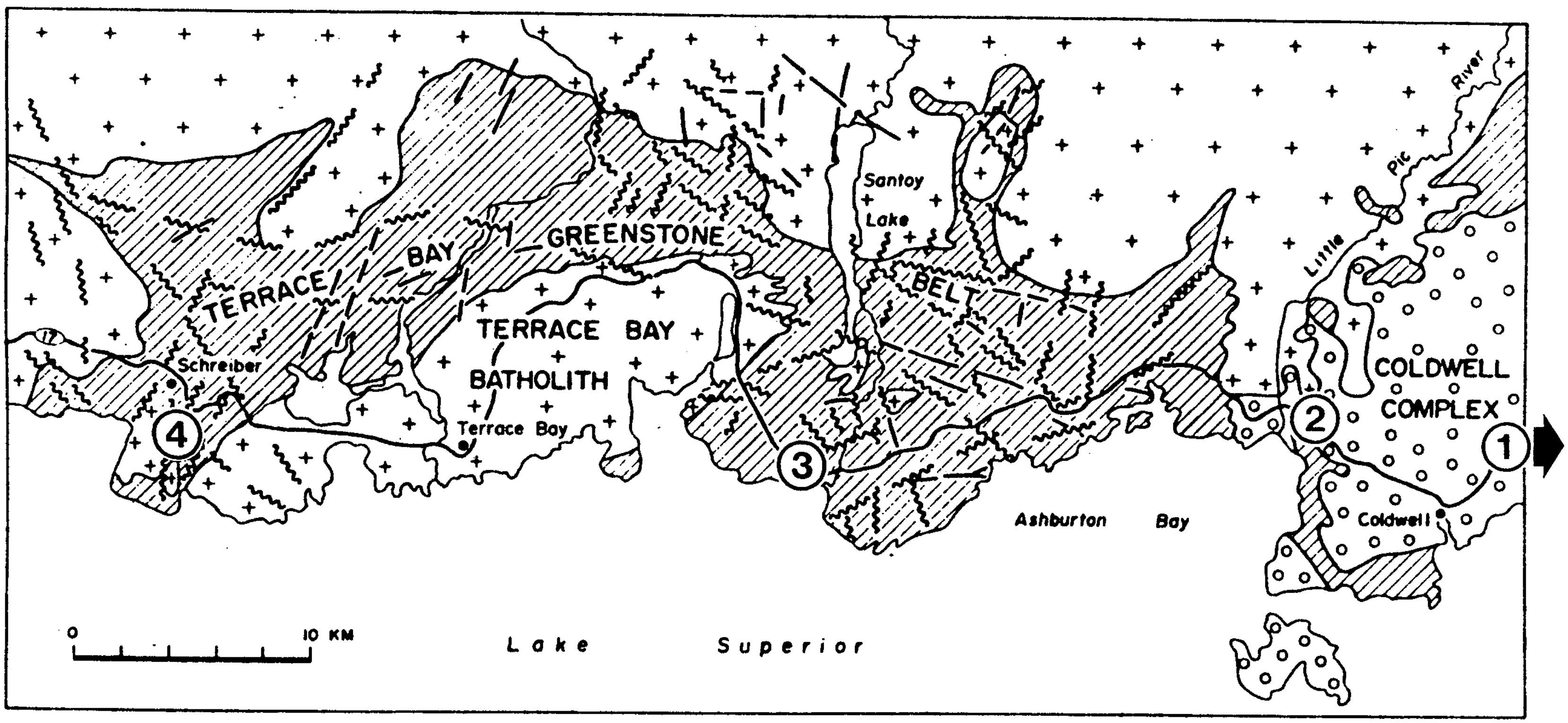


Figure 1 General geology, Schreiber to Coldwell.

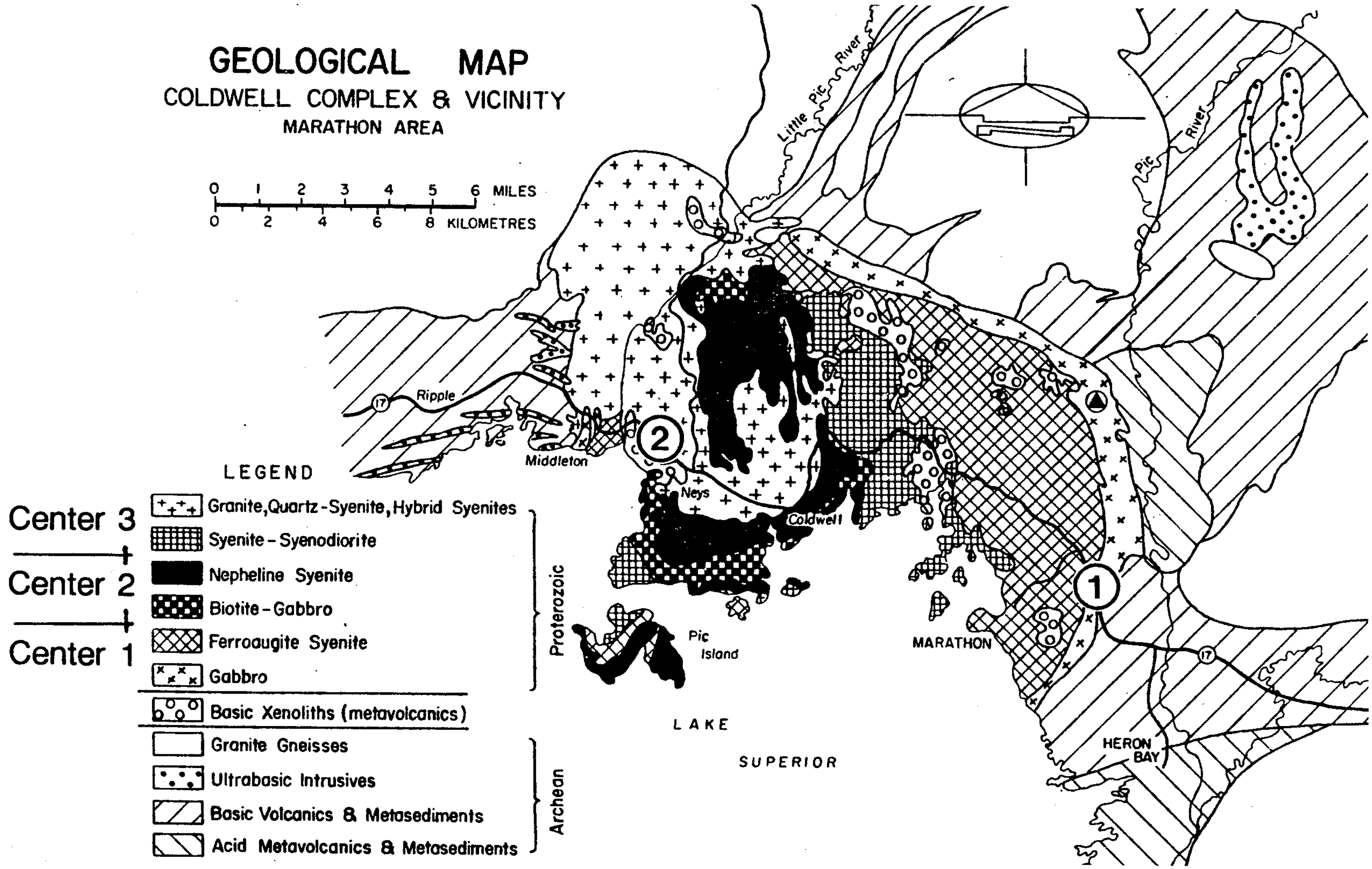


Figure 2 General geology map of the Coldwell Complex and vicinity (modified after Mitchell and Platt 1982).

STOP 2: LITTLE PIC RIVER BRECCIA ZONE This locality within the Coldwell Alkaline Complex has been described by Mitchell and Platt (1982):

"Parking Lot

To the southwest can be seen cliffs of xenolith-free ferroaugite syenite along the west bank of the Little Pic River. The river probably occupies a fault zone with the east bank representing a down-faulted block of Centers -2 and -3 rocks. To the south lie the Coldwell Peninsula and Pic Island. Densely wooded shores are alkali gabbro and nepheline syenite. The distant barren shores are syenites and quartz syenites.

Highway Cuts

The highway cuts on the north side of Highway 17 provide excellent examples of the complex multiple igneous breccias so characteristic of the Little Pic - Redsucker Cove block. Center 2 breccias consist of alkali gabbro and nepheline syenites similar to those exposed on the west side of the Coldwell Peninsula. These breccias are found as large xenoliths in the later Center 3 quartz syenite breccias. Xenoliths in the quartz syenite are oligoclase basalts. These show all stages of assimilation from relatively unaltered but sericitized basalt to almost completely digested xenoliths of amphibole-rich rocks. Development of clots of biotite and amphibole is a characteristic metasomatic feature of xenoliths. The oligoclase basalts probably are remnants of Proterozoic extrusive rocks which originally capped the complex.

These outcrops demonstrate conclusively that Center 3 quartz syenites are younger than Center 2 undersaturated rocks. Two types of lamprophyre can be found crosscutting the breccias:

(a) Porphyritic lamprophyre, characterized by greenish phenocrysts of Al-Cr-augite, possibly of high-pressure origin;

(b) Ocellar lamprophyre, characterized by ocelli of carbonate, quartz and fluorite."

Please also refer to Figure 3.

STOP 3: JACKFISH (STEEL RIVER) PILLOW LAVA

A roadcut exposure of Archean pillow lava occurs on the south side of the Trans-Canada Highway approximately 2 km west of the Steel River. This exposure affords an excellent three-dimensional view of this variolitic, pillowed unit, typical of a submarine environment. The pillows have well-preserved, fine-grained selvages with cusps, with some locally developed hyaloclastite in the interpillow spaces. Local flows young to the east-southeast. Lava tubes and quartz-filled, "drain-away" cavities occur locally.

Varioles are generally present within a 20 cm envelope adjacent to the selvages. The varioles commonly display a radiating or fibrous internal structure, although metamorphism and deformation have destroyed many of these primary features. Both the varioles and the pillow selvages are white to light green, while the pillow matrix varies from a medium to dark green. The varioles appear to have an affinity to coalesce, producing clots and layers. The term variole itself has been commonly used in defining a pea-sized spherule in basic volcanic rocks. Past workers have attributed the development of spherules, spherulites and varioles to the devitrification of glass around scattered nuclei or centres due to rapid crystallization of a viscous magma or melt (Williams et al., 1954) or in response to undercooling of the liquid during cooling (Fowler et al., 1987).

STOP 3: MORLEY PYRITE OCCURRENCE: ARCHEAN SULPHIDE-FACIES IRON FORMATION

The Morley pyrite occurrence is located on the north shore of Lake Superior, approximately 4 km southeast of Schreiber (Fig. 4a). This occurrence is an extremely well-preserved, pyrite- and chert-rich sedimentary deposit situated within a late Archean metavolcanic suite. The _5m thick, lenticular chemical sedimentary unit consists of Algoman-type, sulphide-facies iron formation, underlain stratigraphically by intermediate flows and pyroclastic rocks and overlain by minor turbidites and mafic flows (Schnieders, 1987) (Fig. 4a and 4b).

The Morley occurrence has most recently been described by Fralick et al., (1989):

"The chemical precipitates of the deposit formed during a hiatus in eruptive activity, with pyrite and interbanded light and dark chert in the lower part of the

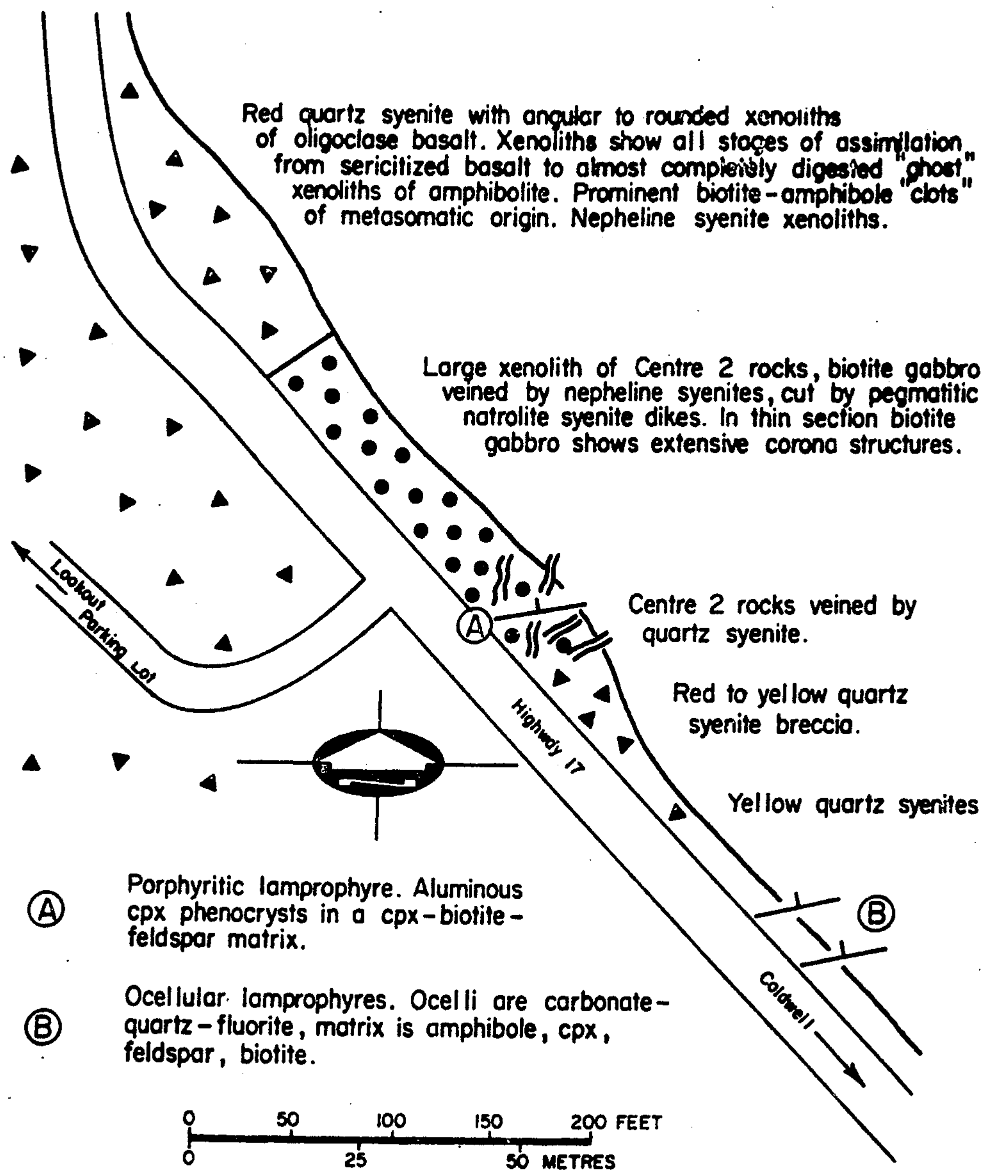


Figure 3 Geology of the Little Pic River lookout area (Mitchell and Platt 1982).

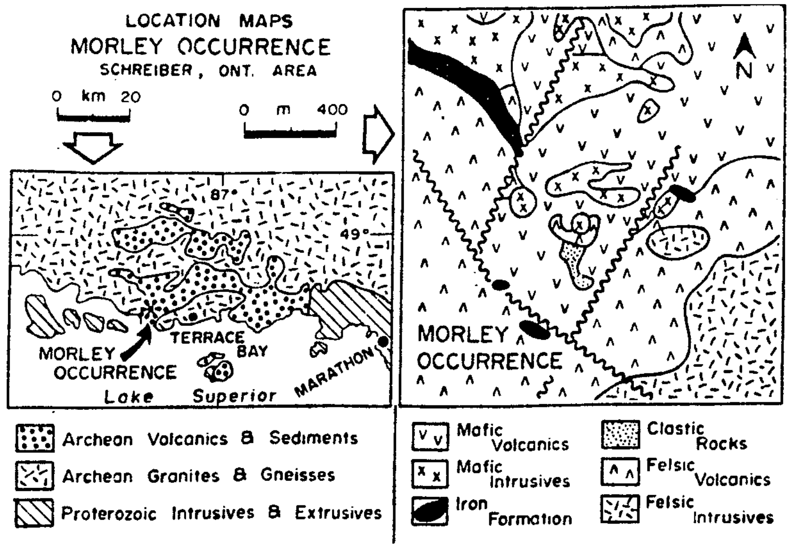


Figure 4a Location of the Morley pyrite occurrence (geology after Harcourt and Bartley 1939).

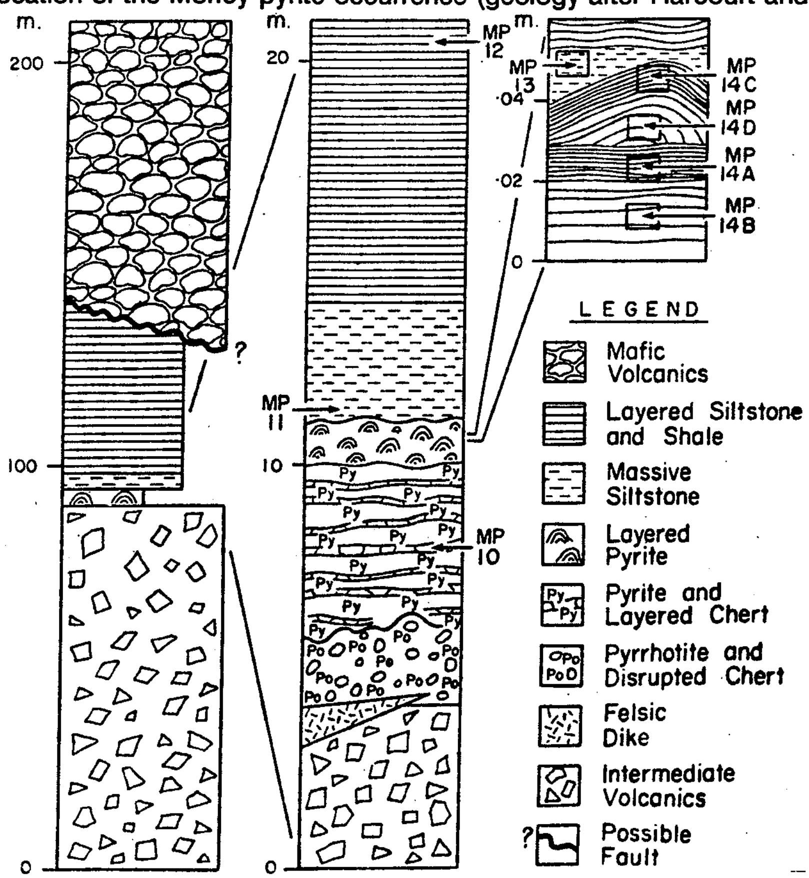


Figure 4b Generalized stratigraphic sequence, Morley occurrence (Fralick et al. 1989).

chemical unit, and bedded to laminated pyrite in the upper part. Black, carbonaceous mudstone and thin-bedded, fine-grained(DE) turbidites are also present. Within the pyrite rich upper part of the sequence, a variety of bedding structures is developed; one five cm-thick section (right side of Fig. 4b) was investigated in detail.

The upper pyritic unit contains cm-scale layers of pyrite separated by mm-scale layers of carbonaceous, cherty mudstone. The pyrite layers contain delicate internal laminations of pyrite and carbonaceous chert from 0.02 to 1 mm thick. Near the tops of individual pyrite laminations, the proportion of chert and disseminated clastic debris is greater. The mm-scale mudstone layers drape irregular pyritic surfaces, thin over colloform pyrite domes and thicken in flanking depressions. Commonly the mudstone layers thin to partings only a few hundred micrometres thick, or pinch out entirely where the top of a pyrite dome contacts the base of the overlying pyrite layer. Thin section and EDS data indicate that the mudstone layers contain quartz, muscovite, albite, carbonate, chlorite, disseminated pyrite, iron oxides and carbon.

The largest colloform pyrite dome observed is 2.5 cm high by 5 cm long, and contains about 60 percent pyrite laminations. The outer half of this dome is separated from underlying (and concordant) laminations by a 0.4 mm band of carbonaceous mudstone. Both halves exhibit thinning-upward successions of about 30 laminations. Thicker, basal, pyritic laminations in the successions are free of chert, clastic debris and carbon, whereas such material is common in the thinner overlying laminations. The top of the dome is overlain by massive pyrite that thickens over the flanking depressions. The systematic change in lamination thickness...is common in the Morley pyrite. However, some pyritic layers (either domed or flat) contain thin, monotonously similar laminations, and rare layers contain laminations that thicken upward.

Flat, finely laminated pyrite crusts, where traced laterally, commonly bend upward to form domal structures that may coalesce. Elsewhere, laminations in the domes intersect pyritic substrates sharply and at a high angle. Many colloform structures contain internal, low-angle discordant contacts, with individual layers or bundles of laminations truncated at angles of less than 20^{o} by overlying layers. These truncations generally occur near the outer edge of a dome. The outer portion of some domes also exhibits smaller, pyritic overgrowths resembling potato eyes. In these overgrowths, laminations rise abruptly and at a high angle from the surface of the underlying domal structure; after a few millimetres, they bend sharply and become parallel to the underlying laminations of the dome. This produces a highly asymmetrical overgrowth.

The domal structures generally are oriented in a convex-upward direction, with only a minor percentage exhibiting reversed directions. The latter appear to have sagged downward into underlying muddy sediments. Some pyritic layers contain contorted laminations that in extreme cases are bent into small, recumbent folds. In addition to such small-scale load and slump structures, pyritic layers display variable degrees of brecciation, some of which may be post-burial. However, thin lenses (< 1 cm thick) of pyritic gravel locally are present within the depressions between colloform pyrite domes, together with hemispheroidal fragments, that probably represent toppled material from the domes. This indicates that at least some brecciation took place on the seafloor."

In addressing the deposit's genesis, Fralick et al., (1989) state:

"A modest hydrothermal signal appears to be preserved in the pyritic sediments. Sulfur isotope ratios of sulfides and whole rocks from the Morley occurrence can be accounted for by hydrothermal leaching of magmatic sulfide from the host volcanic rocks, together with a component of heavier sulfur derived from the reduction of seawater sulfate. We interpret each lamination in the pyritic beds as reflecting a short-term hydrothermal injection into a stagnant bottom layer of water. Upward-thinning cycles of laminations, indicative of gradual, but oscillatory waning of precipitation, correspond to medium-term hydrothermal events. The domal to irregular structures of the Morley pyrite provide evidence for the existence of relatively deep-water organic mats during chemical precipitation. The carbon in the pyrite laminations and associated mudstones is interpreted as a relic of this organic activity, which may have been localized around hydrothermal vents. The pyritic layers in the Morley deposit seem most analogous to sulfidic sediments in the Red Sea brine deeps in terms of their finely laminated nature and sulfur isotope and REE compositions, but differ in their monomineralic nature and possibly biogenic structures."

GEOLOGICAL SETTING OF THE WINSTON LAKE MASSIVE SULPHIDE DEPOSIT

P.W.A. Severin, F. Balint, and R. Sim
Minnova Ltd.
2606 Victoria Ave. East,
Thunder Bay, Ontario.
P7C 1E7

INTRODUCTION

The Winston Lake deposit is located 145 km NE of Thunder Bay, Ontario (Fig. 1) in the Big Duck Lake volcanic sequence (Fig. 2) within the Archean Wawa Subprovince in the Superior Province of the Canadian Shield (Severin and Balint, 1985).

Geological mapping (Pye, 1964) defined a westerly to northwesterly trending volcanic belt, the Big Duck Lake volcanic sequence (Fig. 3), comprised of three parts:

1) a lower part consisting mainly of "metasediments";

2) a central part consisting of north and northeasterly facing pillowed mafic volcanics;

3) an upper part consisting of interbanded metavolcanic and metasedimentary rocks. A large lopolithic-shaped gabbro sill intrudes the contact between the lower "metasediments" and the overlying mafic volcanic rocks. A small massive sphalerite occurrence, the Zenith deposit, was discovered within this gabbro sill by surface prospecting in 1879 (The Mining Review, 1899). No significant work was done on the property until Zenmac Metal Mines Ltd. was incorporated in 1952 to further explore and develop this deposit. The Zenith zinc deposit was mined during the late 1960's and produced 165,000 tonnes of ore grading 16.5% zinc.

Corporation Falconbridge Copper (CFC)

In October, 1978 CFC completed reconnaissance geological and lithogeochemical surveys to assess the exploration potential of the Zenith mine area (Mattinen, 1979; Severin, 1979). The lower "metasediments" (Pye, 1964) were identified as altered felsic volcaniclastic rocks with a lithogeochemical signature (Fig. 4) similar to that recognized at CFC's Sturgeon Lake mine, where the ratio Zinc (ppm)/(% Na_20 x 10) and (% K_20/% Na_20) x 10 within the felsic volcanics 1200 metres west of the Zenith deposit, adjacent to the west contact of the gabbro. Forty-two claims were staked to the west and north of the old Zenith mine and an option agreement was negotiated on the thirty-two Zenith mine property.

In 1979-80 detailed (1:5000) geological, lithogeochemical, magnetometer, VLF and Maxmin II surveys were completed. The geophysical results were disappointing but the geology and lithogeochemical surveys discovered zones of cherty bedded ash within the calc-alkaline felsic volcanic strata and delineated the zone of hydrothermal alteration previously identified by the reconnaissance work (Pirie, 1979; Unger, 1980).

Detailed geological mapping defined the lower "metasediments" as a sequence of calc-alkalic (Fig. 5) felsic volcanic flows and volcaniclastic rocks and the overlying mafic rocks as a series of magnesium to iron-rich massive and pillowed tholeiitic basalt. The contact between these two contrasting sequences is occupied by a composite sill-like gabbro intrusion. The Zenith zinc deposit occurs at a transition between gabbro and meta-pyroxenite phase of the gabbro. Areas of cherty ash occur at the top of the felsic volcanic sequence 1200m west of the Zenith deposit, and close to the gabbro contact. The volcanic rocks have been metamorphosed to amphibolite grade.

Hydrothermal processes altered the felsic rocks, forming well defined zones of strong Na_2O depletion, FeO and MgO enrichment and locally moderate to strong zinc enrichment. At the prevailing amphibolite metamorphic grade this alteration has formed a mineral assemblage of cordierite-anthophyllite +/-garnet +/- staurolite +/- sillimanite.

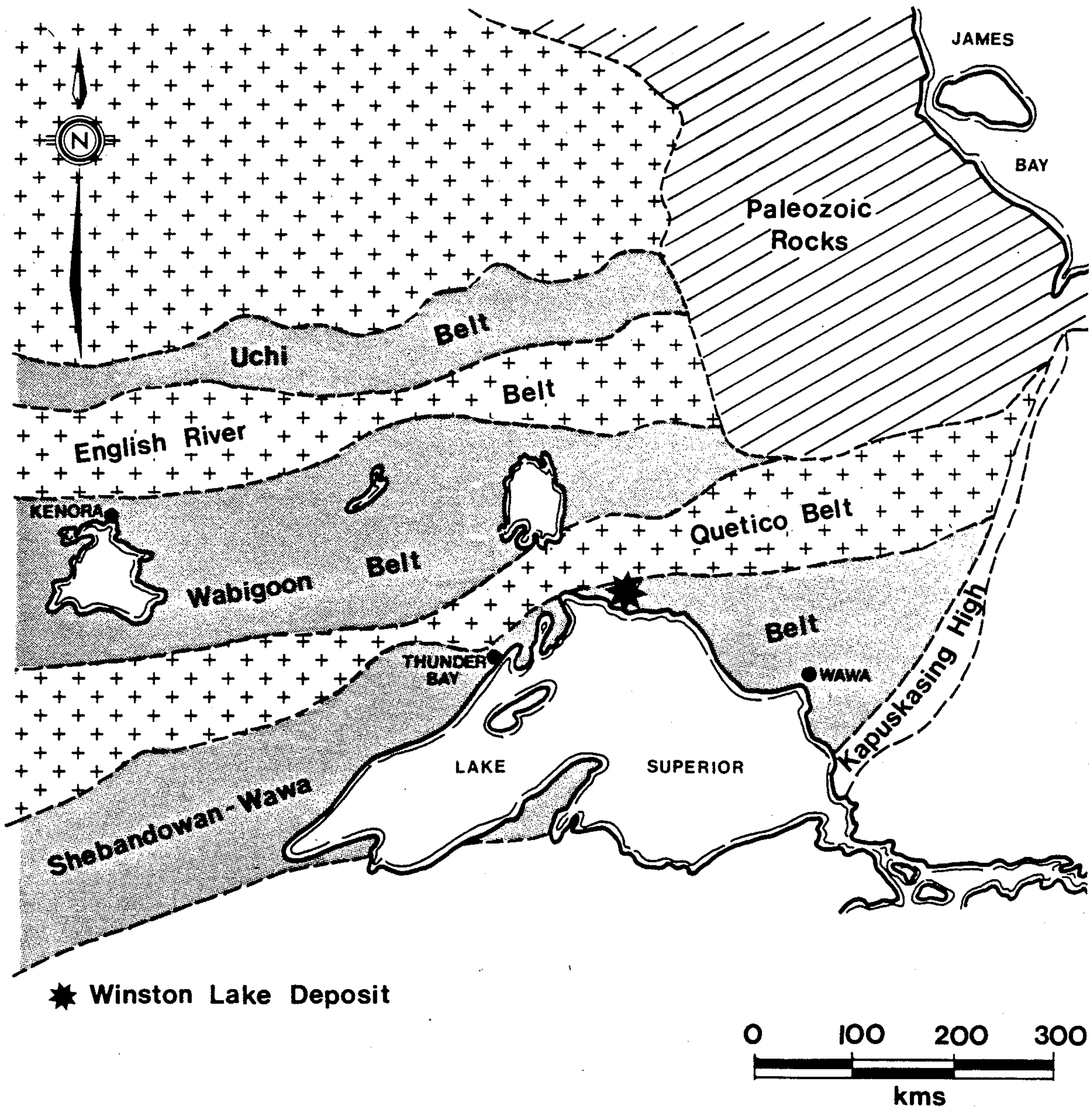


Figure 1 Location Map.

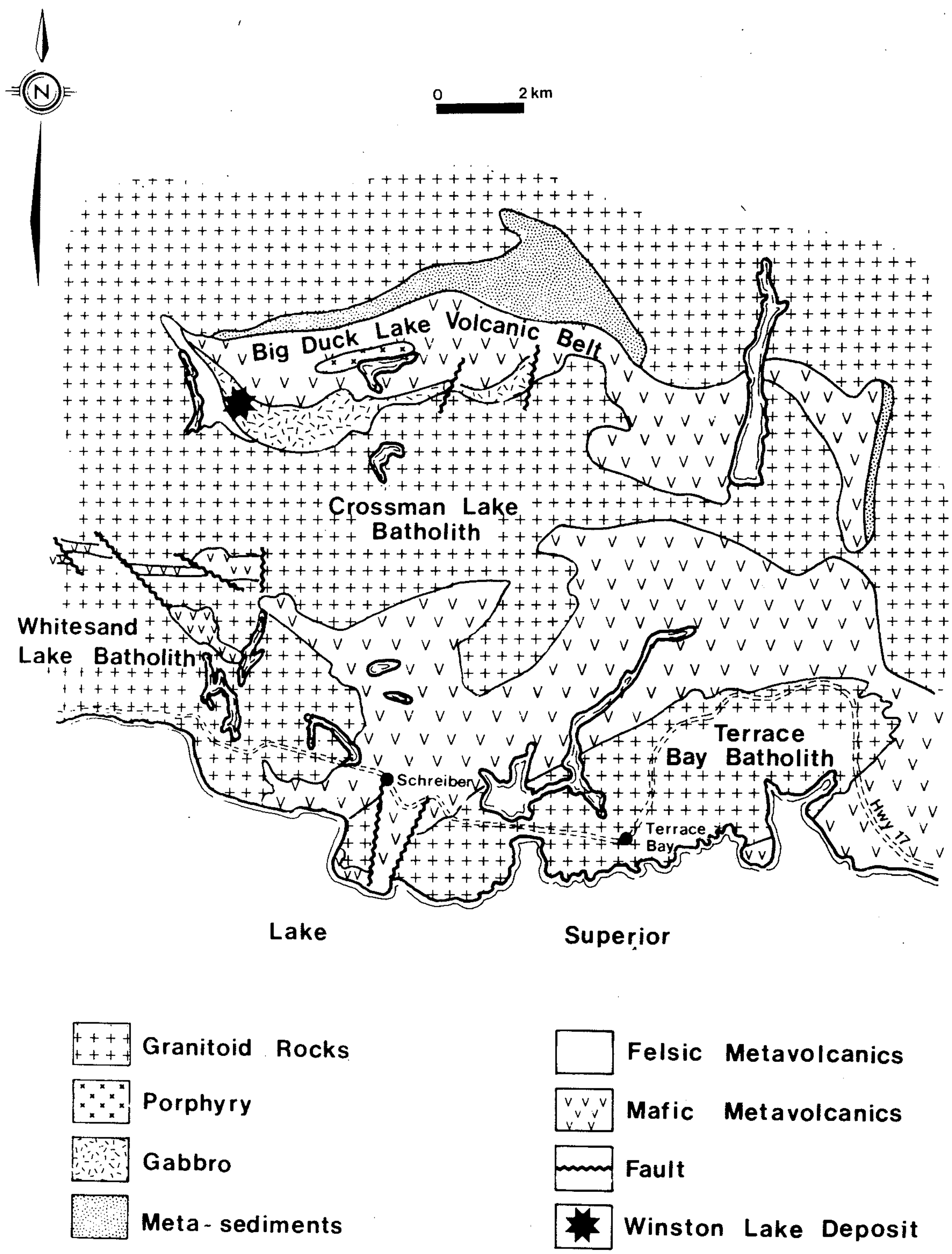


Figure 2 Regional Geology.

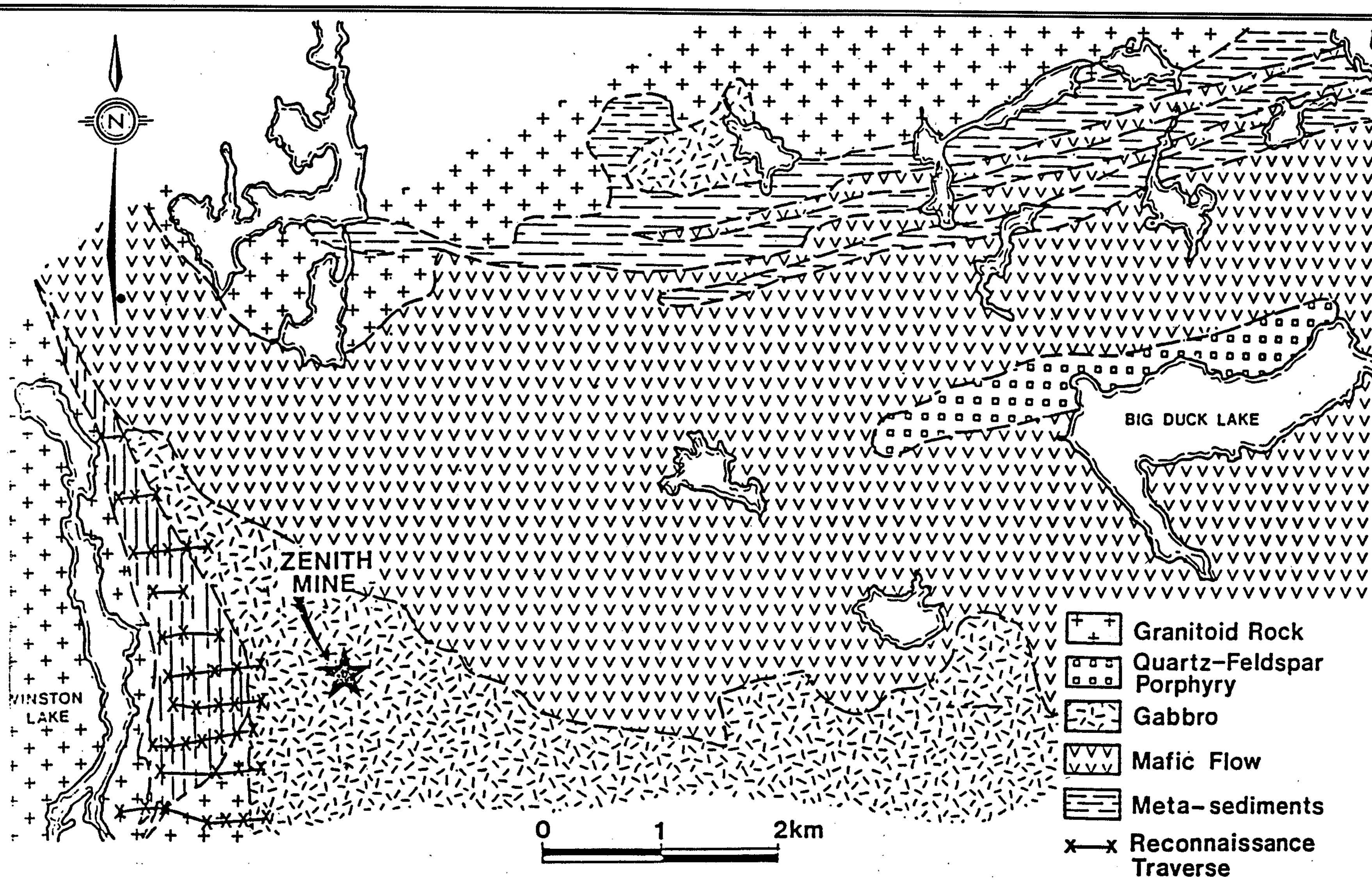


Figure 3 Big Duck Lake Volcanic Belt (after ODM Map 2023).

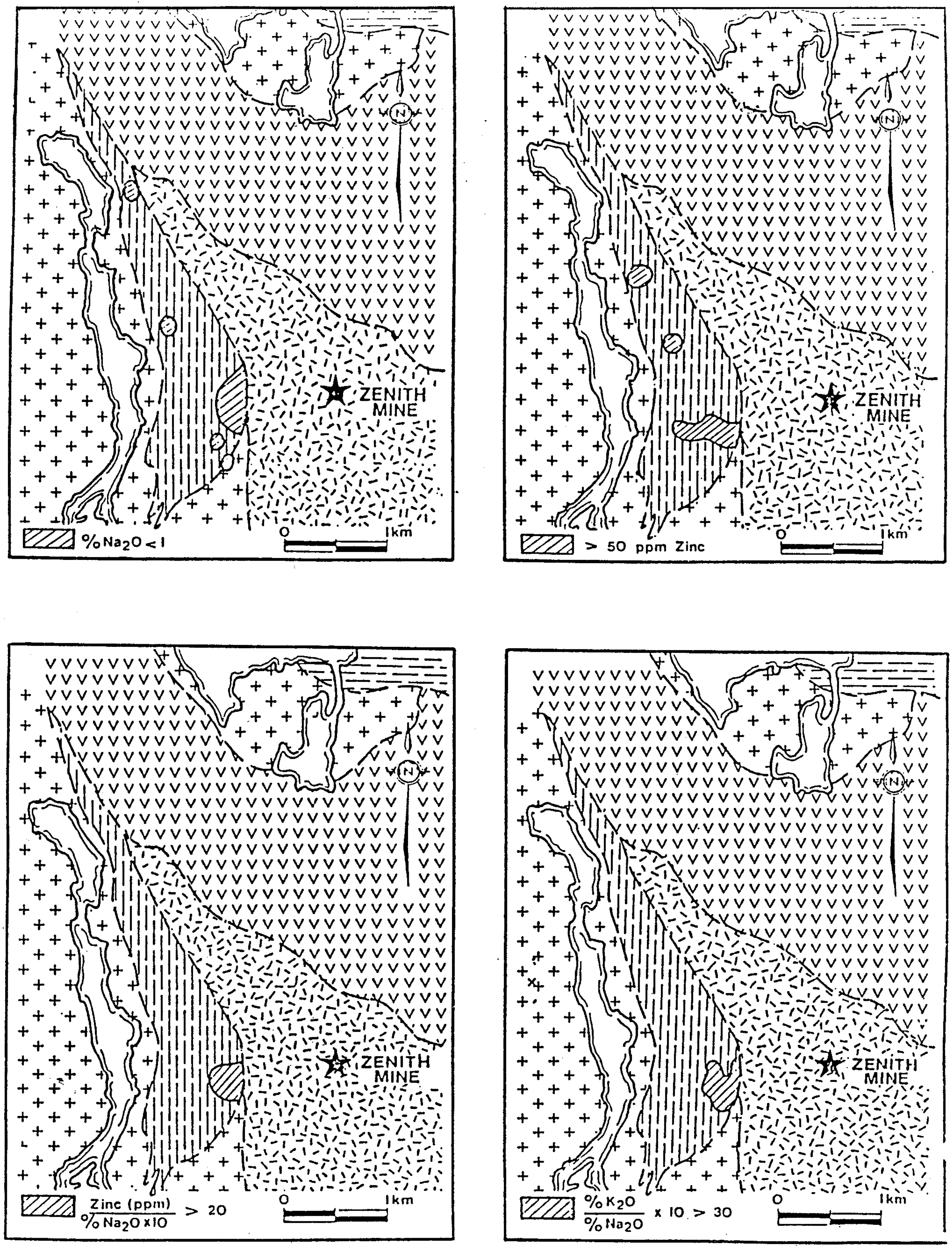


Figure 4 Reconnaissance lithogeochemical Results.

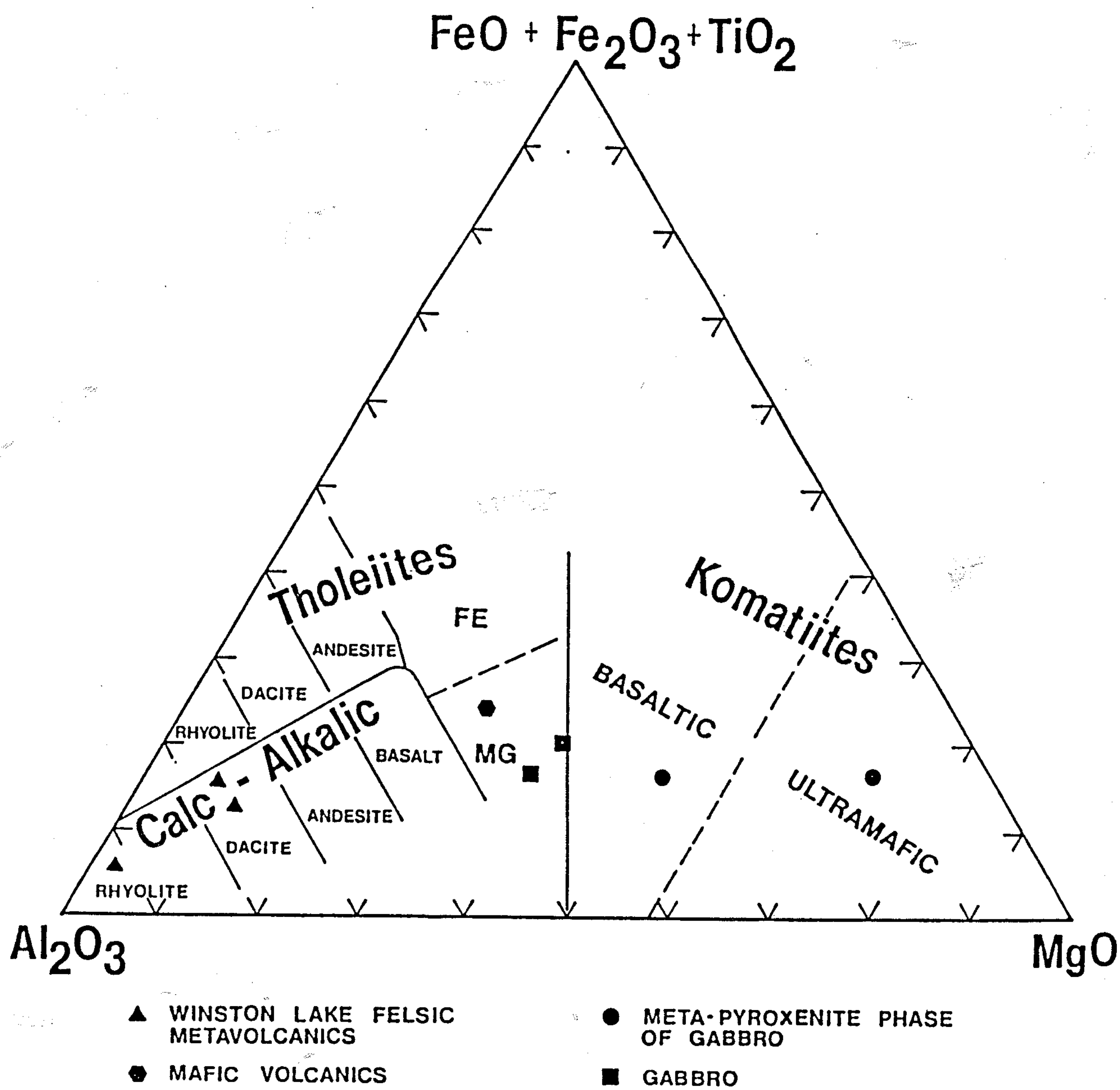


Figure 5 Jensen Plot.

Analysis of altered and unaltered Winston Lake felsic volcanic rocks plotted on an unfolded tetrahedron indicate trends (Fig. 6) similar to those described by Riverin and Hodgson, (1980) for the Millenbach deposit, near Noranda, Quebec. The geological map and lithogeochemical results are illustrated in figures 7 and 8.

The Zenith zinc-rich massive sulphide deposit is hosted by gabbro, an unusual occurrence in the Canadian Shield. This gabbro sill intrudes the contact between underlying altered calc-alkaline felsic volcanic rocks and overlying unaltered tholeiitic pillowed mafic volcanics. Genetic interpretations by previous workers (Tanton, 1930, p. 193; Halet, R.C. in Pye, E.G., 1964, p42) suggested a vein type epigenetic source or a magmatic zinc model to explain the Zenith deposit. Corporation Falconbridge Copper's (now Minnova Inc.) work clearly defined a number of features consistent with volcanogenic massive sulphide deposition. The geological model was developed that presumed the Zenith deposit was a large xenolith of volcanogenic massive sulphides derived from a potential larger in-situ deposit located at the top of the felsic volcanic pile. Assuming a simple dilation process by the gabbro, the schematic section shown as figure 9 was proposed (Severin, 1980).

In 1981 eight diamond drill holes were completed. Four holes were drilled on property wholly-owned by CFC to investigate a pyritic horizon (3 holes) spatially related to the anthophyllite-cordierite zone within the felsic volcanics and a weak Maxmin II, VLF and magnetic anomaly (1 hole) which occurs within the gabbro northwest of the Zenith mine. The results were disappointing. The other four holes were drilled to test the geological model. They were drilled on the adjacent Zenith mine property to explore 125-250m down-dip of the cherty ash horizon which occurs at the top of the felsic volcanic package east of and overlying the cordierite/anthophyllite alteration. The results were encouraging; an exhalative horizon intersected by the holes contains up to 0.5% zinc over 4.3 metres.

Borehole pulse electromagnetic surveys (PEM), which had previously proven successful at CFC's Lac Dufault division in northwestern Quebec, were completed in May 1982 and detected a strong edge-type anomaly (Fig. 10). Directional PEM surveys indicated that a strong conductor was situated down-dip from the exhalite intersected by diamond drilling in 1981 (Balint, 1982).

Diamond drilling resumed on June 2, 1982 to follow-up the borehole PEM anomaly. On June 9th diamond drill hole ZO-5 intersected 2.1 metres of massive sulphides containing 1.10 % Cu, 19.11 % Zn, 22.2 g/t Ag and 0.73 g/t Au. This zone is located at the base of the gabbro sill, 300 metres below surface.

GEOLOGY

The Winston Lake calc-alkalic sequence, which hosts the Winston Lake deposit, is composed mainly of felsic volcaniclastic strata and flows, with minor intercalated sediment and mafic flows. Strata strike 340 degrees, dipping 50 degrees to the east.

The immediate hanging wall to the Winston ore body varies throughout the deposit. 40% of the surface area is in direct contact with the gabbro. The other 60% is composed of <1m to >20m thickness of bimodal ash, mm/cm banded mafic and felsic "cap" unit, typical of these types of deposits. The immediate footwall also varies throughout the deposit. Approximately 30% of the deposit sits on top of chlorite/biotite/anthopyllite/cordierite (altered mafic flow), while the other 70% is in contact with a unit composed of biotite altered "clotted" rhyolite (a bimodal volcaniclastic unit similar to the hanging wall cherty ash unit but generally banded on a coarser scale). The contacts of the ore zone are generally very sharp and commonly marked by a 1 to 3mm chloritic seam.

Original mineable reserves (all categories, including 20% dilution at zero grade) reported November 1, 1987 were 3,076,339 tonnes at 1.00 % Cu, 15.60 % Zn, 30.87 g/t Ag and 1.02 g/t Au.

The ore zone varies in thickness from 2m to more than 20m (horizontal) with an average of 7m. There are two main ore types: A massive to locally banded, fine to medium grained homogeneous mix of sphalerite, pyrrhotite, pyrite and chalcopyrite with 10-20 % included fragments ranging from <1cm-5cm. This is considered "low-grade" with values generally in the 7-14 % Zn range. The second "high-grade" ore type is predominantly composed of medium to coarse grained sphalerite, massive to locally banded with chalcopyrite and/or pyrrhotite. Grades in this material can be as high as 54 % Zn. An overall zonation of ore types is not presently apparent, possibly due to multiple source points.

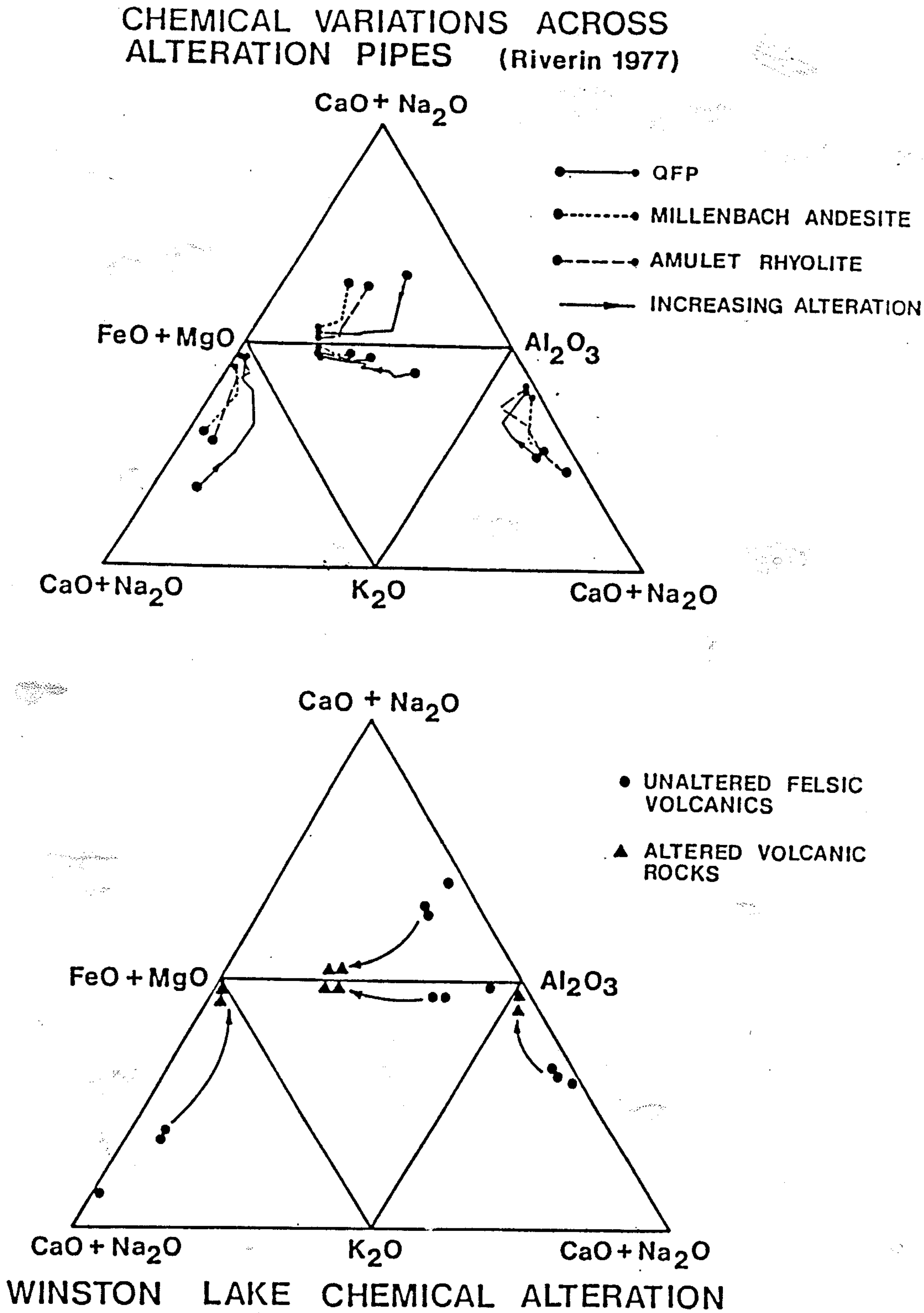


Figure 6 Unfolded tetrahedron.

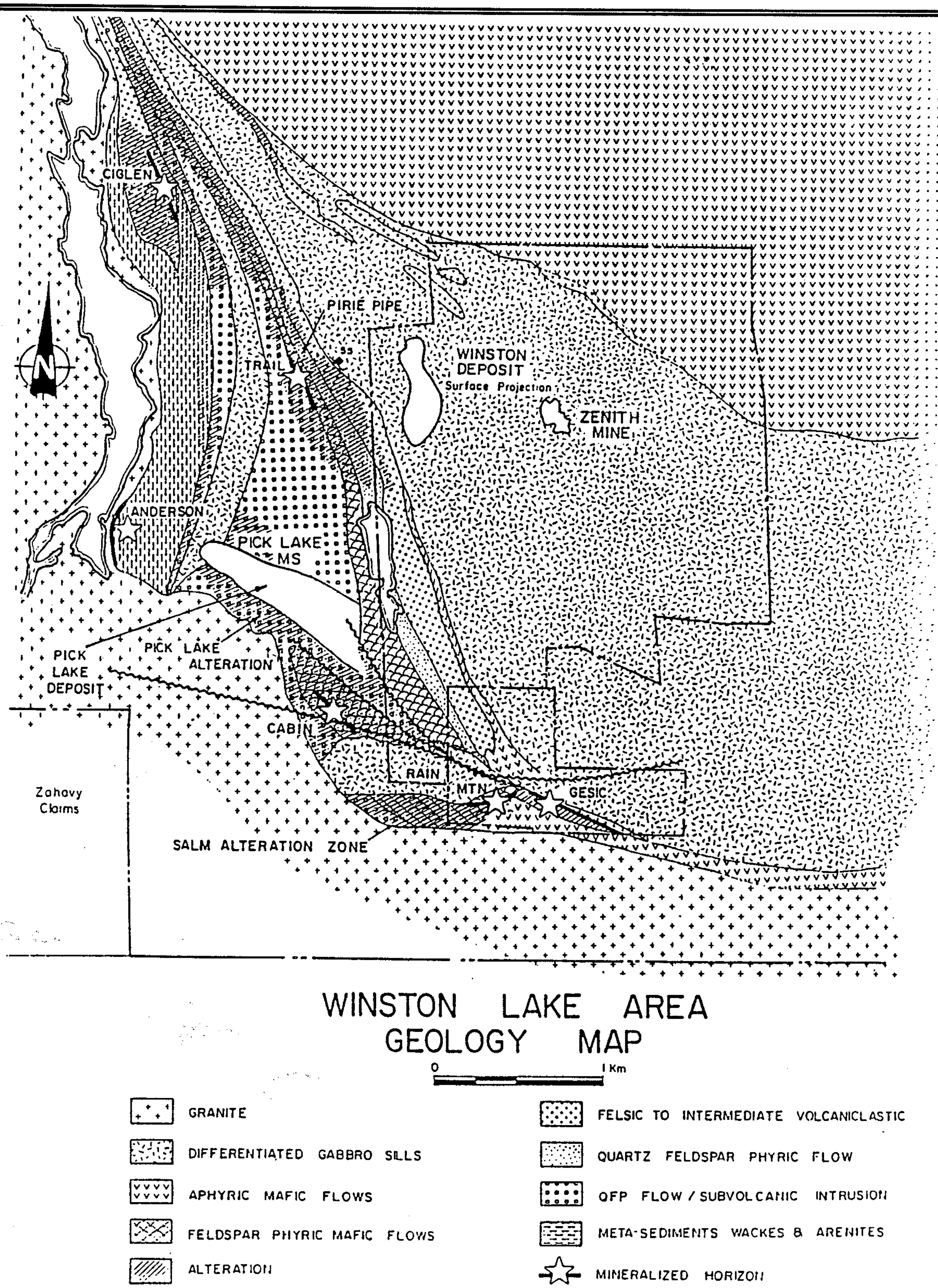


Figure 7 Winston Lake area geology map.

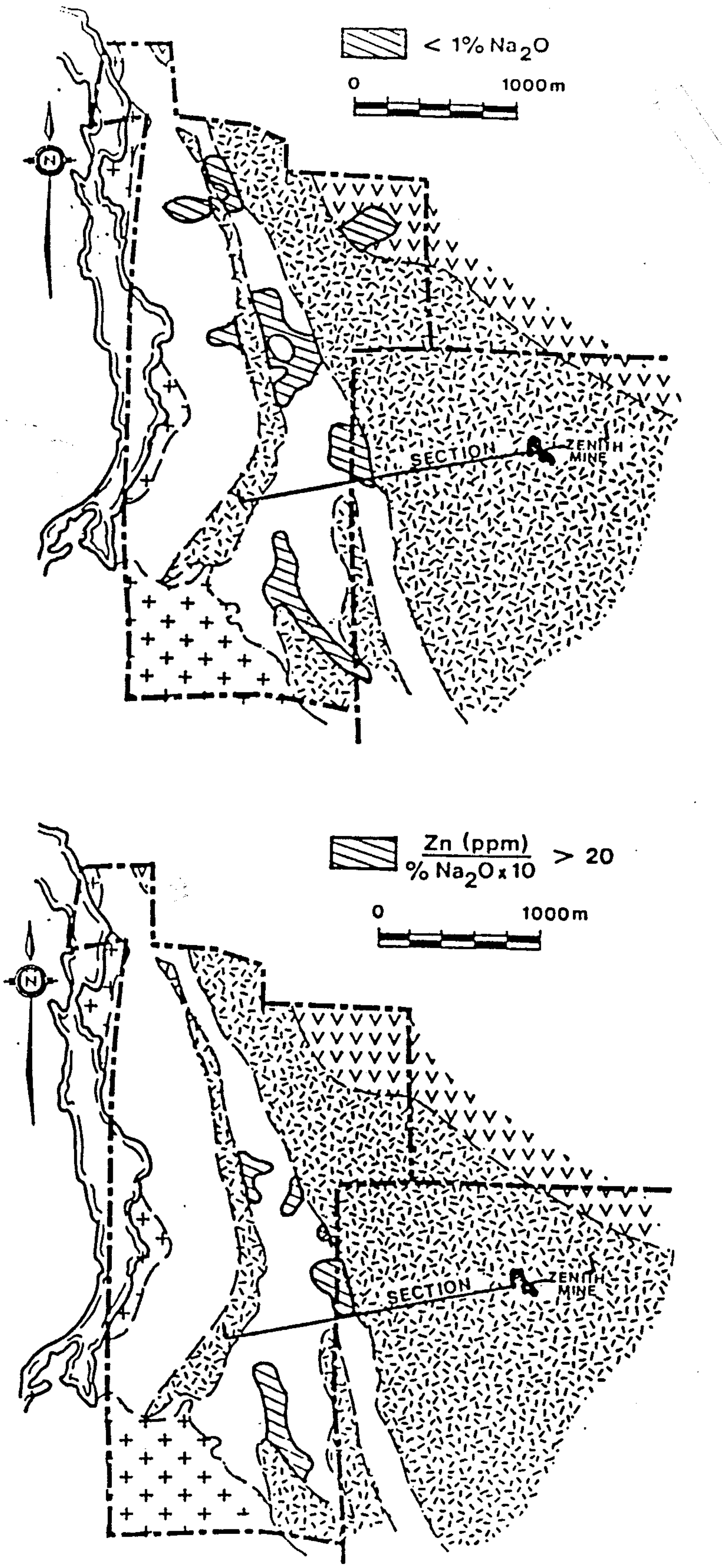


Figure 8 Detailed lithogeochemical results

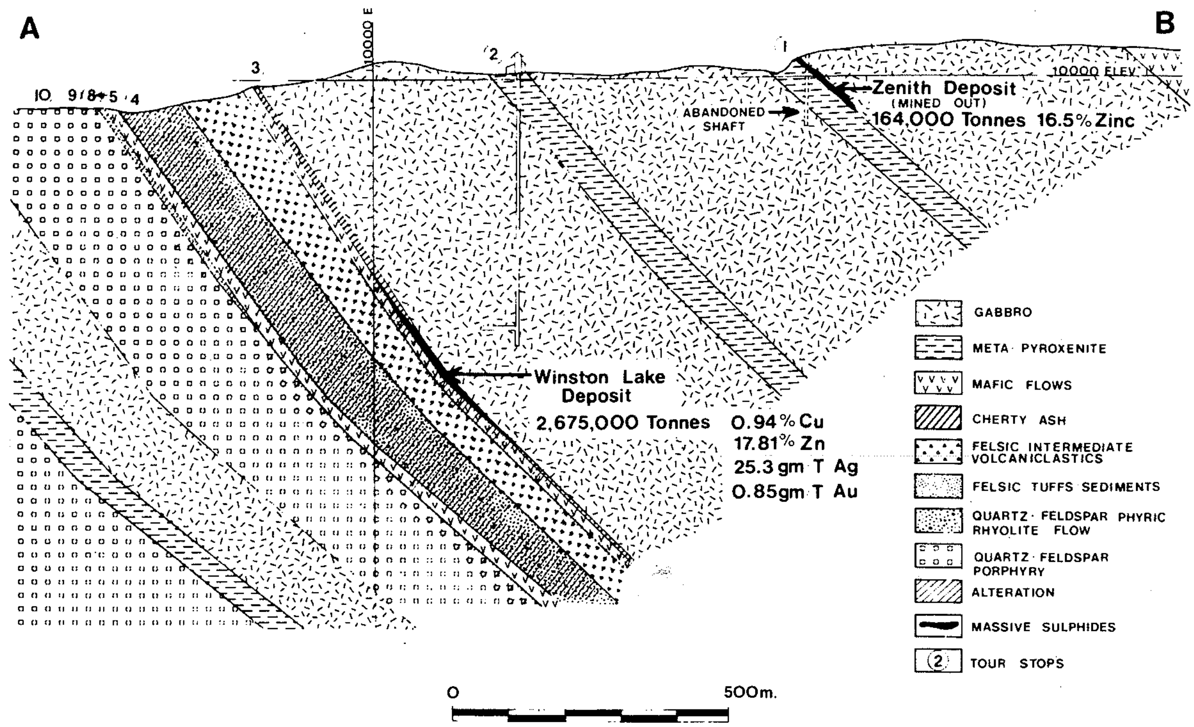


Figure 9 Composite section of the Winston Lake deposit.

SURFACE TOUR

The tour stops are shown on figures 7, 9 and 11.

STOP 1: ZENITH DEPOSIT (Fig. 7 and 9).

Exploration in the area dates back to at least the 1870's. The Zenith Zinc Deposit is reported (The Mining Review, 1899) to have been discovered in approximately 1879.

Minor production is reported for the period 1899 to 1902 but no serious exploration was initiated until Zenmac Metal Mines Ltd. acquired the property during the early 1950's. Diamond drilling indicated a mineral inventory of 128,000 tonnes grading 23% Zn and 0.25% Cu but planned development was suspended due to low ($0.105/lb) zinc price. By the autumn of 1963 the forecasts were optimistic. A decision was made to proceed to production and a 21.8 km gravel road was completed by July of 1964. Shaft sinking commenced in September 1964 and was completed to 425 feet (130m) by November. Three levels were established at depths of 150, 275 and 400 feet (46, 84 and 122m) during the period April 1, 1966 to April 29, 1970, 164,000 tonnes of 16.5% Zn were milled.

The two Zenith deposits (Fig. 7) occur at a transition between gabbro and meta-pyroxenitic phase of the gabbro (Fig. 9). The following description is summarized from Pye (1964) and Halet (1965). The deposit has an overall lenticular shape, dipping 35-45

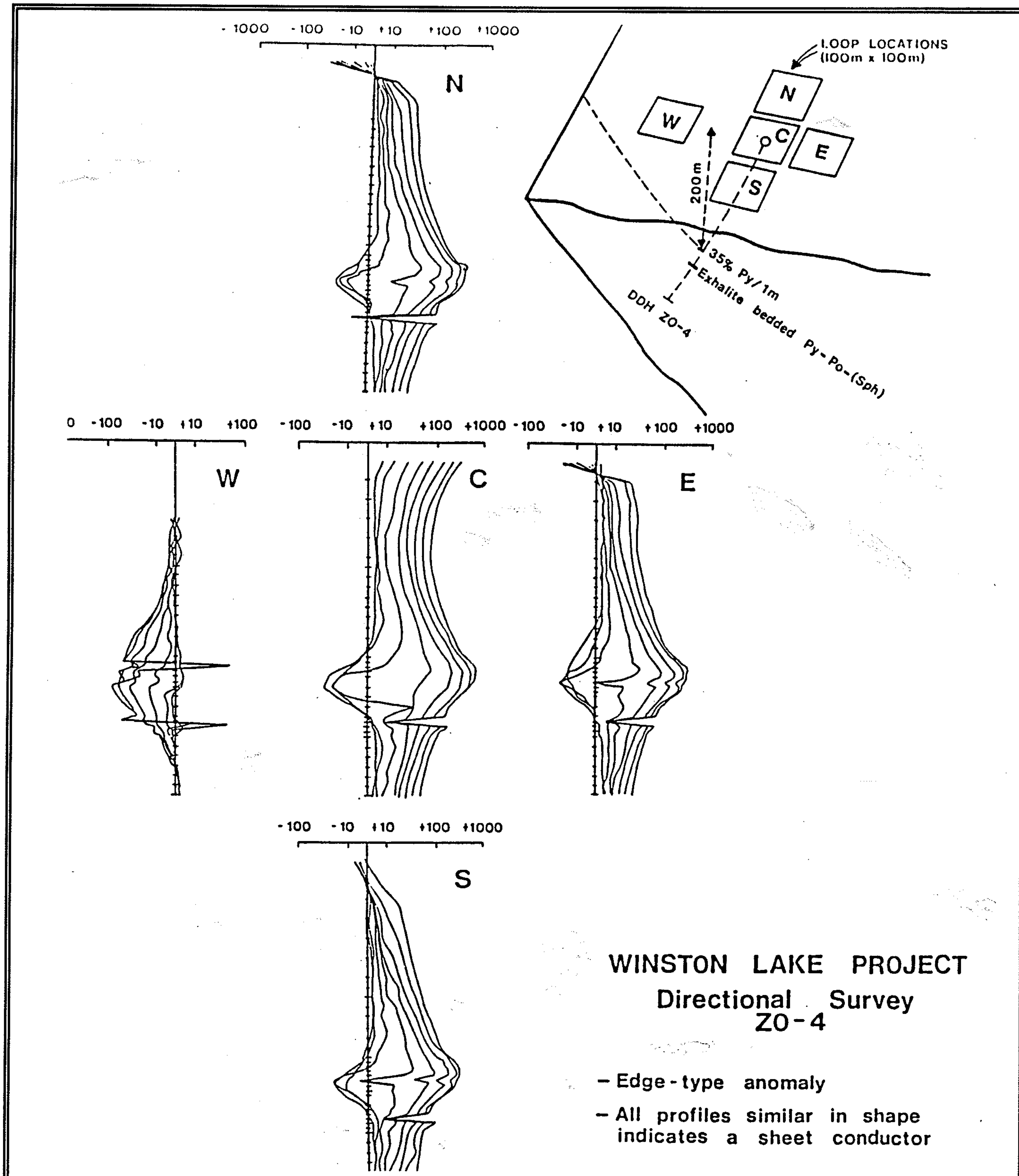


Figure 10 Borehole Pulse EM Survey.

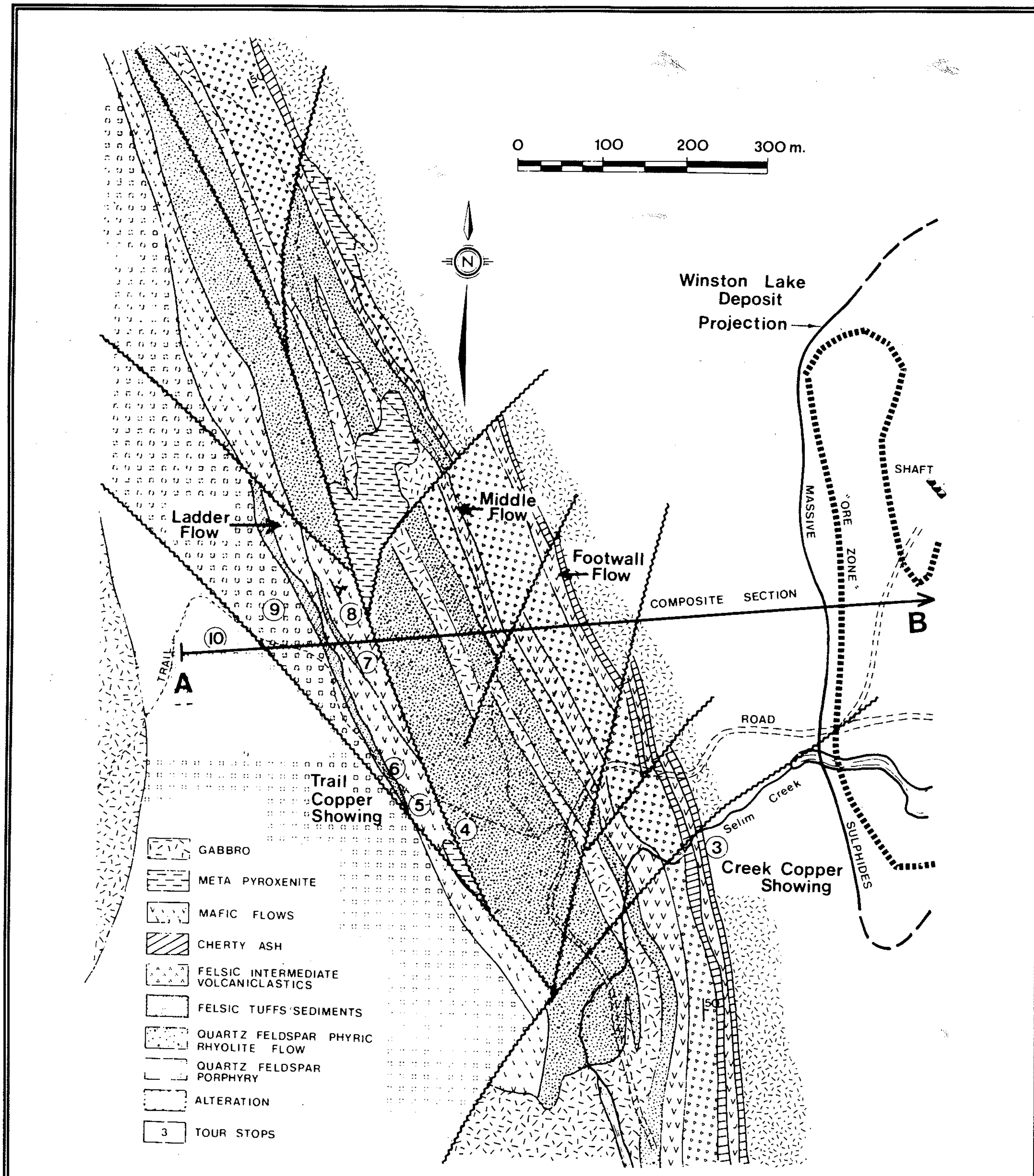


Figure 11 Detailed surface geology of the Winston Lake deposit.

degrees NE and varying in thickness from several centimetres to 13.4 metres. Three types of mineralization have been reported:

1. Irregular lenses of coarse-grained sphalerite with minor pyrrhotite, pyrite and chalcopyrite.

2. Dyklets or tongues of sphalerite along fractures in host rock and interstitial to fragments of gabbro.

3. Sparse disseminations of sphalerite and other sulphides in the host rock adjacent to the massive sulphide occurrences. This mineralization was divided into two types by Halet (1965):

a) Zinc ore composed of massive sphalerite with minor pyrrhotite and chalcopyrite.

b) Copper ore composed of massive pyrrhotite with minor chalcopyrite. The copper ore consisted of up to 1.5% Cu.

Stop 1, at Kenabic Lake, will provide access to meta-pyroxenite, gabbro and several open cuts (that date back to 1900) that will give some idea of the character of sulphides that comprised the Zenith Deposits.

STOP 2: DISCOVERY OF THE WINSTON LAKE DEPOSIT (Fig. 7 and 9)

This stop is 300m SSW of the Winston Lake headframe. It also marks the location of the collar of discovery hole ZO-5 which intersected 2.1m of 1.10% Cu, 19.11% Zn, 22.2 g/t Ag, and 0.73 g/t Au on June 9, 1982.

STOP 3: GABBRO - WINSTON LAKE HORIZON THIN MAFIC FLOWS - FELSIC VOLCANICLASTIC ROCKS (Fig. 7, 9, and 11).

The Winston Lake volcanic package has been defined by Pirie, (1979), Unger, (1980), Balint, (1982), Balint and Salm, (1983), Morton, (1984) and Severin and Balint, (1984). It consists of an eastward facing monoclinal sequence of intercalated felsic and mafic metavolcanic strata underlain by metasedimentary rocks. All strata dip from 45 to 65 degrees to the east. Although metamorphosed to lower amphibolite grade, the prefix meta will not be used in this summary.

For the purposes of comparison, partial whole rock and trace element geochemistry is presented in Table I. The top of the predominantly felsic sequence varies from 35 to 125 m thick and is composed of several volcaniclastic units (pyroclastic flows and/or debris flows) that consist of more massive basal sections overlain by finely laminated ash deposits. The upper laminated unit is referred to as the "Winston Lake Horizon" (WLH). The Winston Lake massive sulphide deposit occurs approximately 450 m down dip along this horizon. The WLH consists of a series of very finely laminated, felsic beds, cherty beds and to a lesser extent, more mafic-rich beds.

At least two thin (up to 20 m) massive to possibly pillowed, fine to medium-grained, mafic flows and associated mafic sedimentary units occur within the felsic-volcaniclastic package. In the vicinity of the Winston Lake deposit, both flows are intensely altered to cordierite-anthophyllite-biotite.

The basal beds of the volcaniclastic unit consist of 25-70% lithic fragments of which 40-75% are mafic with the remainder quartz-eye bearing felsic fragments. These elongate flattened fragments are up to 4 x 15 cm in size with some felsic fragments up to metre size. Laminated sections vary in thickness from 15 mm to 7 metres with individual beds ranging from 1 mm to 15 cm. Where hydrothermal activity has penetrated this sequence, the mafic component, the smaller felsic fragments and ash component have been altered to a biotite-cordierite assemblage. The larger felsic fragments tend to be preserved in a grey biotite-cordierite matrix. The flattened mafic volcanic clasts are often rich (3%) in iron sulphides and magnetite.

The mafic sill that partially "sill-out" the Winston Lake deposit and hosts the Zenith deposit is a composite intrusion with meta-gabbroic and meta-pyroxenitic phases. The gabbro grades into meta-pyroxenite which forms two distinct layers within the sill as illustrated in figure 9.

STOP 4: ALTERED QUARTZ-FELDSPAR PHYRIC RHYOLITE FLOW.

One of quite possibly a series of quartz feldspar phyric rhyolite flows from 50 to 150 m thick, underlies the volcaniclastic sequence. Unaltered, this rock type is a massive pink to grey rhyolite with 10 to 20 %, 0.5 to 2 mm clear to grey quartz eyes and up to 20 %, 1 to 2

Table 1 Partial whole rock and trace element geochemistry of Winston Lake volcanics.

	Sample No.	SiO_2 %	Al_2O_3 %	FeO_T %	MgO %	CaO %	Na_2O %	K_2O %	TiO_2 %	P_2O_5 %	MnO %	Cu ppm	Zn ppm
	TBD												
Footwall Mafic Flow	4611	48.7	15.25	8.49	6.94	7.89	4.06	0.75	1.01	–	–	94	35
Altered Footwall Mafic Flow	3473	34.3	19.72	11.83	19.85	0.80	1.12	1.88	1.26	–	–	104	50
Middle Mafic Flow	3604	52.1	15.49	8.49	5.01	7.96	3.71	0.48	1.34	–	–	45	10
Altered Middle Mafic Flow	2747	51.2	14.69	10.00	11.57	0.11	0.54	1.39	0.89	–	–	109	43
Felsic-intermediate Volcaniclastics	3601	72.2	10.96	3.25	1.19	2.53	5.04	0.26	0.41	–	–	36	12
Altered Felsic-intermediate Volcaniclastics	3258	68.1	11.35	6.04	5.92	0.53	1.13	1.79	0.36	–	–	441	125
Quartz Feldspar Phyric Rhyolite Flow	3608	77.3	12.18	1.63	0.32	1.23	6.06	0.09	0.23	–	–	3	7
Altered Quartz Feldspar Phyric Rhyolite Flow	3282	74.9	10.32	3.63	6.19	0.62	0.47	1.13	0.27	–	–	83	17
'Ladder' Mafic Flow	3612	46.7	18.00	10.51	5.06	7.25	3.59	1.67	0.94	–	–	233	21
Altered 'Ladder' Mafic Flow	3264	49.6	14.35	19.29	7.65	0.34	0.52	0.19	0.82	–	–	42	76
Quartz Feldspar Porphyry	3615	74.6	11.43	2.22	0.91	1.88	4.16	0.65	0.30	–	–	5	23
Altered Quartz Feldspar Porphyry	3272	79.9	7.86	5.60	2.64	0.06	0.06	2.03	0.27	–	–	20	58
Gabbro	1643	50.9	14.30	11.60	7.94	11.10	1.85	0.30	0.76	0.07	0.18	–	–
Transition Zone	44	45.9	15.70	10.40	12.50	8.99	1.45	0.99	0.44	0.04	0.15	–	–
Pyroxenite	45	41.5	6.03	15.50	24.20	5.05	0.07	0.02	0.50	0.04	0.16	–	–

mm subhedral white feldspar laths in a massive aphanitic matrix. Locally the rocks are flow banded with grey to black biotite-magnetite rich streaks. Good exposures of this unit(s) occur along the road near the Cleaver Lake camp site. With increasing hydrothermal alteration, these rocks can contain:

1.QUARTZ, MUSCOVITE, BIOTITE + feldspar

2.QUARTZ, CORDIERITE, SILLIMANITE, BIOTITE KNOTS + staurolite + garnet

3.QUARTZ, CORDIERITE, ANTHOPHYLLITE + sillimanite + staurolite + garnet

STOP 5: "LADDER FLOW" - MAFIC LAVA LOBES - PERVASIVE AND INCIPIENT ALTERATION.

The "Ladder Flow" is a 20 m- to 200 m-thick series of mafic rocks consisting of thin sheet flows, thicker massive flows and flow lobes to units of pillow lava, pillow breccia and hyaloclastite. A felsic interflow sedimentary unit is present locally between individual mafic flows. These mafic rocks underlie the quartz feldspar pyritic rhyolite flows of Stop 4. Unlike the mafic rocks of the rest of the Big Duck Lake Volcanic terrain, these mafic rocks are highly feldspar phyric, commonly containing 10 to 30%, 1 to 5 mm tabular laths of plagioclase. At this stop, elongate pillows probably representing feeding tubes of lava lobes and massive feldspar phyric flows are partially altered to anthophyllite-biotite-cordierite + garnet assemblages. The alteration shows a preference for the more hyaloclastic margins of lava lobes. Remnant patches of a laminated rusty siliceous material are exposed along the base of the outcrop. This represents an interflow sedimentary rock which is better exposed at Stop 7.

From Stop 6 to 7 the contact between the base of the "Ladder Flow" and these altered sedimentary rocks is exposed. Outcrops include unaltered, mafic flow bottom breccia, unaltered massive flow with epidote-silica patches, partially altered broken pillow breccia and unaltered pillowed flow.

STOP 6: TRAIL SHOWING - ALTERED MINERALIZED FELSIC SEDIMENTS/TUFF - ALTERED "LADDER FLOW".

Similar rock to that which underlies the "Ladder Flow" at Stop 6, occurs between individual mafic flows. At this site, a 0.5 m- to 1.0 m-thick, altered, predominantly felsic sedimentary unit containing pyrite, pyrrhotite and minor chalcopyrite occurs between two altered mafic flows. The base of the interflow sediment is altered to a garnet-rich assemblage. Since much of the outcrop represents a contact dip slope, this garnet-rich material outlines amoeboid shapes which reflect the texture of the underlying mafic flow top.

STOP 7: ALTERED MINERALIZED INTERFLOW SEDIMENTS - FLOW TOP "LADDER FLOW": Similar rocks to those which underlie the "Ladder Flow" at Stop 6, occurs between individual mafic flows. At this site, a 0.5 to 1.0 m thick, altered, predominantly felsic sedimentary unit containing pyrite, pyrrhotite and minor chalcopyrite occurs between two altered mafic flows. The base of the interflow sedimentary rock is altered to a garnet-rich assemblage. Since much of the outcrop represents a contact dip-slope, this garnet-rich material outlines amoeboid shapes which reflect the texture of the underlying mafic flow top.

STOP 8: ALTERED PILLOWED "LADDER FLOW".

This outcrop is an exceptional example of a hydrothermally altered and subsequently metamorphosed basaltic pillow lava. Well formed, undeformed meter-scale pillows indicating a facing direction to the east are well preserved. The centers of these pillows are composed of coarse anthophyllite and cordierite crystals. The bladed to sheaf like anthophyllite crystals commonly have undergone retrograde alteration to chlorite and talc. The large, elongate, light grey to blue grey cordierite crystals and aggregates of crystals are commonly retrograded to pinite. The pillow selvages consist of massive biotite and these readily distinguish pillowed from massive flows. Locally, silicious material occurs at pillow borders and in pillow interstices.

STOPS 9 AND 10: ALTERED AND UNALTERED QUARTZ FELDSPAR PORPHYRY "DOME".

As exposed on surface, the Cleaver Lake QFP "Dome" conformably overlies a metasedimentary sequence and underlies the "Ladder" mafic flow. It appears to be up to 1000 m thick west of Cleaver Lake. Unaltered QFP is remarkably homogeneous, massive to foliated, pale pink to greyish pink with 20 to 50% rounded to stretched quartz phenocrysts and 1 to 25 % white to pale pink, lath-shaped 1 to 3 mm feldspar crystals. Alteration has destroyed the feldspar crystals, but the quartz eyes are preserved except in very extreme cordierite-anthophyllite facies alteration. Mineral assemblages in altered QFP include: quartz-muscovite, quartz-biotite-sillimanite, cordierite-quartz-biotite-sillimanite and cordierite quartz-anthophyllite with staurolite, garnet, spinel, magnetite, zircon and rutile as associated constituents. (Morton, 1984).

PHYSICAL VOLCANOLOGY AND HYDROTHERMAL ALTERATION OF THE STURGEON LAKE CALDERA COMPLEX

R.L. Morton[1], G.J. Hudak[1], J.S. Walker[1], and J.M. Franklin[2]

INTRODUCTION

The South Sturgeon Lake area of northwestern Ontario (Fig. 1) is underlain by a well preserved, though partially eroded, Archean submarine volcanic caldera. This caldera is host to six massive sulphide deposits (Mattabi, F-Group, Sturgeon Lake Mine, Creek Zone, Lyon Lake, and Sub-Creek Zone) (Fig. 2, Table I) as well as numerous subeconomic massive sulphide occurrences. The Sub-Creek Zone may represent the down plunge extension of the Lyon Lake Zone (Koopman et al., this volume). The Sturgeon Lake Mine, F-Group and Mattabi deposits were depleted of reserves in 1981, 1984 and 1988 respectively, whereas the Lyon Lake and Creek Zone deposits are currently being mined.

TABLE 1. Grade and tonnage figures of the ore deposits in the South Sturgeon Lake area.

Deposit	Tonnage* (10^6 tons)	Zn %	Cu %	Pb %	Ag oz/ton
Mattabi	12.55	8.28	0.74	0.85	3.31
F-Group	0.38	9.51	0.64	0.58	1.92
Sturgeon Lake	2.28	9.17	2.55	1.21	5.22
Lyon Lake & Creek Zone	3.17	8.67	1.26	0.99	4.50

* Prodution Grade and tonnage figures (from M.R. Patterson)

The combination of well preserved volcanic textures, a variable 55° to 90° dip of a north-facing, essentially homoclinal volcanic sequence and more than 600,000 m of diamond drilling over an apparent 4,500 m stratigraphic interval presents the opportunity to examine successive stages of caldera evolution. This includes initial subaerial-shallow subaqueous collapse, subaqueous silicic explosive volcanism and ore formation, and terminal dome building with sedimentary and lava flow fill and burial.

The main purpose of this field trip is to illustrate the physical volcanology of an ancient, ore-hosting caldera and to stress the importance of volcanism and synvolcanic structures in controlling the occurrence and location of hydrothermal alteration assemblages and volcanogenic massive sulphide deposits.

REGIONAL GEOLOGY

The south Sturgeon Lake area is located within the Archean Wabigoon volcano-sedimentary greenstone

[1] Deparment of Geology Economic Volcanology Research Laboratory, University of Minnesota-Duluth, Duluth, MIN.

[2] Geological Survey of Canada, 601 Booth Street, Ottawa Ontario, K1A 0E8

belt within the Superior Province. The Wabigoon subprovince is bounded to the north by the English River Gneiss Belt and to the south by the Quetico Gneiss Belt (Trowell and Johns, 1986). The volcanic and sedimentary rocks of the area have been subjected to regional greenschist facies metamorphism with almandine-amphibolite assemblages towards both the eastern and southern extents of the Sturgeon Lake area (Trowell, 1974, 1983; Groves, 1984). The South Sturgeon Lake sequence strikes west-northwest, dips steeply (65°-75), faces north, and forms the southern limb of a syncline with an axis centered through Sturgeon Lake (Trowell, 1983).

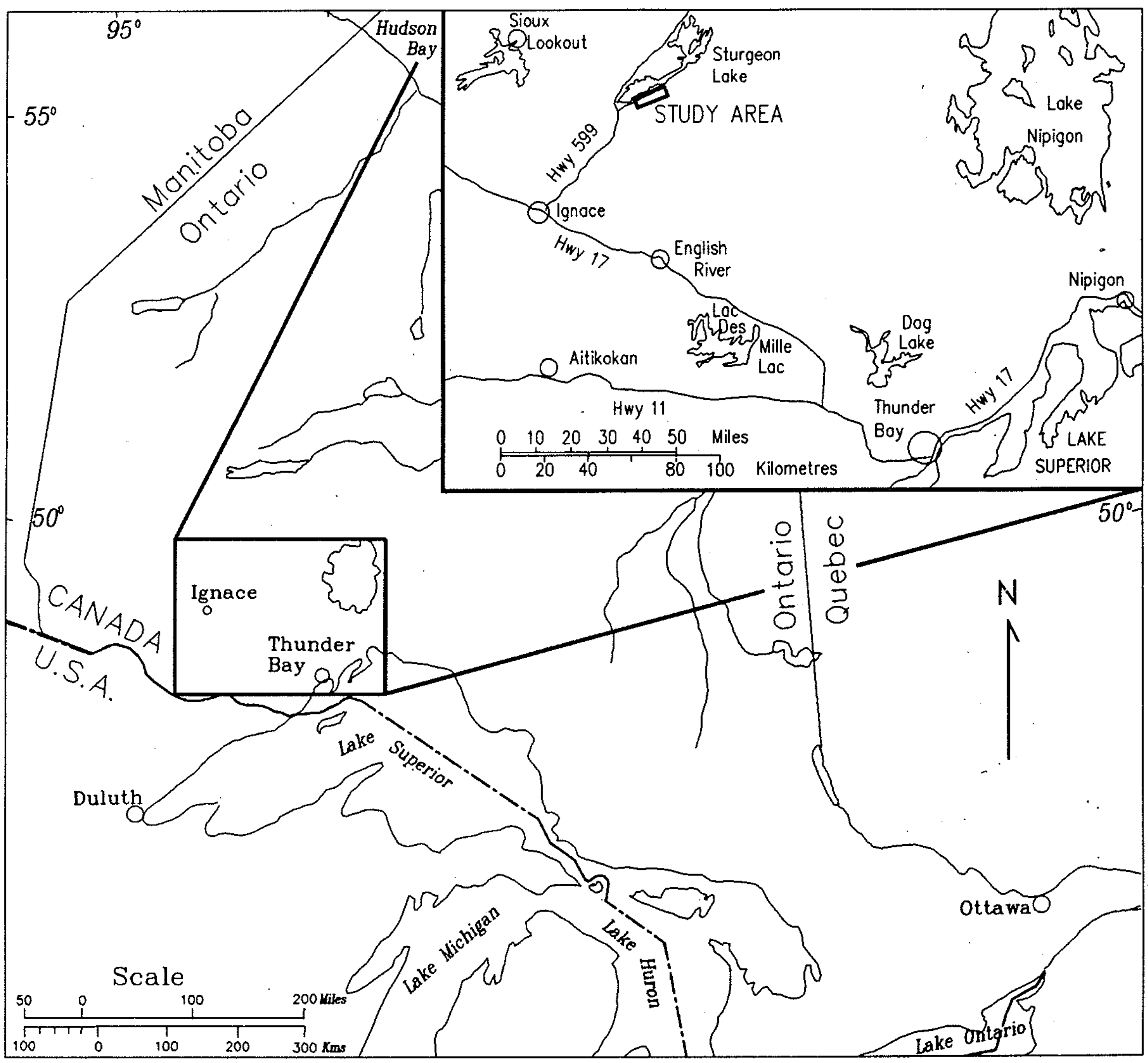


Figure 1 Location of the south Sturgeon Lake area.

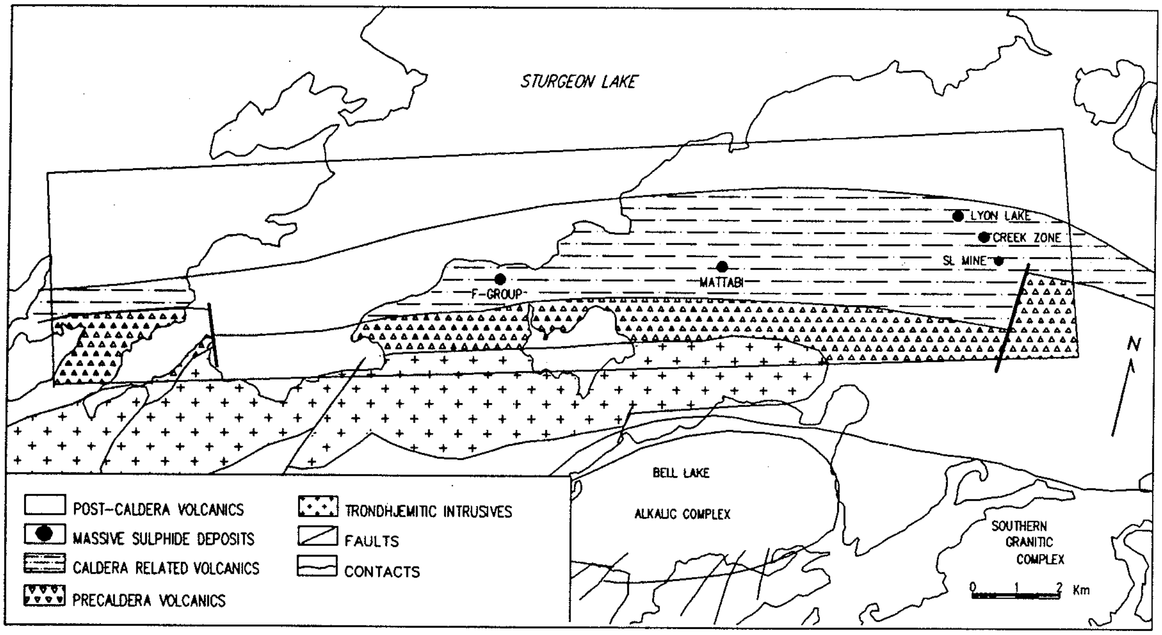


Figure 2 Generalized geological map illustrating the regional extent of the Sturgeon Lake Caldera Complex.

Detailed mapping of the volcanic rocks in the south Sturgeon Lake area coupled with the relogging of 200,000 m of diamond drill core, with emphasis on the physical volcanology of the rocks, has led to the recognition and description of a well preserved Archean submarine caldera complex. This complex, which has been named the Sturgeon Lake Caldera (Morton et al., 1988, 1989, 1990), is approximately 30 km in strike length and contains up to 4500 m of caldera fill material. Five separate, major ash flow tuff units have been defined, and each can be traced for kilometers across the complex with individual thicknesses ranging from 100 to more than 1200 m. Based on the stratigraphic distribution and thickness of the five ash flow tuff units and associated debris flow deposits, it is believed that the Sturgeon Lake Caldera consists of a series of smaller nested or overlapping calderas and that each ash flow unit is associated with a collapse event. This interpretation is supported by studies of more recent caldera complexes which show that major ash flow tuff units are related to individual collapse events and that nested or overlapping calderas are common (Cas and Wright, 1987).

Based on detailed stratigraphic mapping and core logging, numerous synvolcanic faults have been defined; those with major stratigraphic displacement (>150 m) may represent individual caldera boundaries. In general the eastern and western margin of the complex has been located and the back (northern) bounding wall has been partially defined. A series of late, north-south-trending dip-slip faults has broken the complex into a number of blocks which allow the caldera complex to be observed at different stratigraphic levels. Individual volcanic and intrusive rock units are traceable across the faults, but thicknesses change dramatically (Fig. 3). Zircon ages of the ash flow tuff deposits and late dome lavas yield a similar age of 2,735 m.y. ± 1.5 m.y. (Davis et al., 1985).

A large, sill-like intrusion (Beidelman Bay Complex) has a similar age to the felsic volcanics and

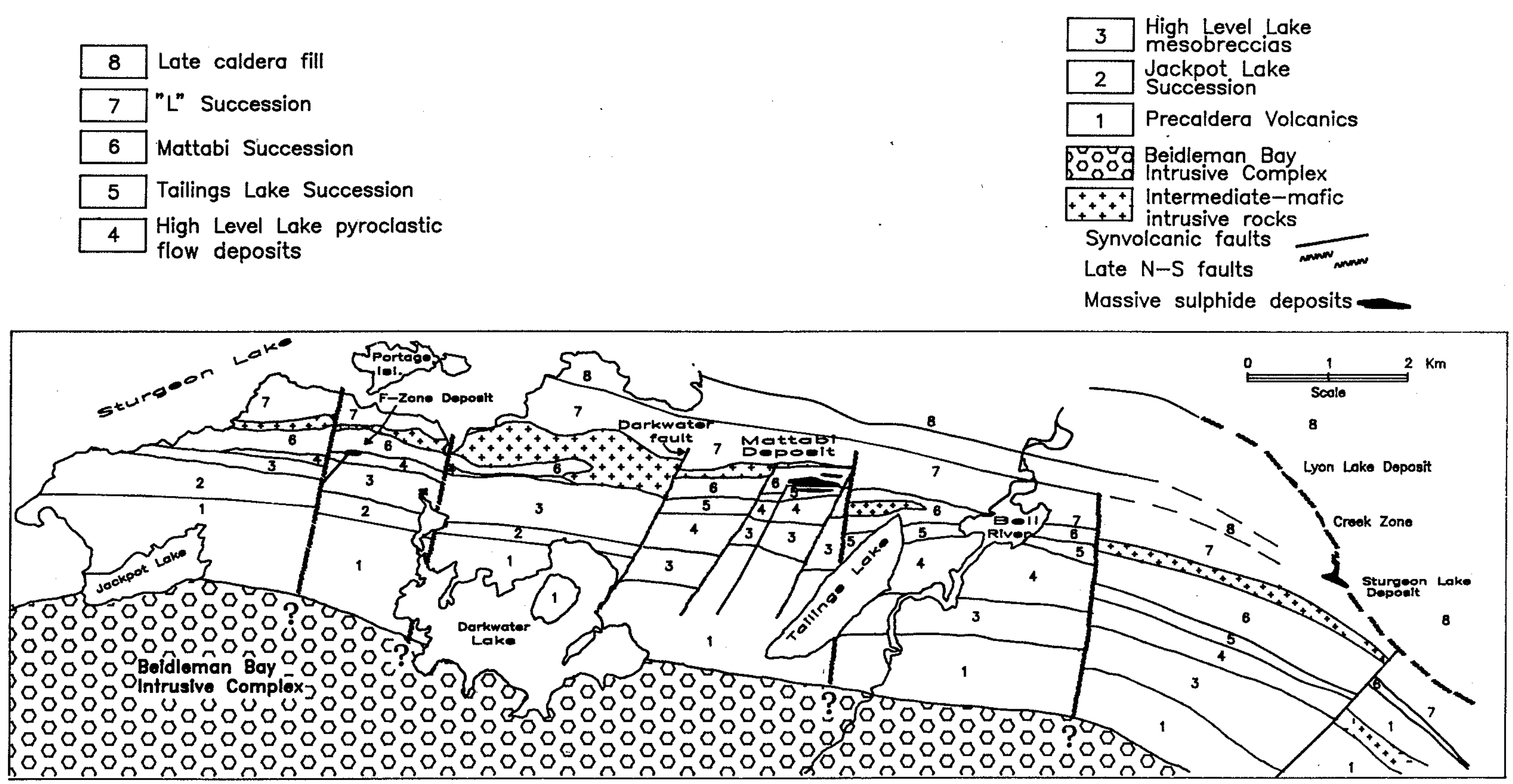


Figure 3 Simplified geological map of the Sturgeon Lake Caldera Complex.

represents, in part, the magma chamber for the eruptive material. This intrusive body can be traced along strike for 20 km and has an average width of 2.5 km; its composition varies from trondhjemite to quartz diorite with feldspar and quartz ± feldspar porphyry phases. The Beidelman Bay intrusive complex also hosts minor occurrences of porphyry copper-type mineralization, and the upper portions locally contain stringers of zinc-silver mineralization. A mafic (gabbroic-quartz dioritic) dyke- to sill-like intrusion occurs across the length of the caldera complex (Fig. 3). Trace element and rare earth analyses (Davis et al., 1985) indicate that the caldera rocks have a similar petrogenetic history and are cogenetic with the subvolcanic Beidelman Bay trondhjemite sill which intrudes the Darkwater mafic flows (Poulsen and Franklin, 1981).

Metamorphosed hydrothermal alteration is widespread within the complex and in the upper part of the Beidelman Bay intrusion. Discrete alteration assemblages form zones that are a) widespread and largely conformable to the volcanic stratigraphy, b) locally lens- or pod-like beneath sulphide occurrences and deposits and c) narrow and elongate, cross-cut stratigraphy and are associated with synvolcanic faults (Groves, 1984; Hudak, 1989; Jongewaard, 1989; Morton et al., 1988; Morton and Franklin, 1987).

As described above, there is only local evidence of significant post-volcanic deformation in the southern part of the South Sturgeon Lake sequence. In the upper part of the caldera sequence and northward (Fig. 2), there is evidence of greater structural complexity.

The South Sturgeon Lake strata have been conventionally determined to comprise a simple north-facing homocline (Franklin, et al., 1975), consisting of several "cycles", each representing a major volcanic episode. Facing indicators, mainly pillows in all parts of the pile, graded beds in the rare sedimentary units, and alteration pipe distribution, are consistently north. Recently, however, several studies have indicated that interpretation as a single homocline may be an over-simplification of the structure.

1) Dusanowskyj and West (1976) interpreted the Bouger gravity map (Barlow et al., 1975) to indicate that the greenstone belt extends downwards only 4 to 5 km, and is underlain by granitic crust.

2) Davis et al. (1985) noted that the ages of Franklin et al's (1978) lower three cycles are within error of being the same (together with the subvolcanic Beidelman Bay sill). Thus the lower 4 to 6 km of the South Sturgeon Lake sequence formed in less than about 2 Ma (the resolution of the age determination method), essentially a single episode, and a remarkably short time for so much volcanic rock.

3) The Lyon Lake and Creek Zone orebodies do not have a clearly defined vertically extensive footwall alteration zone. Alteration usually extends only a few meters below ore.

These data, combined with the structural observations of Dube et al. (1989) and Koopman et al. (this volume) are suggestive that repetition of some parts of the stratigraphic package, and truncation of some units, probably by faulting, has resulted from structural "stacking". Although this hypothesis has not been tested, the apparent motion on the hanging wall fault to the Lyon Lake orebody (Koopman et al., this volume) may indicate that the Lyon Lake Andesite was emplaced in its present position by faulting, possibly an early (Pre-folding?) thrust fault. Other faults of this type are postulated to be present.

There is an apparently consistent eastward plunge to all orebodies in the area, as well as elongated fragments and amygdules. Koopman et al. (this volume) illustrate the important effect that this plunge has on the distribution of ore at Lyon Lake. The consistency of plunge and facing directions precludes major folding as an explanation for the possible structural "thickening". Minor folds are important, however (Dube et al., 1989), making difficult the upward projection of strata intersected in deep drill holes.

A regional foliation cuts all rocks, including the Beidelman Bay Complex. This foliation is predominantly east-west, but displays a slight warping, from ENE in the western part of Sturgeon Lake, through EW in the Mattabi area, to WNW in the eastern extremities of the belt. This slight warping of foliation indicates a late deformation about a NS axis. The dominant deformation, forming the foliation, resulted from N-S transpression, however.

A major fault transects the Beidelman Bay Complex and adjacent strata, just to the west of the edge of figure 3 (Poulsen and Franklin, 1981). This fault strikes northeasterly, and has deformed a late Archean alkaline complex (Sturgeon Narrows), as well as the aforementioned regional foliation. Less prominent parallel late faults occur throughout the area.

PHYSICAL VOLCANOLOGY OF THE STURGEON LAKE CALDERA COMPLEX

PRE-CALDERA VOLCANISM (UNIT 1, FIG. 3 AND 4)

Pre-caldera volcanic rocks, referred to as the Darkwater Succession, have been interpreted to represent part of a large subaerial-subaqueous shield-type volcano composed of a thick sequence of basalt lava flows, scoria-tuff cone deposits and debris flows, with minor amounts of interlayered rhyolitic lava flows and pyroclastic fall deposits (Groves et al., 1988, Morton et al., 1988, 1990). The basaltic lavas beneath and lateral to the caldera complex are composed of amygdaloidal to massive flows which commonly exhibit brecciated and scoriaceous tops; pillow lavas and hyaloclastites have been identified only along the eastern margin of the caldera leading to the interpretation that the mafic lavas are dominantly subaerial (Groves et al., 1988). The scoria and tuff cone deposits represent part of a field of small, monogenic volcanos which formed on the upper flanks of the major shield edifice. One of these has been described in detail (Groves et al., 1988) and all are interpreted as subaerial to shallow subaqueous deposits formed from magmatic and phreatomagmatic eruptions. Caldera collapse caused large blocks and smaller clasts of these lavas and scoria-tuff cone deposits to slide into the caldera where they now form a major component of the various mega-and mesobreccia units (Unit 3).

JACKPOT LAKE SUCCESSION (UNIT 2, FIG. 3 AND 4)

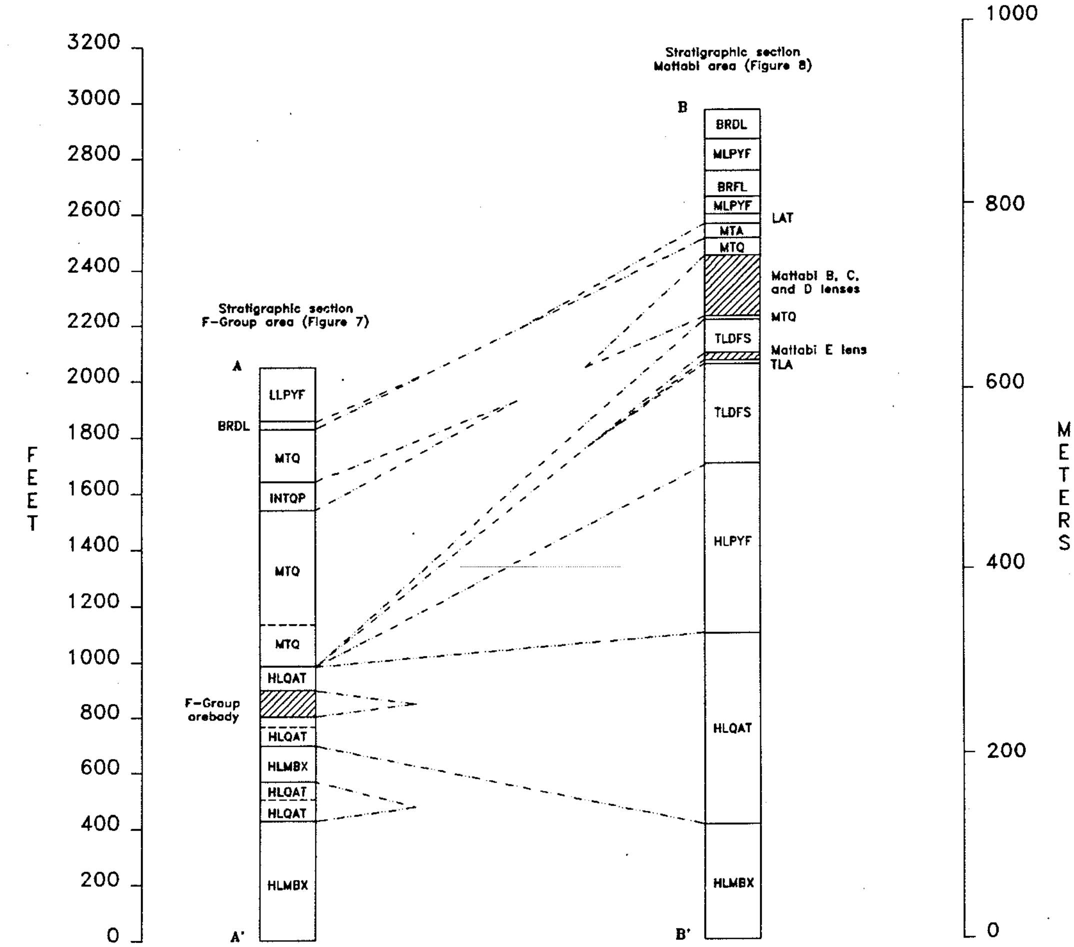
Pyroclastic flow and fall deposits of the Jackpot Lake Succession overlie the pre-caldera volcanic rocks and represent the initial pyroclastic eruptions that led to the first caldera collapse event. These pyroclastic rocks vary from 50-300 m thick and are most abundant in the Jackpot Lake area; they thin eastwards and appear to terminate north of Darkwater Lake (Fig. 3). The extent of these rocks westward, beneath Sturgeon Lake, is not known. These rocks vary from massive to well-bedded and are composed predominantly of fine grained, aphyric ash tuff. Small quartz crystals (<1 mm, 1%) occur locally and are typically angular and splintered or broken; subround-irregular shaped pumice comprises from 1 to 50% of the rocks and is typically recrystallized to small quartz mosaics. The ash-size matrix is composed of interlocking mosaics of quartz and feldspar with a wide variety of secondary alteration minerals. Locally these deposits are separated by thin debris flow units and associated fine grained sedimentary rocks indicating that the eruptions were episodic.

HIGH LEVEL LAKE SUCCESSION (UNITS 3 AND 4, FIG. 3 AND 4)

The High Level Lake Succession has been subdivided into two units: a) mesobreccia and b) quartz crystal-rich and pumice-rich pyroclastic flow deposits. Pre-caldera volcanic rocks and the Jackpot Lake pyroclastic units are immediately overlain by beds of coarse heterolithic breccia that are interlayered with, and grade upwards into, quartz crystal ash flow tuff deposits. The breccia units have a strike length of at least 22 km and exhibit rapid changes in thickness from 80 to more than 900 m.

The breccia deposits (Unit 3) are composed of block-and lapilli-size clasts of mafic and felsic volcanic rocks which are mineralogically and chemically similar to underlying pre-caldera and Jackpot Lake pyroclastic units. Recognizable clasts are poorly sorted and vary from measurable blocks 70 m by 40 m down to matrix-size material; typical size is 3 to 25 cm. Detailed mapping and core logging also indicates that there are large blocks of the Darkwater scoria-tuff cone deposits that are more than 1 km in length and 300 m in thickness. Matrix material to the clasts varies from abundant alteration and metamorphosed alteration minerals (chloritoid, iron carbonate, andalusite, pyrophyllite, sericite, chlorite) to a composition very similar to the interlayered and overlying ash flow tuffs. Overall the composition of this unit is basaltic-andesitic. In the upper 5-50 m the breccias contain sparse quartz crystals and lapilli-size pumice.

These breccia deposits are similar to heterolithic breccias (mega- and mesobreccias) described by Lipman (1976, 1989) from calderas in the San Juan Mountains. In the Sturgeon Lake Caldera Complex these coarse heterolithic breccias are believed to represent debris derived from caldera walls as collapse occurred; collapse and breccia formation were



LEGEND OF GEOLOGICAL UNITS

	Intrusive rocks – Fine grained quartz diorite	(INTQP)
8	No Name Lake Succession	
7	L Succession	
	Upper L – Quartz and plagioclase crystal ash flow tuffs	(ULPYF)
	– Volcaniclasitc sedimentary rocks	(LSED)
	Middle L – Dacitic lava flows	(BRDL, BRFL)
	– Quartz crystal ash flow tuff	(MLPYF)
	Lower L – Andesitic to dacitic lava flows	(BRDL)
	– Quartz and plagioclase crystal ash flow tuffs	(LLPYF, LAT)
6	Mattabi Succession	
	Massive to bedded ash flow tuff	(MTA)
	Bedded quartz crystal ash flow tuff	(MTQ)
5	Tailings Lake Succession	
	Heterolithic debris flow deposits	(TLDFS)
	Felsic ash tuff	(TLA)
4	High Level Lake Succession	
	Pumice-rich ash flow tuff	(HLPYF)
	Quartz crystal ash flow tuff	(HLQAT)
3	Heterolithic breccia (Mesobreccia)	(HLMBX)
2	Jackpot Lake Succession	
1	Precaldera rocks	

SYMBOLS

- Geological Contact (dashed where inferred)
- Massive Sulphide Deposit
- Synvolcanic Fault
- Correlation Line

Figure 4 Stratigraphic correlations between the F-Group and Mattabi areas.

simultaneous with eruption and deposition of the High Level Lake quartz crystal ash flow tuff. Thick deposits of mesobreccia have been related to catastrophic caldera subsidence of a kilometer or more (Lipman, 1976, 1989). Such catastrophic collapse in the Sturgeon Lake area could readily have moved the floor of the Sturgeon Lake Caldera from an initial subaerial-shallow subaqueous environment to a deeper submarine environment.

The High Level Lake ash flow tuff deposits (Unit 4) represent the second pyroclastic event associated with the caldera complex and can be subdivided into two separate units: a) a lower quartz crystal-rich rhyolite (average SiO_2 of 74%) and b) an upper pumice-rich dacite (average SiO_2 of 67%). The quartz crystal ash flow tuff forms lenses (up to 1000 m in strike length by 100 m thick) within deposits of mesobreccia and an 80-300 m thick unit which overlies the mesobreccia. The contact between the mesobreccia and the quartz crystal-rich rhyolite is gradational and is marked by a gradual upward increase in the size and percentage of quartz crystals and a marked decrease in the number of lithic clasts (< 5%). The ash flow deposits are composed of 3-25% broken and angular quartz crystals (0.8-2 mm), which locally are enclosed by 0.1 to 0.8 mm thick ash rims, and 5-30% silicified and/or carbonated lapilli-size pumice. The ash rims surrounding the quartz crystals suggest the presence of water in the eruptive environment.

These quartz crystal-rich ash flow tuffs grade upwards into quartz-crystal poor, pumice-rich ash flows that vary from 50 to 370 m in thickness. These massive deposits are composed of 35-75% silicified and/or chloritoid-andalusite-pyrophyllite-rich lapilli-size pumice and 5-25% mafic and chlorite-carbonate-rich lithic fragments set in a fine-grained quartz-rich matrix. The upward change in composition from rhyolite to dacite, decrease in quartz crystals, massive nature of the deposits and the presence of abundant pumice can be interpreted to represent rapid eruption of magma from a zoned or layered, gas-rich magma chamber.

The quartz crystal unit can be traced for 20 km across the caldera, whereas the pumice-rich unit is not observed west of the major northeast-trending synvolcanic fault (Darkwater fault). This fault has an apparent offset of 400 m and forms one side of a major topographic depression that hosts the ore lenses of the Mattabi massive sulphide deposit. The pumice flow ponded within but did not fill this depression.

TAILINGS LAKE SUCCESSION (UNIT 5, FIG. 3 AND 4)

The High Level Lake pumice flows represent the end of the second explosive eruptive cycle and are overlain by 60-400 m of subaqueous debris flow deposits and bedded epiclastic rocks. Like the High Level Lake pumice flows, rocks of the Tailings Lake Succession are not found west of the Darkwater fault; however, they can be traced eastward from this structure for more than 10 km. The debris flow deposits are composed of thick (5-120 m) massive basal units overlain by well-bedded sedimentary material which may exhibit grading. The debris flows are composed of clasts derived from the underlying ash flow tuffs and from the pre-caldera rocks. The size and percentage of clasts increases westward toward the Darkwater fault and, in deep drill holes, downdip toward the back margin of the caldera. Locally the debris flow deposits and associated sedimentary rocks are separated by dacitic ash tuffs and by dacitic to andesitic lava flows; these units only rarely crop out and are primarily observed in drill core. The ash tuffs are aphyric and fine-grained, have SiO_2 contents that range from 65 to 77% and TiO_2 contents that are consistently greater than 0.6%; they have locally undergone intense hydrothermal alteration. These rocks range from 5 to 80 m in thickness and are the host rocks to the Mattabi E ore lens. The lava flows range in thickness from 10 to 50 m and vary from massive to amygdaloidal and brecciated; locally these rocks are feldspar-phyric.

The debris flow deposits represent the periodic avalanching of material from caldera walls, whereas the ash tuffs indicate sporadic eruption of felsic material, most likely along leaky ring(?) fractures. The uppermost debris flows are composed of more than 60% pre-caldera mafic volcanic and High Level Lake ash flow tuff clasts, and these are interlayered with abundant ash deposits. These units appear to represent renewed activity at caldera margins which immediately preceded the eruption of the overlying Mattabi ash flow tuffs.

MATTABI SUCCESSION (UNIT 6, FIGS. 3 AND 4)

The Mattabi Succession represents the third and most voluminous eruptive event within the caldera, forming deposits that exceed 1200 m in thickness. The deposits can be traced across the caldera complex for more than 20 km and directly overlie debris flow or ash deposits of the Tailings Lake Succession east of the synvolcanic Darkwater fault; they overlie the High Level

Lake quartz-rich ash flow tuff and mesobreccia deposits west of this structure. East of the Darkwater fault the Mattabi ash flow tuffs range from 150-1200 m thick; they thin rapidly west of this structure (with the notable exception of the F-Zone area), ranging in thickness from 20-80 m.

It is believed that the Darkwater fault represents one segment of the caldera wall for the Mattabi eruptive event and that deposits west of this structure represent outflow sheets. In the vicinity of the F-Zone ore deposit the Mattabi ash flow tuffs are 125-200 m thick and are believed to have ponded in, and filled, the fault-bounded basin which hosts this massive sulphide deposit (Hudak, 1989).

Based on flow morphology and composition, the Mattabi ash flow tuff succession may be divided into two distinct units: a) a bedded quartz crystal-rich unit and b) a massive to poorly-bedded ash unit. The bedded quartz-crystal rich unit varies from 150 to 1100 m thick east of the Darkwater fault and 20-75 m thick west of the fault. The bedded ash flow tuffs have an average Sio2 content of 75%, always contain less than 0.5% Na_2O and exhibit a pronounced cyclicity within individual bedded sections. Beds are subdivided into basal quartz crystal- (5-35%) and pumice-rich (5-45%) lower sections overlain by bedded ash deposits which contain sparse quartz crystals (1-3%) and rare pumice. The lapilli-size pumice are well-vesiculated (40-70% amygdules) and, with subangular to irregular shapes, exhibit no evidence of flattening that might be indicative of welding. The lower units, in core from relatively deep diamond drill holes that approach the back wall and eastern margin of the caldera, contain up to 20% pre-caldera mafic and felsic volcanic clasts and possibly indicate mixing, close to caldera margins, with debris derived from caldera walls.

Detailed core logging in the eastern part of the area (below the Sturgeon Lake Mine) has defined 14 such flow units which vary in thickness from 15 to 180 m; the bedded ash component is 1 to 14 m thick. More than 280 chemical analyses have been completed on these units with one sample collected for about every three meters of stratigraphy. Analyses of standard trace elements clearly and dramatically show that individual flows (bed sets) have high Zr (550-1025 ppm), Y (75-120 ppm) and Nb (30-80 ppm) at and/or near their base, and that these elements gradually decrease upwards towards the ash beds which have low Zr (80-230 ppm), Y (24-45 ppm) and Nb (1-4 ppm) contents; TiO_2 and Ba appear to have opposite trends. This zoning pattern holds true regardless of the size or percentage of quartz crystals found in the basal beds or the thickness of the beds. At Mattabi the flow units exhibit a similar thickness, composition and chemical zonation with the B, C and D massive sulphide ore lenses situated at breaks between quartz-rich beds in the hanging wall and ash beds in the footwall. Numerous other massive sulphide occurrences are found throughout the lateral and vertical extent of this unit, making it a prime exploration target.

The bedded quartz-crystal deposits are overlain by massive, very fine-grained ash flow tuff deposits which vary from 20 to 150 m in thickness. These units are composed of zero to 2% small (< 1 mm), broken and sliver-like quartz crystals in a quartz-rich matrix; it is common for these units to be intensely silicified and iron metasomatized. This massive unit is overlain by 5-35 m of bedded, fine-grained ash material which is compositionally and chemically similar to the underlying massive ash flow tuffs. The massive unit contains no known massive sulphide mineralization and lacks the pronounced geochemical zoning exhibited by the bedded deposits. In morphology these units are similar to those described by Busby-Spera (1984) from the Mineral King roof pendant and by Morton and Nebel (1983) from the Wawa area, and represent flow units deposited in a relatively shallow subaqueous environment.

L SUCCESSION

This is a complex succession of rocks which range from quartz- and quartz-plagioclase-bearing pyroclastic flow deposits through plagioclase-phyric lava flows and domes to sedimentary sequences. The entire succession ranges from 250 to 1200 m thick with the domes, lava flows and dome related sediments and debris flow deposits making up the bulk of the stratigraphy. The pyroclastic flow deposits form two distinct units, each of which may represent a major pyroclastic eruption (caldera forming ?). The distribution of these pyroclastic rocks and their physical volcanology and alteration is currently the subject of a doctoral thesis. Work completed so far shows that the ash flows associated with the first eruptive event were quartz- and plagioclase-crystal rich whereas the ash flows associated with the second eruptive period were dominantly quartz crystal-rich. These quartz-rich pyroclastic flows are the host rocks for the Sturgeon

Lake Mine, Lyon Lake and Creek Zone massive sulphide deposits. Feldspar-bearing lava flows and domes (up to 400 m thick and 3 km long), along with a variety of dome-derived volcaniclastic rocks and debris flow deposits, overlie the pyroclastic flow units. The domes occur close to caldera margins and are believed to represent the last felsic eruptive products associated with caldera formation. Locally iron formation (banded magnetite-chert and iron carbonate) and graphitic sedimentary rocks are interlayered with dome-derived volcaniclastic rocks and represent more quiet sedimentation and low temperature hydrothermal activity in basins that formed between the growing domes and caldera walls.

The L succession of rocks is overlain by extensive basaltic-andesitic lava flows which have been termed the No Name Lake Andesite (at Mattabi) and possibly the Lyon Lake Andesite (at Lyon Lake). Previous workers believed that the No Name Lake Andesite separated the Mattabi and L pyroclastic rocks from what had been termed the NBU rhyolites. It was believed that the NBU rhyolites were the host rocks for the Lyon Lake and Creek Zone ore bodies, and that the Lyon Lake andesite stratigraphically overlaid these ore deposits (Mumin, 1988, Franklin et al., 1977). Detailed volcanological facies mapping and core relogging, along with geochemical studies, now shows that No Name Lake and Lyon Lake Andesites are compositionally similar, and possibly correlative. Structural complexities separating these units have not been resolved. The No Name Lake and Lyon Lake basalts and andesites form amygdaloidal to massive flows, pillow lavas, hyaloclastites and scoriaceous rocks and represent the end of caldera pyroclastic activity and the start of a new, more quiescent cycle of volcanism.

SUMMARY OF THE VOLCANOLOGICAL HISTORY OF THE STURGEON LAKE CALDERA COMPLEX

1) Subaerial basaltic shield volcanism.

2) Subaerial-shallow subaqueous scoria and tuff cone formation as an end stage in the development of the shield volcano.

3) Formation of the ancestral Beidelman Bay silicic, high level magma chamber and initial eruption of the Jackpot Lake pyroclastic rocks along incipient ring(?) fractures.

4) Major eruption of the Jackpot Lake pyroclastic flow and fall deposits, triggering instability and initiating the first caldera collapse event.

5) Eruption of quartz crystal-rich ash flow tuffs along major ring(?) faults, coupled with the avalanching and sliding of pre-caldera material from growing caldera walls, to form the mega- and mesobreccia deposits and the mixed breccia and pyroclastic units.

6) Continued eruption of pyroclastic material from a zoned magma chamber with a gradual change from rhyolitic quartz crystal-rich to dacitic pumice-rich material.

7) High temperature hydrothermal activity and the formation of the F-Zone massive sulphide deposit with the High Level Lake quartz crystal units as the host rocks.

8) Periodic movement along old and newly developed ring(?) fractures to form the debris flows and associated bedded epiclastic rocks of the Tailings Lake Succession. This was coupled with the periodic eruption of ash tuffs from the developing ring(?) fracture system.

9) Formation of the E ore lens at Mattabi.

10) Major movement on newly developed ring(?) structures coupled with the eruption of fine ash to form the upper part of the Tailings Lake Succession.

11) Pulsating subaqueous eruptions to form the bedded Mattabi ash flow tuffs coupled with continuous intense hydrothermal activity and formation of the B, C and D ore lenses at Mattabi. The hydrothermal system was active both during and between eruptive pulses.

12) Continuous, shallow subaqueous eruption of the massive Mattabi ash flow tuff and continued lower temperature hydrothermal alteration.

13) Generation of andesitic lava flows and debris flows that separate the Mattabi ash flow tuffs from the first pyroclastic event of the L Succession.

14) Eruption of the lower L pyroclastic units, which are quartz- and plagioclase-phyric and mark the

first pyroclastic activity where plagioclase crystals are present. The formation of the subeconomic A zone at Mattabi was associated with this pyroclastic event.

15) Deposition of plagioclase-phyric lava flows and volcaniclastic sediments derived from the L and Mattabi pyroclastic flows.

16) Eruption of quartz crystal-rich pyroclastic material (second L pyroclastic event) and renewed high temperature hydrothermal activity leading to the formation of the Sturgeon Lake Mine, Lyon Lake and Creek Zone ore deposits.

17) Lava dome building and periodic dome collapse with reworking and erosion of the domes occurred. Associated with dome formation were iron- and graphite-rich moat sediments that were related to the last stages of hydrothermal activity in the caldera. Again, structural complexities have not been resolved, and these sedimentary strata may have been deposited prior eruption of the quartz-crystal-rich pyroclastic material and the hydrothermal event which formed the ore bodies.

18) Final caldera fill by mafic lavas that represent the start of a new cycle of more quiescent volcanism.

HYDROTHERMAL ALTERATION

Hydrothermal solutions, both ore-forming and non-ore-forming, have altered the volcanic rocks associated with the Sturgeon Lake Caldera Complex. Following alteration, these rocks were metamorphosed to greenschist facies. Therefore, the alteration minerals now present are in metamorphic rocks derived from the alteration minerals formed in the hydrothermal system.

Hydrothermal fluids formed five distinct, mappable alteration assemblages in the area. From least- to most-altered, these assemblages are: a) widespread, semiconformable carbonatization and silicification; b) widespread, semiconformable iron carbonate ± iron chlorite; c) widespread, semiconformable chloritoid ± iron carbonate and/or iron chlorite; d) lens- to pod-like, locally linear zones of aluminum silicate (pyrophyllite, andalusite, and/or kyanite) + chloritoid; and e) linear cross-cutting, and stratiform zones of aluminum silicate. Late sericite and/or magnesium-rich chlorite alteration locally overprints these five alteration assemblages.

Iron carbonate ± iron chlorite assemblage rocks contain at least 10% iron carbonate + iron chlorite with less than 5% chloritoid or aluminum silicate minerals. Outcrops which contain abundant iron carbonate are easily identified by their orange to orange-brown stained, commonly pitted surfaces. Staining varies from irregular, lens-shaped patches up to 15 cm in diameter, to veins and veinlets 1-15 mm in width that are aligned parallel to the foliation. Pumice that has been replaced by carbonate can be recognized as rounded to oval, orange-brown stained pits up to 5 cm in diameter which commonly contain rounded to lens-shaped quartz-filled amygdules (10-60%). In thin section, this assemblage contains iron carbonate ± iron chlorite (10-60%), sericite (up to 30%), Mg-rich chlorite (0-50%), and locally, traces of chloritoid or aluminum silicate minerals.

Chloritoid ± iron carbonate and/or iron chlorite assemblage rocks contain greater than 5% chloritoid, and are characterized by the presence of 1-3 mm dark green to black chloritoid prisms and rosettes. The presence of the chloritoid commonly gives the rocks a "salt and pepper" appearance. Locally, chloritoid porphyroblasts occur with sericite (10-55%) in 1-5 mm wide grey-green veinlets which can be conformable to or cross-cut the strata. Other minerals present include iron carbonate (up to 60%), iron chlorite (2-20%), magnesium chlorite (1-20%) and pyrite (0-12%).

Aluminum silicate + chloritoid bearing rocks typically contain 1-3 mm chloritoid porphyroblasts (up to 33%) in a grey to grey-pink matrix composed of massive pyrophyllite (5-20%), 1-3 mm blocky pink andalusite (0-8%), and/or bluish tabular porphyroblasts of kyanite (0-8%) up to 2 mm long. Chloritoid occurs as 1-3 mm prismatic crystals or rosettes disseminated throughout the rock, or in chloritoid-andalusite-rich veinlets up to 1 cm in width. In addition, iron chlorite (up to 7%), magnesium chlorite (up to 35%), iron carbonate (3-35%), pyrite (trace - 30%) and sphalerite (up to 5%) are also associated with this alteration assemblage.

Rocks containing greater than 5% aluminum silicate minerals (pyrophyllite, andalusite, and/or kyanite) or micaceous pseudomorphs of aluminum silicates with less than 5% chloritoid represent the aluminum silicate assemblage. Generally, rocks bearing this alteration assemblage are light grey to pink in colour. Andalusite is the most commonly observed aluminum silicate mineral, and is present as 1-3 mm grey to pink, blocky to rounded porphyroblasts in micaceous domains. Kyanite occurs in three forms: a) as ragged grains in the matrix of strongly altered,andalusite-rich rocks; b) as subhedral to euhedral grains associated with pyrite, sphalerite, and/or chalcopyrite mineralization; and c) as euhedral, commonly bent blades up to 2 cm in length within or along the margins of quartz ± carbonate veins and veinlets up to 3 cm in width. Other phases present within this assemblage include quartz (10-80%), sericite (0-40%), iron carbonate (0-10%), chloritoid (0-4%), magnesium chlorite (0-30%), pyrite (trace - 15%), and sphalerite (locally up to 7%).

Figure 5 illustrates the distribution of the ore-forming alteration assemblages in the F-Group area. Iron carbonate ± iron chlorite assemblage rocks form a large, semiconformable zone along the southern, eastern, and northern boundaries of the area. Locally, pod-like areas (up to 50 m in diameter) of this assemblage occur within the chloritoid assemblage zone, which exists as a large, semiconformable zone up to 100 meters thick in the hanging wall to the F-Group orebody, and is present as pipe-like alteration zones in the F-Group footwall. More proximal to the mineralization, semiconformable and locally pipe-like zones containing the aluminum-silicate ± chloritoid alteration assemblage are present. Aluminum silicate assemblage rocks are most closely associated with the mineralization at the F-Group deposit. These rocks are distributed in two distinct patterns: a) in linear, 5-30 meter wide, pipe-like zones that trend NNE and cross-cut the stratigraphy; and b) in a broad, semiconformable zone (700 by 500 meters at surface) located both in the footwall and the hanging wall rocks to the F-Group massive sulphide deposit.

Figure 6 illustrates the distribution of ore-forming alteration at the Mattabi deposit. Semiconformable carbonatized and silicified rocks form a broad zone in the lower footwall rocks (Franklin et al., 1975). This zone is cross-cut by several westward plunging tabular zones outwardly zoned with an aluminum silicate-rich core and aluminum silicate + chloritoid margin. These zones commonly surround synvolcanic fault zones which lead upward to and cross-cut a broad semiconformable zone of aluminum silicate + chloritoid altered rocks, which crudely surrounds the Mattabi deposit. The aluminum silicate-rich tabular zones, associated with the synvolcanic faults, spread out into a semiconformable zone directly beneath and lateral to the deposit. Stratigraphically overlying the deposit is a semiconformable zone of chloritoid ± Fe-carbonate ± Fe-chlorite up to 150 m thick.

FIELD TRIP STOP DESCRIPTIONS

F-GROUP AREA

The F-Group volcanogenic massive sulphide deposit was discovered from airborne and ground geophysical surveys combined with exploration diamond drilling conducted by Mattagami Lake Mines in 1969. The orebody was mined from 1981-1984 and yielded 377,565 tons of ore which contained 0.64% Cu, 9.51% Zn, 0.58% Pb, and 1.92 oz/t Ag (M. Patterson, personal communication, 1990).

STOP F-1: HIGH LEVEL LAKE ASH FLOW TUFF AND MESOBRECCIA

At this location (Fig. 7), approximately 300 meters southwest of the F-Group pit, intensely altered High Level Lake ash flow tuff and mesobreccia deposits crop out. The High Level Lake ash flow tuff deposits are the host rocks for the F-Group massive sulphide orebody. These rocks vary in colour from grey-green to greyish-pink and contain 5-20% 1 mm quartz phenocrysts in a quartz-rich, locally chlorite-bearing recrystallized ash matrix. Light grey pumice fragments, from 3-20 mm in diameter, are subrounded to oval in shape, and are composed primarily of recrystallized quartz. These fragments typically contain 30-50% <1 mm round to oval quartz-filled amygdules.

High Level Lake ash flow tuff deposits overlie and are locally interbedded with mesobreccia deposits. Mesobreccias formed from material that slumped off oversteepened walls of a caldera during and after caldera collapse. By definition, these deposits contain fragments dominantly less than 1 m in diameter (Lipman, 1976). Megabreccia deposits are formed more proximal to the caldera wall, and contain fragments generally greater than 1 m in diameter. Mesobreccia deposits which occur in the footwall to the Mattabi orebody (see Stop M-1a) and the Sturgeon Lake Mine are laterally equivalent to the mesobreccia deposits on this outcrop.

At this location, the mesobreccia deposits vary from green to grey-green and contain up to 50%

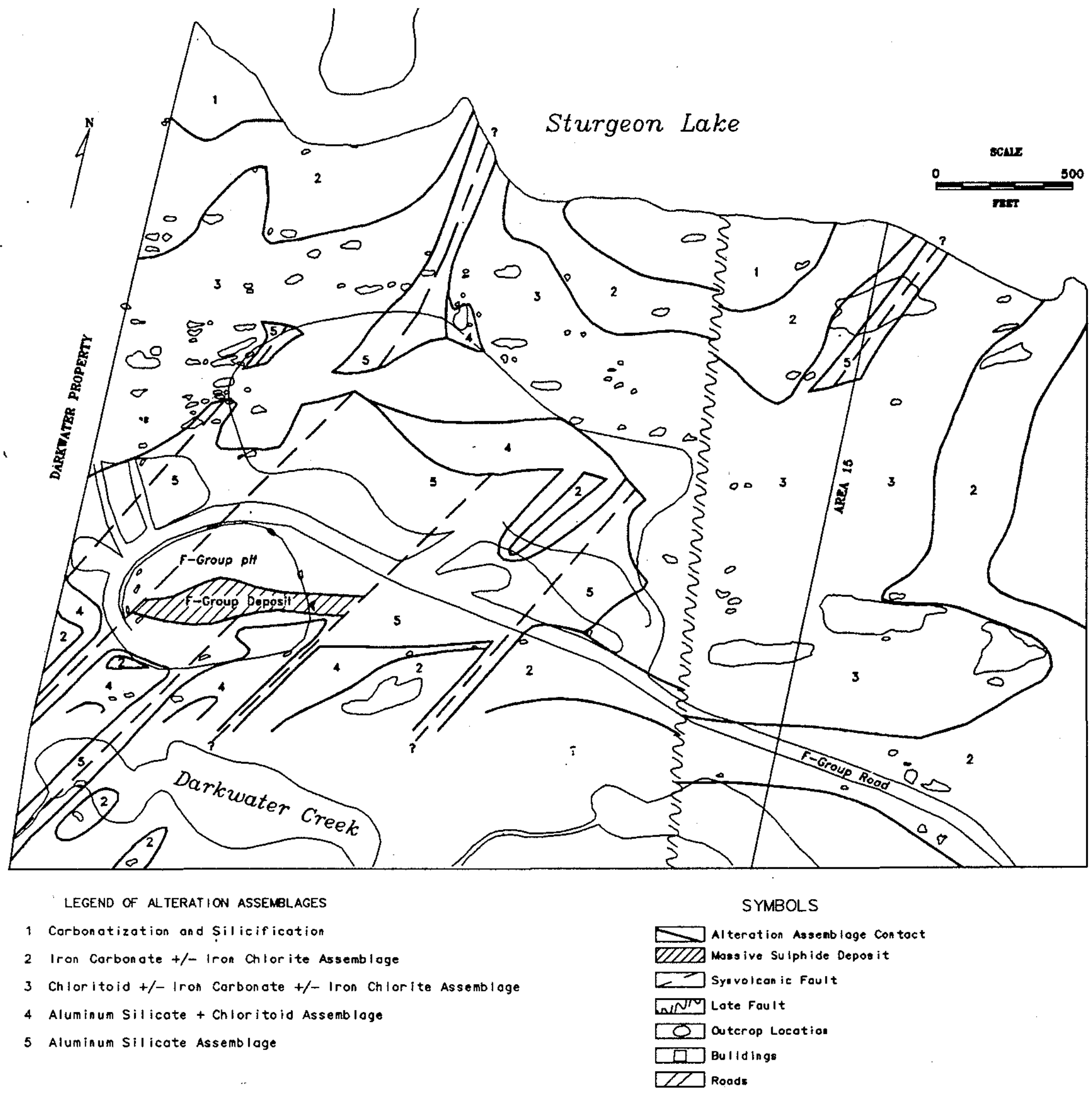


Figure 5 Distribution of ore-forming alteration assemblages associated with the F-Group area.

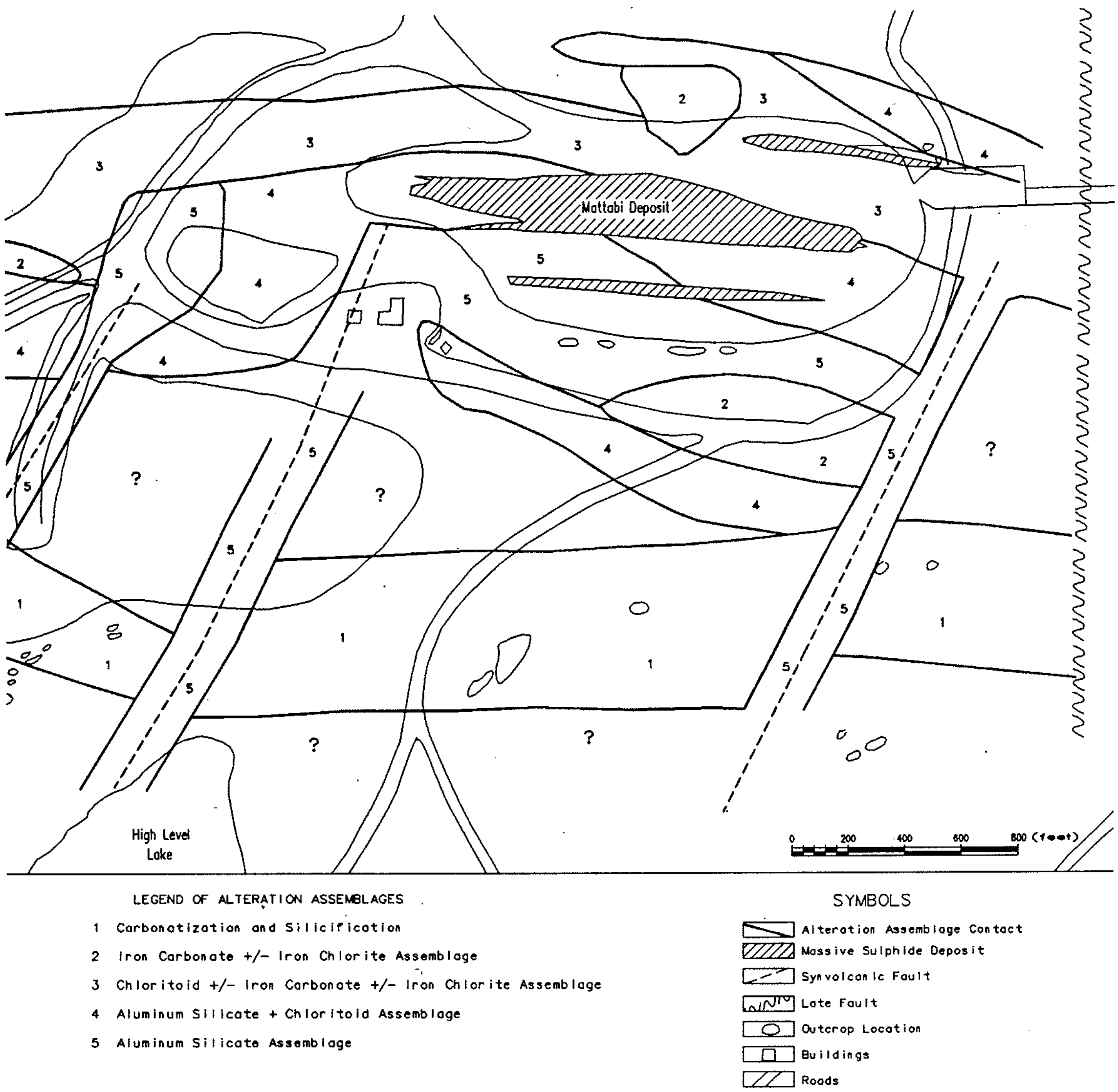


Figure 6 Distribution of ore-forming alteration assemblages associated with the Mattabi area.

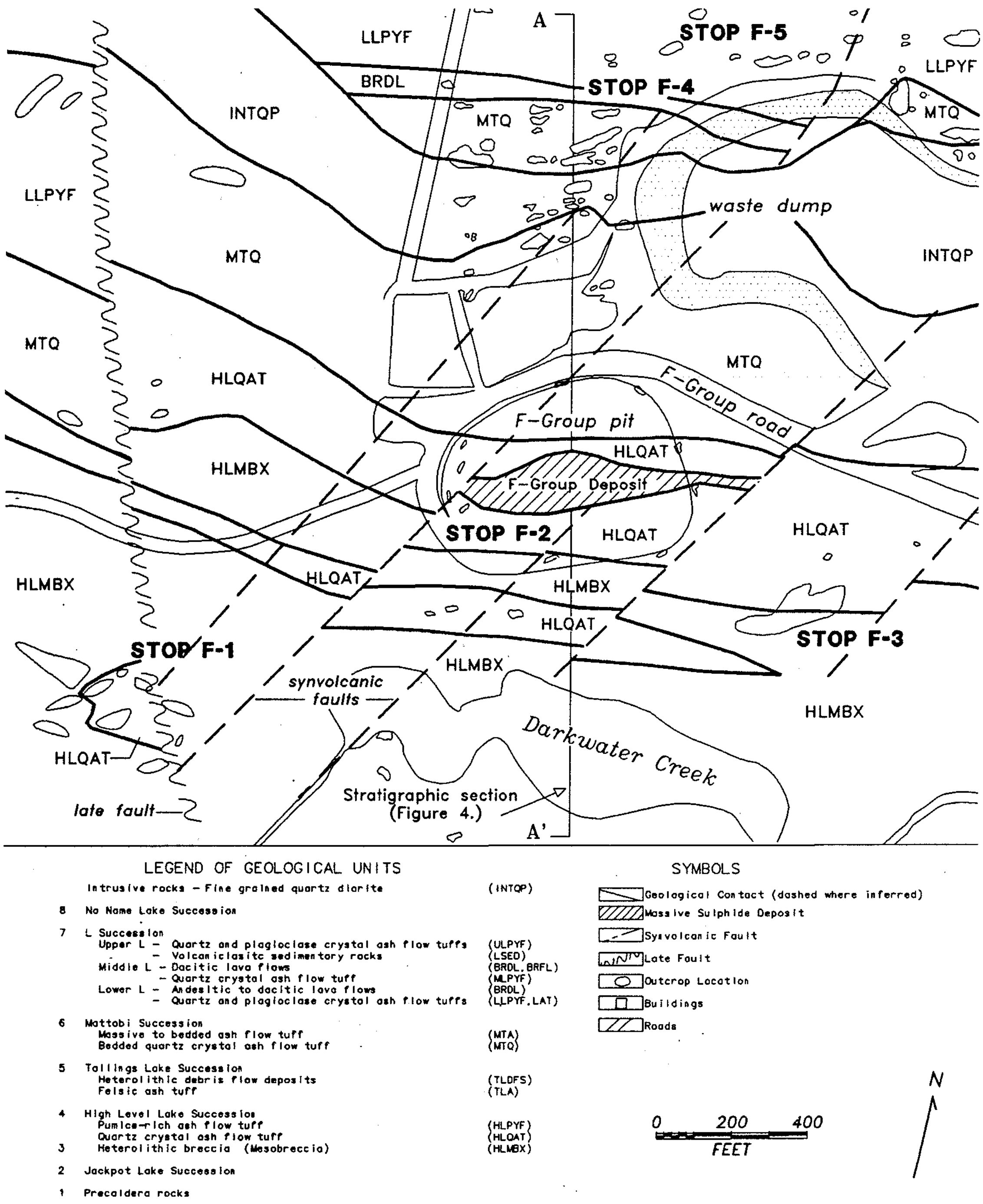


Figure 7 Geological plan map of the F-Group area, with field trip stop locations.

fragments composed principally of three types: a) 5% rounded, 3-10 mm in diameter, amygdaloidal quartz- and sericite-rich fragments interpreted to be pumice;b) 5-10% light grey lapilli-sized subrounded felsic lava fragments; and c) up to 30% rounded to oval, 3-10 mm diameter chlorite-rich amygdaloidal fragments believed to be scoria and amygdaloidal basalt. The matrix to this unit is composed primarily of magnesian chlorite, quartz and sericite.

Both lithological units present in this outcrop have undergone pervasive aluminum silicate alteration (ie., Na, Ca, and Mg loss). Three different aluminum silicate minerals are present. Andalusite (5-35%) occurs as 1-6 mm equant pink porphyroblasts. In thin section, andalusite is present as ragged, inclusion-rich porphyroblasts with sericite/pyrophyllite rims, and as fresh porphyroblasts that occur in patches and veins up to 5 mm wide. Kyanite is present in two forms: a) as ragged tabular porphyroblasts up to 5 mm in length within the altered matrix of the tuffs; and b) as pale blue blades (3-20 mm in length) within white to red-brown quartz-iron carbonate veins and veinlets that are up to several cm in width. Pyrophyllite is commonly found along the margins of both the andalusite and kyanite porphyroblasts, and can also occur in veins up to several cm in width as soft, pale green radiating micaceous aggregates. Pyrophyllite is often present where quartz veins intersect kyanite veins. Minor amounts of chloritoid (generally <5%, but locally up to 10%) are associated with these aluminum silicate minerals.

This series of outcrops is believed to be proximal to and part of a synvolcanic fault. These faults are believed to be the channnelways in which high temperature, metalliferous hydrothermal solutions traveled to the seafloor. The aluminum silicate alteration is believed to be produced when these high temperature fluids leached cations from the rocks (for example, during the alteration of feldspar or volcanic glass), leaving them rich in aluminum and silica.

STOP F-2: F-GROUP TRENCH

The F-Group trench was unearthed in the summer of 1987 to channel runoff waters from nearby waste dumps into the F-Group pit. This trench exposes intensely altered High Level ash flow tuff deposits, which are the host rocks to the F-Group orebody. These rocks contain a well-foliated, locally quartz-phyric, pink groundmass containing andalusite (up to 30%), pyrophyllite and sericite. Green, prismatic porphyroblasts of chloritoid, ranging from 1-4 mm in length, make up 5-25% of the unit, and give the rock its distinctive "salt and pepper" appearance. This assemblage is commonly associated with massive sulphide mineralization at Mattabi.

STOP F-3: MESOBRECCIA - HIGH LEVEL LAKE PYROCLASTIC FLOW CONTACT

The southern section of stop F-3 (Fig. 7), which is located approximately 75 meters southeast of the F-Group pit, is composed in part of heterolithic debris flow deposits that contain a green chlorite-and sericite-rich matrix. These deposits comprise part of the major mesobreccia unit on the regional geology map (Fig. 3). Fragments make up 30-70% of this outcrop and are of two types: a) easily recognizable, angular to rounded light grey to pale white felsic lava flow and ash fragments; and b) green, difficult to recognize, lapilli-sized rounded to angular scoria and amygdaloidal basalt fragments. Both types of fragments are believed to have been derived from the underlying pre-caldera Darkwater Lake Succession rocks.

To the northeast, the mesobreccia deposits become less fragmental, and 1 mm quartz phenocrysts can be observed in the matrix over a zone which varies from 1-5 m in width. This zone represents the gradational contact between the mesobreccia deposits and the overlying quartz-phyric pyroclastic flow deposits. This gradational contact is believed to have formed from mixing of these two units. This contact zone, combined with the interlayering of the High Level Lake ash flow tuff and mesobreccia deposits, suggests simultaneous deposition of these two units. Similar relationships have been documented in mesobreccias that occur in the San Juan Mountains of Colorado (Lipman, 1976).

The light grey quartz- and sericite-rich, recrystallized ash matrix of the High Level Lake ash flow tuff units contains 5-25% 1 mm quartz phenocrysts. Irregular patches and lenses of red-brown iron carbonate alteration vary from 1-10 cm in length and locally comprise 10-15% of the outcrop. In thin section, many of the 1-2 cm carbonate patches appear to be altered pumice.

STOP F-4: MATTABI ASH FLOW TUFF DEPOSITS

At the northern edge of the lime pond, two outcrops composed of Mattabi ash flow tuff are exposed (Fig. 7). Locally, lenses containing up to 60% pyrite are present. These lenses vary from 5-10 m in length, and may reach a thickness of 1 m. To the north, silicified Mattabi pyroclastic deposits are in contact with a dark green, iron carbonate-rich, amygdaloidal andesite intrusion. This intrusion dilates the Mattabi pyroclastic deposits in the F-Group area.

To the north of the intrusion, Mattabi ash flow tuff deposits again crop out. These deposits are grey to pale green in colour and contain 1-3% 1-2 mm quartz phenocrysts in a sericite- and quartz-rich recrystallized felsic ash matrix. Pumice fragments vary in size from 5 mm-2 cm and are carbonatized or silicified. The well developed foliation in the rock trends east - northeast.

At this location, the pyroclastic deposits contain abundant chloritoid and iron carbonate. Rusty red-brown iron carbonate alteration locally comprises 20% of the outcrops and occurs in lenses and patches from 2-30 cm in diameter. Thin, 1-2 mm wide veinlets of prismatic dark green chloritoid and sericite are typically aligned parallel to the foliation. Locally, 15-20% of the unit is composed of these chloritoid-sericite veinlets.

STOP F-5: "L" PYROCLASTIC FLOW DEPOSITS

This is an outcrop of the lower "L" quartz-phyric pyroclastic flow deposit. This unit correlates to the east with the immediate hanging wall rocks to the Mattabi deposit "A" lens (see Fig. 9).

At this location, 15-20% 1-5 mm pale blue quartz phenocrysts (average diameter approximately 3 mm) are set in a massive grey or grey-green altered ash matrix. The well developed foliation again trends east-northeast.

The alteration assemblage chloritoid + sericite is responsible for the colour of this outcrop. This assemblage is present in thin, wispy veinlets that trend parallel to the foliation.

MATTABI - AREA "16"

The Mattabi volcanogenic massive sulphide deposit was discovered by Mattagami Lake Mines Ltd. in 1969 from follow-up diamond drilling of an airborne geophysical anomaly.

The upper sections of the orebody were mined from 1972 to 1980 using open pit methods, and the deeper ore was mined using underground methods from 1980 until reserves were depleted in 1988. The deposit produced 12.55 million tons of ore grading 8.28% Zn, 0.74% Cu, 3.31 oz/ton Ag and 0.85% Pb. The orebody comprised 5 stratiform lenses of massive sulphide ore separated by stringer mineralization or barren host rock. The lenses occurred within three distinct stratigraphic successions (Fig. 9).

STOP M-1 HIGH LEVEL LAKE MESOBRECCIA (HLMBX) DEPOSITS AND QUARTZ CRYSTAL ASH FLOW TUFFS (HLQAT)

This series of outcrops is located approximately 500 m stratigraphically below the lower lens of the Mattabi orebody. The southern portion of these outcrops is composed of coarse heterolithic debris flow deposits, which are interpreted to be mesobreccia deposits formed from caldera collapse (See Stop F-1). The far northeastern outcrops (Fig. 8) comprise felsic quartz-phyric pyroclastic flow deposits, which overlieand are interfingered with the mesobreccia deposits and are believed to represent the first intercaldera pyroclastic material. Figure 8 shows the location of the outcrops described below.

M-1A - MESOBRECCIA (HLMBX)

These outcrops consist of coarse heterolithic lithic-rich breccia that contain up to 50% 1-25 cm light-coloured subangular felsic lithic fragments, <10% block-sized pumiceous fragments commonly with silicified rims, and <5% amygdaloidal mafic fragments up to 5 cm in diameter, set in a matrix composed dominantly of alteration minerals. In thin section, the matrix mineral assemblage consists of 40% fine-grained quartz, 12% chloritoid, 30% Mg-chlorite, 15% sericite ± pyrophyllite and 3% opaque minerals. The northeast section of these outcrops contains an outcrop-sized felsic fragment (>10m) with similar composition and texture as the smaller felsic fragments.

M-1B - HIGH LEVEL LAKE QUARTZ CRYSTAL ASH FLOW TUFF (HLQAT)

The HLQAT at this location consists of massive light-tan to grey coloured quartz-phyric ash flow tuff.

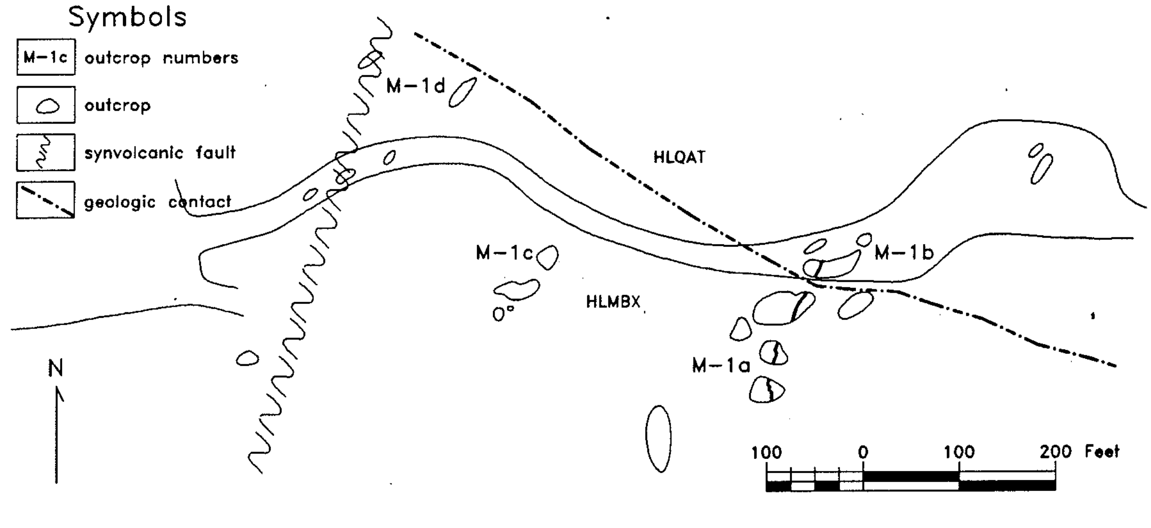


Figure 8 Geological sketch map of the High Level Lake Succession outcrops south of the Mattabi Mine.

Quartz crystals (2-5%) range in size from 0.5 to 1.5 mm. In thin section, only "ghosts" of 2-10 mm pumice (20%) can be observed in a matrix composed of recrystallized inequigranular "chunky" quartz (up to 65%), 25% fine sericite, Mg-chlorite (6%), 3% pyrite, and 1% fine opaque minerals. Further to the east, a coarse fragmental texture is observed in outcrops of this unit. The rock consists of coarse (0.5-60 cm) subangular quartz-phyric felsic lithic fragments (60-70%) contained in a matrix petrographically similar to the fragments. This texture has been interpreted as a post-depositional slumpage feature of preconsolidated ash flow tuff. Only minor variations in the alteration mineralogy can be found between the matrix and fragments.

M-1C - HIGH LEVEL LAKE MESOBRECCIA (HLMBX)

This exposure comprises pumice-rich mesobreccia, which contains 20-30% 2-30 cm rounded highly-vesicular (scoria?) fragments and <10% 1-15 cm felsic lithic fragments. The felsic fragments are similar to fragments seen further east (Stop M-1a) and are indicative of the HLMBX unit.

In thin section, the matrix consists of 30% quartz, 20% chlorite, 10% Fe-rich carbonate, 15% sericite ± pyrophyllite, 10% chloritoid, 10% ragged andalusite and 5% opaque minerals. Chlorite alteration increases toward the center of the outcrop where massive chloritic veining up to 50 cm wide occurs.

This chloritic veining is possibly the result of fresh seawater recharge into a hot hydrothermal system causing abrupt increases in magnesium content and thus forming chlorite alteration. The quartz-filled tension fractures evident in this outcrop are probably due to volume changes during hydrothermal alteration. Kyanite can be found in these fracture-fillings on the south section of the outcrop. A 10-20 cm wide band of silicified rock, striking N-S, bisects the outcrop. Samples from this band contain 50% quartz, 15% chloritoid, 15% sericite or pyrophyllite (probably replacing andalusite), 8% Fe-rich carbonate, 7% chlorite, 2% relict andalusite, and 2% opaque minerals. This band of more intense alteration probably represents a conduit for upward high- temperature fluid movement, which was later overprinted by sericite (+K) and chlorite (+Mg) alteration.

M-1D - MESOBRECCIA (HLMBX) AND SYNVOLCANIC FAULT ZONE

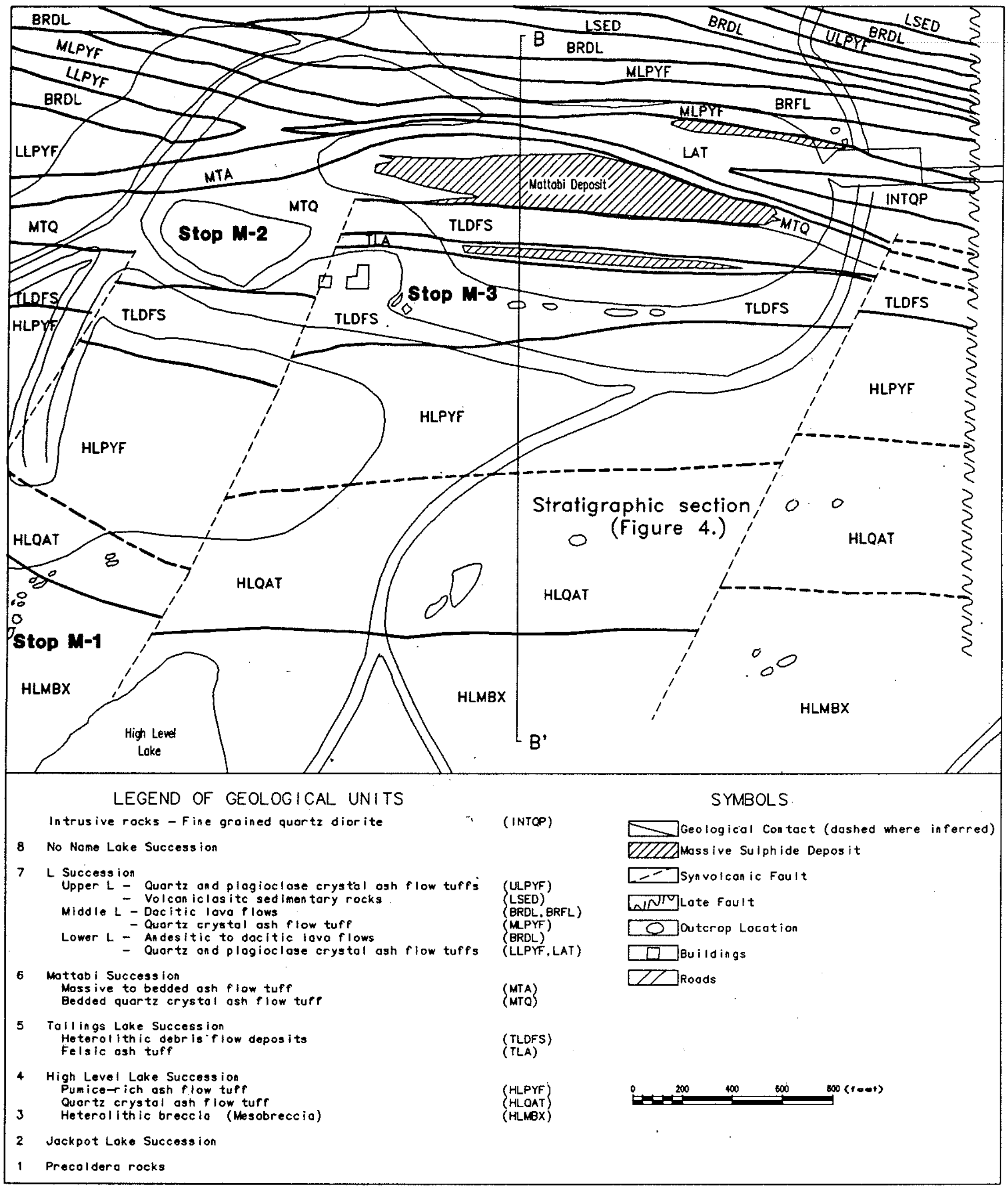


Figure 9 Geological -plan map of the Mattabi area, with field trip stop locations.

This series of outcrops shows a definable, confined and symmetrically zoned increase in alteration intensity through a synvolcanic fault structure within the HLMBX unit. The texture of the HLMBX unit near the synvolcanic fault is strongly overprinted but is still recognizable due to the abundance of relatively unaltered felsic lithic fragments. Alteration in the small outcrop on the road and immediately north of the road consists of 1-4 cm clots of Fe-carbonate-rich material surrounded by anastomosing veinlets of quartz, chloritoid ± andalusite. The dominant change in rock mineralogy from the previous outcrop (M-1c) is an increase in the amount of Fe-rich carbonate and chloritoid.

STOP M-2 - MATTABI ASH FLOW TUFFS (MTQ)

This outcrop consists of Mattabi Succession quartz crystal ash flow tuff. This exposure contains 10-50% 5-30 mm rounded cherty juvenile felsic fragments, 5-15% 5-30 mm rounded pumice, filled by quartz ± sulphide, and 2-7% 0.5-1.5 mm colourless quartz crystals. The recrystallized altered ash matrix consists of fine quartz (40-50%), chloritoid (10-30%), sericite ± pyrophyllite (15-35%, after andalusite?), andalusite (0-5%) and opaque minerals (1-5%). Lithic fragments consist of quartz ± sericite ± pyrophyllite and the pumice are filled and/or replaced by quartz ± sulphide.

The outcrop shows some alignment of pumice (weathering pits) striking 105° and dipping steeply to the north. This feature may indicate primary bedding.

STOP M-3 - TAILINGS LAKE HETEROLITHIC DEBRIS FLOW DEPOSITS (TLDFS)

The Tailings Lake Succession consists dominantly of highly-variable heterolithic mafic debris flow deposits (TLDFS) and lesser bedded ash horizons (TLA). The debris flow deposits vary in fragment composition, fragment abundance and bedding characteristics. Three fragment types are dominant: lithic mafic fragments, commonly replaced by chlorite and carbonate (0-50%); fine-grained cherty felsic fragments (0-30%); and rounded pumiceous fragments (0-20%). Bedding is uncommon, but where present, is usually defined by sorting of the various fragment types.

In the outcrop exposure beside the mine ventilation shaft many of the characteristics of the TLDFS unit are conspicuous. This exposure is atypical in its cross-sectional view and in the presence of bedding in the TLDFS unit. Here this unit contains 2-30 mm mafic fragments (5-30% carbonate-replaced weathering pits) and 2-30 mm felsic lithic or pumice fragments (5-30%). In thin section, the matrix consists of 35% quartz, 20% carbonate and 10% andalusite which has been overprinted by 20% Mg-chlorite and 10% sericite ± pyrophyllite. Bedding is defined by sorting of mafic vs. felsic fragments. It appears that mafic fragments are normally-graded and felsic fragments (pumice) are reversely graded. Bedding trends approximately 100° and dips 60-70° to the north.

M-4 - NO NAME LAKE ANDESITE (NNL)

The No Name Lake Succession consists of andesitic pillow lavas, pillow breccias, sheet flows and interflow sedimentary rocks. Only the upper section of this succession is exposed in outcrop; the lower sections are observed only in drill core and appear to consist dominantly of thick amygdaloidal flows.

The first outcrop (behind the core racks, Fig. 10) of this stop consists of thin (30-70 cm thick) sheet flows with 5-25% oval-shaped carbonate-filled amygdules (2-30 mm in diameter) generally aligned in the strike direction. In thin section, these rocks are composed of 20% aligned fine laths of plagioclase and 15% quartz in a secondary matrix of 40% chlorite, 10% biotite, and 2% epidote.

Pillow breccia and hyaloclastite are exposed in the outcrop east of the water tower. These rocks consist of approximately 25% amygdaloidal pillow fragments (5-50 cm) in a matrix consisting of 30% hyaloclastite and 70% amygdaloidal flows. The pillow breccia fragments contain 10-20% carbonate filled amygdules (now weathering pits) which range in size from 1-4 mm. The hyaloclastite portion of the matrix consists of 0.5-2 cm recrystallized massive angular andesite fragments. The flow portion of the matrix consists of vaguely defined amygdaloidal (20-50% amygdules) flows. Flows become the dominant rock type on the north portion of the outcrop.

Pillowed flows are exposed in the outcrop north of the water tower. This rock consists of well-formed 1-4 m (in the long dimension) amygdaloidal pillows with 25-30%, 1-20 mm carbonate-filled amygdules. The amygdules show a bimodal size distribution; most amygdules are 1-4 mm, with another distinct grouping at 5-20 mm. Massive pillow selvages vary from 5-15 cm thick.

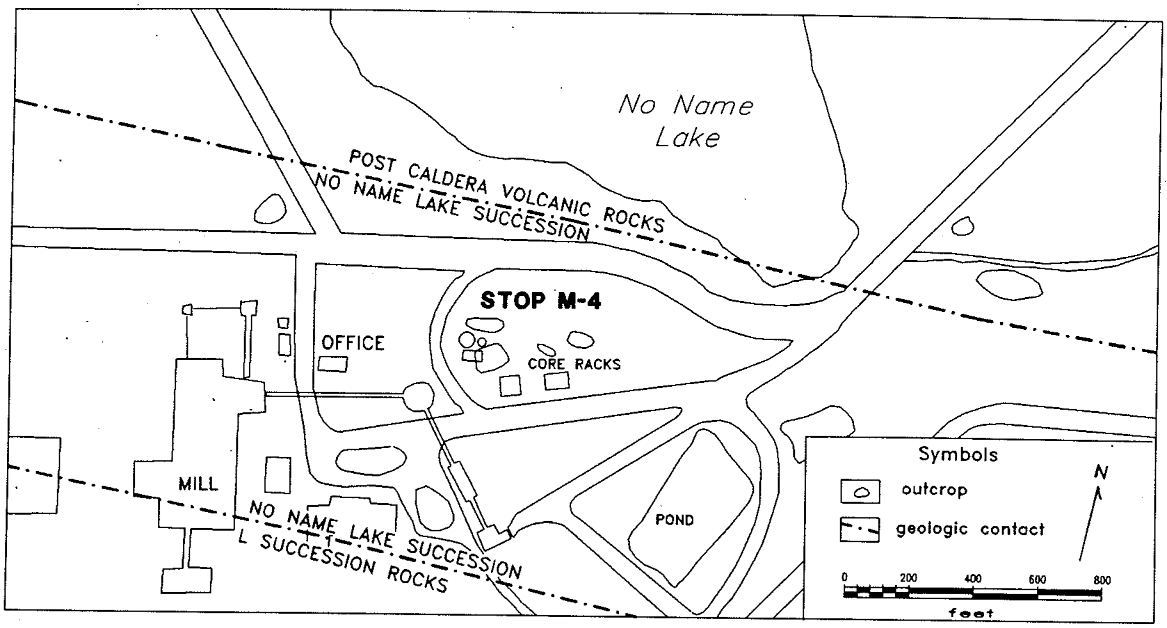


Figure 10 Geological plan map of the No Name Lake Succession outcrops north of the Mattabi Mine (Stop M-4).

STRATIGRAPHIC AND STRUCTURAL GEOLOGY OF THE LYON LAKE MASSIVE SULPHIDE DEPOSIT, STURGEON LAKE, ONTARIO[1]

E.R. Koopman[2] M.R. Patterson[3], J.M. Franklin[2] and K.H. Poulsen[2].

INTRODUCTION

The Lyon Lake massive sulphide deposit is the second largest volcanogenic massive sulphide deposit in the South Sturgeon Lake district. It has many of the typical characteristics of volcanogenic massive sulphide deposits, but lacks a conspicuous alteration zone, and has some unusual stratigraphic relationships, with iron formation in the footwall.

A combined airborne magnetometer and electromagnetic survey was flown over the Sturgeon Lake area for Mattagami Lake Mines Ltd. in March 1969. Ground geophysical follow-up of strong 6 channel anomalies led to the drilling which intersected the footwall iron formation and sediments.

Exploration diamond drilling along the geological contact at the Sturgeon Lake Mine deposit led to the discovery of the Lyon Lake massive sulphide deposit in October 1971. The Creek Zone, halfway between the Lyon Lake and Sturgeon Lake deposits was discovered in February 1972. Deep diamond drill testing for a proposed shaft site late in 1974 discovered a fourth ore zone, the Sub-Creek, at a vertical depth of 500m.

MINING

Production mining at Lyon Lake commenced in November 1980, with total production of 3,172,540 tons at a grade of 8.67% Zn, 1.26% Cu, 4.50 oz/ton Ag, and 0.99% Pb (to December 31, 1989). Mining is currently being carried out at a rate of 32,000 tons per month with current reserves sufficient to maintain production until December 1991. The full extent of the orebody has not yet been defined and exploration drilling and development is being carried out between the 1900 and 2600 foot levels.

The principal mining method is longhole open stoping utilizing trackless equipment, however, significant amounts of "jumbo mining" is being carried out in the flat lying ore zones. The limited size of the individual ore lenses requires intensive development. Production muck is moved to 1 of 5 ore passes which feed the 1800 foot level. This tracked level is the main tramming horizon where 10 ton battery locomotives move the ore to a single set of passes which feed the 2700 level crusher station at the shaft. The 3 compartment shaft was originally sunk to a depth of 1347 feet, but was deepened in 1982 to its current depth of 2950 feet. Most ore is skipped from the 2800 level to surface, but crushing facilities on the 1000 level and skipping from the 1200 level can be used for any ore mined above the 1000 level. Additional access to the underground for men and material is gained via a ramp which currently extends from surface to the 1900 foot level.

Ore is trucked 5 miles west to the Mattabi concentrator for custom milling.

LYON LAKE MINE GEOLOGY

The Lyon Lake deposit occurs at the top of the caldera sequencee (Morton, R.L., this volume). The mafic unit of the overlying cycle is in fault contact with the top of the lowermost cycle and forms the hanging wall to the Lyon Lake, Creek, Sub-Creek Zones and Sturgeon Lake Mine. The stratiform deposit is hosted by a blue quartz phyric fragmental rhyolite. The footwall to the ore consists of an upper rhyolitic unit composed

[1] Contribution to COMDA

[2]Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario, K1A 0E8

[3]Noranda Mines, Mattabi Mines Limited, Lyon Lake Division

of interbedded ash and lapilli tuff, immediately underlain by a lower rhyolitic unit of coarsely fragmental rock. Below the footwall rhyolite, a sequence of sedimentary rocks comprised of banded iron formation, graphitic shale, massive po-py bodies, greywacke, and siltstone, provides a stratigraphic marker unit. In the lower parts of the mine, a locally developed heterolithic breccia is spatially related to the ore horizon as host, footwall and hanging wall rocks to the ore. The sedimentary sequence overlies, and is possibly in fault contact with the underlying massive intermediate to felsic flows. Massive sulphides form several stacked or en echelon lenses, and comprise discontinuous and contorted bands of coarse grained sphalerite and pyrite. Locally centimeter-scale lenses of galena occur within sphalerite bands.

The dominant structural feature controlling ore distribution is a major open fold characterized by a shallowly plunging hinge line trending east-southeast (120/16) (Dube et. al., 1989). Hinge lines of mesoscopic folds, mineral lineations, and striations measured on bedding, foliation, and fault planes are all subparallel to the hinge line of the major fold and result from stretching along the fold axis. The hanging wall basalt truncates the ore, felsic volcanic, and sedimentary rocks. Based on kinematic and geometric structural relationships, this discontinuity is believed to represent a thrust fault related to folding. A structural contour map of the hanging wall - footwall contact indicates that the dip of this contact is shallowing in an eastward direction reflecting the plunge effect of the major open fold (Koopman et. al., 1989). The shallow dip of this contact as a result of folding reflects the deformation at Lyon Lake. It has resulted in the re-orientation of portions of orebodies into attitudes that are much shallower than the steep regional dip. The contour map (Fig. 4) also illustrates that the Sub-Creek Zone is the down-plunge extension of the Lyon Lake Zone.

MINE UNITS (Fig. 1)

FELDSPAR PHYRIC DACITIC FLOWS

This unit is dacitic, and forms the lowermost rock type observed in the underground workings, and only at one locality (2600 level). The unit is massive, characterized by 20 to 30%, 1 to 3mm size euhedral to subhedral feldspar phenocrysts in a siliceous and sericitic matrix. Locally, evidence of fragments of similar composition within massive portions near the top contact of the flow suggest flow top brecciation. The contact with the overlying sedimentary rocks is an apparent 50m thick tectonic zone of highly strained rock characterized by millimeter to centimeter thick alternating bands of sericite-rich and chlorite-rich altered dacite.

SEDIMENTARY STRATA

A sequence of sedimentary rocks including banded iron formation, graphitic shale, massive po-py lenses, greywacke, and quartz-rich pelites occurs 15 to 60 meters below the ore horizon depending on thickness and amount of folding. Most of the deformation observed at the mine is taken up by these sediments; the banded iron formation (as well as the sulphides) being the most strongly deformed due to inherent competency differences. The sedimentary unit ranges in apparent thickness from 20 to 60 meters, and may be thicker since the lower sedimentary contact was observed at only one locality (2600 level east) in the underground workings.

The sedimentary rocks consist of a basal Fe-oxide or Fe-carbonate banded iron formation in possible fault contact with the underlying dacitic flows, overlain by sulphidic graphitic shale, which is overlain by interbedded greywacke, medium-grained felsic quartzose pelite, and graphitic pelite. Locally, massive po-py lensoid bodies occur above the banded iron formation. Sedimentary strata associated with iron formation have an average thickness of 15-45 cm (Dube et. al., 1989) and typically consist of 1 to 5 mm chloritic and pyritic ovoid blebs alligned parallel to the foliation. Several quartz-feldspar rich beds alternate with chlorite-rich and sericite-rich beds. Graphitic shale generally occurs as a massive unit with fine laminations separated by a thin layer of pyrrhotite, but, individual graphitic beds up to a meter thick are also interbedded with greywacke and quartzose pelite.

BANDED IRON FORMATION

Iron-carbonate, -silicate, -oxide, and/or -sulphide banded iron formation forms the basal member of the sedimentary strata in the mine workings, and from drill core intersections, extends laterally along strike east and west of the mine. This unit has not been observed

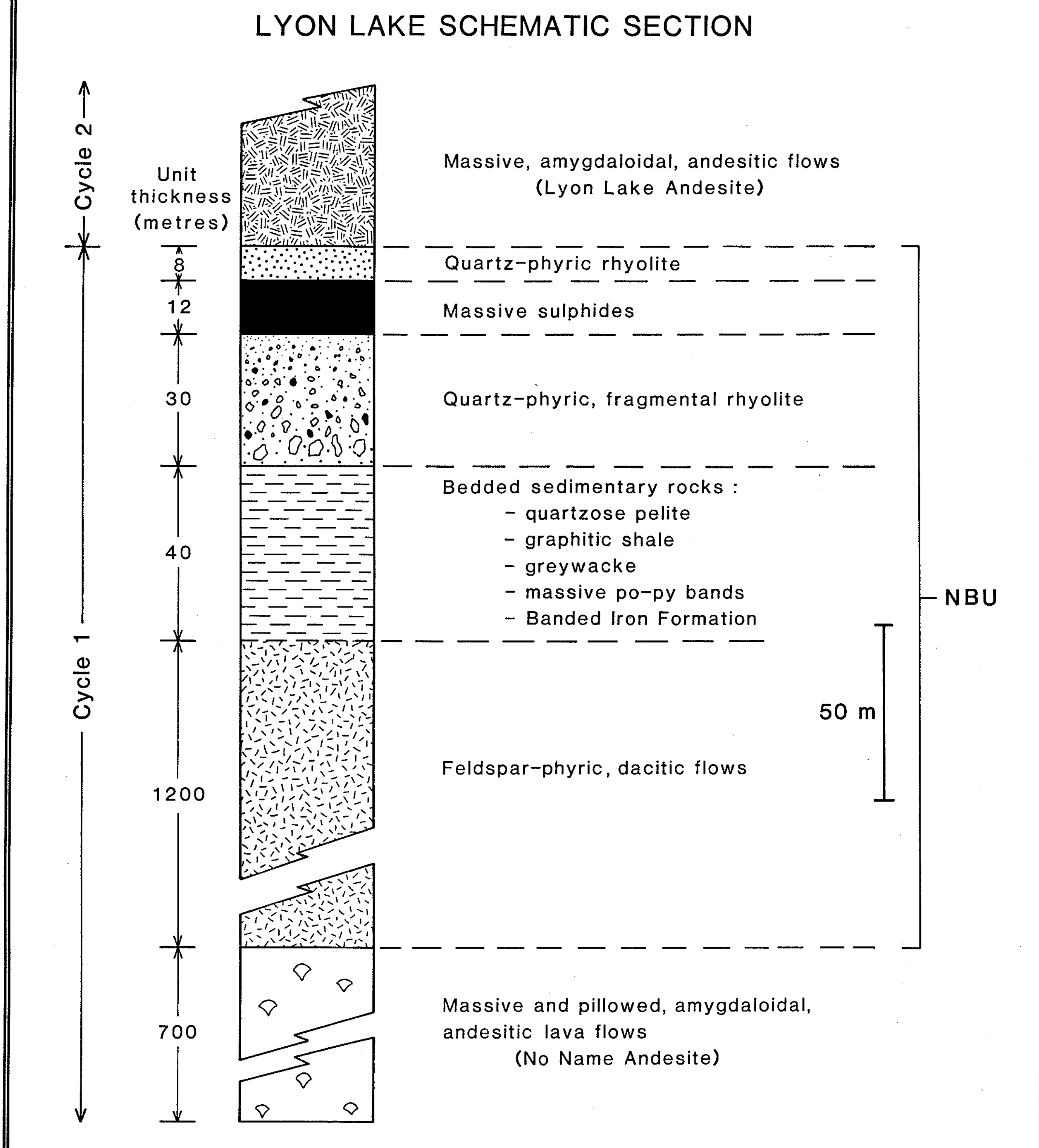


Figure 1 Idealized stratigraphic column of the Lyon Lake mine area. Modified from Dube et. al., (1989).

in outcrop. Based on underground mapping and drill core information, banded iron formation occurs from near surface through to the 2600 level, and probably continues at depth, as a continuous stratigraphic marker. It is typically thin-banded, laminated, or forms laminae within discreet bands. Thin bands and laminations of iron formation associated with greywacke and volcanic rocks, indicate that this banded iron formation is a typical Algoma Type iron formation (Gross, 1965). Compositional variation is observed from level to level, but, in general, carbonate iron formation occurs in the upper (surface to 1000 levels) and lower mine levels (1700 to 1800 levels), and iron-oxide iron formation is developed in the middle levels (1000 to 1700 levels), and deepest level of the mine (2600). The compositional variation may represent lateral facies changes, and not the presence of more than one iron formation because the iron formation can be traced from one level to the next. Iron-sulphide iron formation consists of magnetic or non-magnetic pyrrhotite rich layers locally interlayered with iron-silicate, iron-carbonate, or iron-oxide iron formation. Sulphidization may have occurred locally to produce the iron sulphidic layers.

Schistosity and cleavage are well developed in the iron-silicate and cherty beds, parallel to bedding in the limbs of folds, and at high angle to bedding in the hinge area of folds.

At the 525 to the 825 level, non-magnetic iron-carbonate and cherty beds are interlayered with iron-silicate beds (dominantly chlorite). The thickness of individual bands that are not folded or that are on the limbs of folds varies from 2.5 to 15 cm thick. Stringers and veinlets of pyrrhotite and pyrite are developed in the cherty layers subparallel to the foliation.

From the 900 to 1700 levels the iron formation is composed of strongly magnetic iron-oxide interbedded with chert, iron-silicate, and iron-carbonate. Laminations of strongly magnetic pyrrhotite occur in the iron-carbonate beds, and locally in the iron-silicate beds. Magnetite occurs as fine laminations in cherty beds, or as monomineralic 2 to 3 cm thick beds. From the 1580 to 1700 levels, the banded iron formation is basal to, and intercalated with an apparent thick zone of the complete sedimentary strata described above, whereas, at the 1000 level, sediments other than the iron formation are less abundant and the iron formation is associated with quartz-phyric volcanic flows stratigraphically above and below. The variation in thickness of these sediments may be explained partly by primary deposition, but is more likely a structural thinning and thickening (boudinage) effect resulting from deformation.

At the 1800 level, iron-carbonate banded iron formation similar to that observed in the upper mine (Dube et. al., 1989) occurs, and is associated with graphitic shale and siltstone.

At the 2600 level, banded iron formation is characterized by white cherty bands alternating with iron-silicate and iron-oxide layers of uniform thickness. Locally, near the contact of the underlying feldspar-phyric dacitic dome, the iron formation is segmented and individual fragmented beds are hosted in a graphitic pelite matrix.

HETEROLITHIC FRAGMENTAL UNIT

This unit is spatially associated with the ore, forming the footwall, host, and/or hanging wall, from the 1400 level to the 1850 level. It consists of sub-rounded to angular cherty, chloritic, carbonaceous, sulphidic, and graphitic fragments supported in a graphitic silty matrix. Cherty white fragments are the most abundant and largest fragments and make up 40 - 50% of the rock. Chloritic, graphitic, and sulphidic fragments are less obvious in the graphitic matrix, and less abundant than the cherty fragments. These fragments are generally <1cm to 5 cm in size, and make up approximately 10% of the rock. The fragments are oriented parallel to the foliation and have aspect ratios of 2:1 to 10:1, however, in zones of intense deformation, aspect ratios may be as high as 20:1. This unit is locally associated with the ore. Both fragment and matrix composition its local distribution suggests this unit represents a talus deposit composed of rock fragments similar in composition to the iron formation and graphitic sediments.

RHYOLITE

The sedimentary sequence is overlain by a fine-grained quartz-phyric rhyolite which forms the immediate host and footwall to the massive sulphide bodies. The rhyolite varies from massive to lithic fragmental and is characterized by blue, and, less commonly, grey and black euhedral quartz crystals ranging in size from 1 to 3 mm. The abundance of quartz phenocrysts shows a gradational variation from 1 to 15% depending on stratigraphic position. Typically, the immediate host rock contains the most abundant, larger and dominantly blue quartz crystals, whereas deeper in the footwall, quartz phenocrysts are less abundant, smaller, and grey and black. Lithic fragments grade from coarse volcanic breccia at the base to interbedded felsic ash and lapilli tuff towards the top. The fragments include cherty and felsic lithic clasts ranging from ash to block size in a quartz-phyric, chlorite- carbonate-, and sericite-bearing matrix.

BASALT

The mafic member (Lyon Lake Andesite) of the overlying volcanic cycle forms the hanging wall to the massive sulphide horizon. It is compositionally basalt (Franklin et al., 1977). Wherever the contact between the hanging wall and footwall rocks was observed underground,the base of this unit is strongly schistose, and the top of the footwall rhyolite is strongly cleaved. Based on underground mapping and drill core logging, this basalt is observed to truncate the host rhyolite, ore, and footwall sedimentary sequence. The basalt consists of feldspar-phyric massive to amygdaloidal sheet-flows and fragmental rocks. Amygdules comprise up to 5% of the rock and vary in size from 2 to 20 mm, and shape from rounded to elongate. In the strongly schistose zone, amygdules have aspect ratios up to 5:1. Abundant 1 to 10 mm thick, irregularly distributed quartz and iron-carbonate stringers and fractures occur parallel to the foliation. Locally, in undeformed areas, centimeter thick chlorite layers were observed which may represent sheet-flow contacts, defining 20-30 cm thick sheet-flows. Pillows were not observed underground or in outcrop in the mine area.

INTRUSIVE ROCKS

Two types of medium-grained intermediate to felsic dykes, as well as fine grained mafic dykes are present in the mine. Intermediate calc-alkaline dykes are the most abundant, consisting of 20 -30%, 2 to 5 mm amphibole phenocrysts, altered to chlorite, linearly oriented subparallel to the foliation (Dube et. al., 1989). These dykes, although not ubiquitous, occupy the contact between the hanging wall basalt and footwall rhyolite. Fine-grained, dark-green mafic dykes are most abundant in the middle to lower levels of the mine and are most commonly observed along the upper ore contact, and within the ore lenses, as irregular deformed intrusive bodies. The attitudes of all types of dykes varies from parallel to high angle relative to the host strata and are commonly folded.

SULPHIDE LENSES

Ore lenses within the Lyon Lake Mine are relatively small, irregular massive sulphide lenses averaging approximately 15,000 to 20,000 tons each. The Lyon Lake, Creek and Sub Creek zones are not single ore lenses, but groupings of spatially-related smaller ore zones.

The ore lenses consist of massive and banded fine-grained sulphides with variable amounts of gangue. The principal sulphides are pyrite, sphalerite, chalcopyrite, galena, and pyrrhotite with minor amounts of tetrahedrite, arsenopyrite, and tennantite. Gangue to the ore consists of aggregates of quartz and carbonate and inclusions of sheared wall rock and boudinaged dykes. Quartz is the most abundant gangue mineral, followed by carbonate, with minor chlorite, muscovite, rare phlogopite and possible clinozoisite (Wolfson, 1988).

STRUCTURE

FOLDING

The dominant structural feature controlling ore distribution is a (major) open fold characterized by a shallow hinge line trending east-southeast (Fig. 2), (Koopman, et. al., 1989; Dube et. al., 1989). Although the rocks generally strike east-west, face and dip 65^{o}-75^{o} north, detailed structural underground mapping and drill hole information indicates that all stratigraphic units are folded in the mine area, and this deformation extends southeasterly to the Sturgeon Lake Mine area. At Section 12,200E (Fig. 2), this flexure is located at the 1480 level and has a wavelength of up to 100 meters. In contrast to the upper and lower levels of the mine, all the lithologic units within the flexure are thicker and subhorizontal or shallowly dipping northeast or southeast. In this area, the ore dips shallowly, in the nose of the fold. Figure 3 shows the pole distribution of the ore contact with the surrounding rocks measured on all levels of the mine; defines a near cylindrical fold characterized by a calculated hinge line oriented at $120^{o}/16^{o}$.

Asymmetrically folded ore, iron formation, and

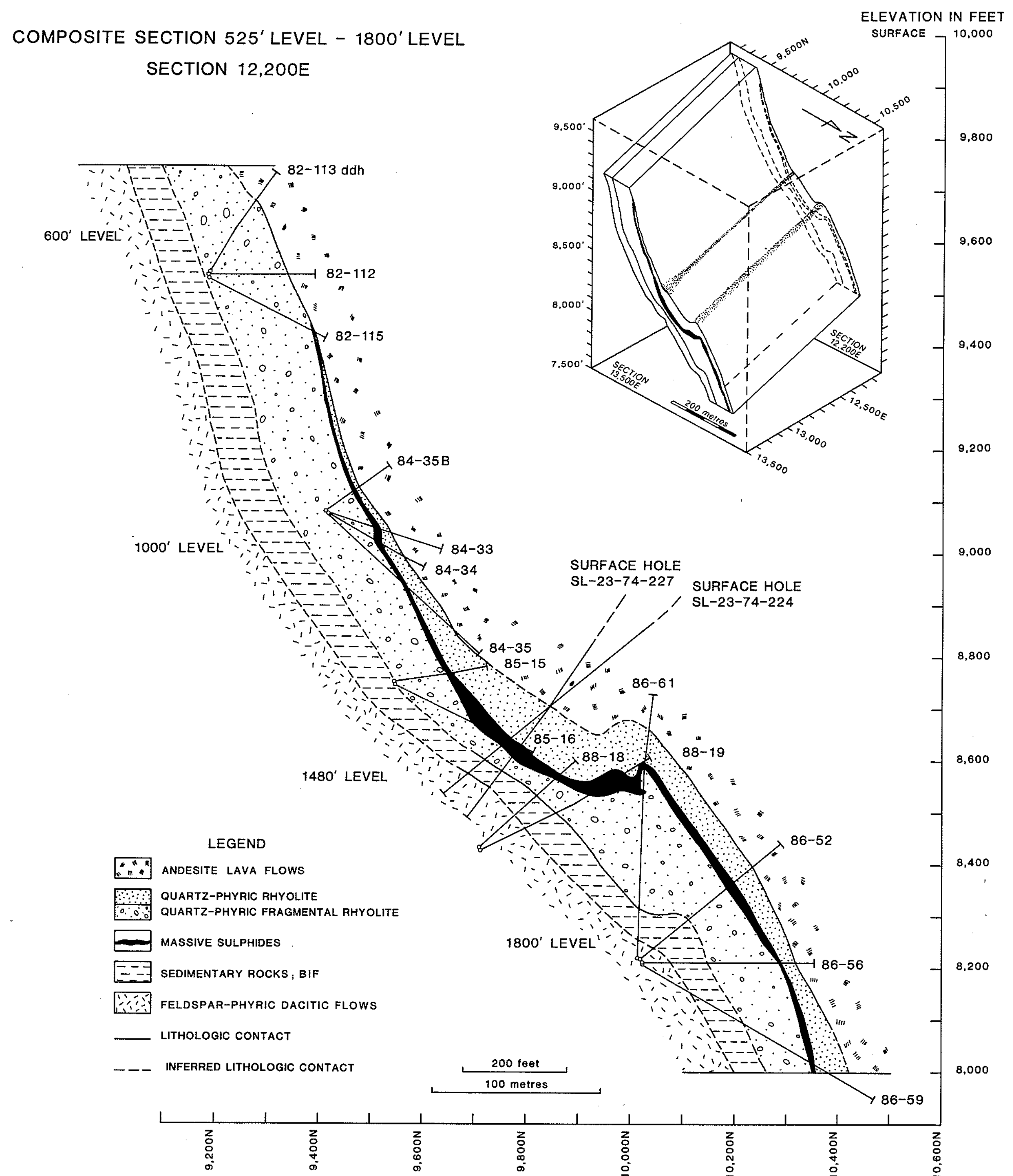


Figure 2 Composite cross section (12,200E) of the Sub Creek Zone. Inset: isometric projection of the deposit showing the plunge of the major flexure. North arrow indicates Mine North which is 40° east of true north. (Note: Section 13,500E is a projection of Section 12,200E).

sedimentary strata, with wavelengths of up to 10 meters are abundant on all levels of the mine and have a southward vergence (Bell, 1981) on north dipping limbs and a northward vergence on south dipping limbs of the larger fold. In the iron formation, the mechanism of folding is flexural and the folds belong to classes 1b and 1c of Ramsay (1967). Figure 3 illustrates the pole distribution of bedding and the hinge lines measured on these smaller folds. The measured hinge lines trending 094° and plunging 38° east are very close to the pole distribution which is characterized by a calculated hinge line trending 102° and plunging 35° east. The axial surface related to this fold is oriented at 295°/72° (Fig. 3), (Dube et. al., 1989), which is subparallel to the regional foliation. The plunge of these mesoscopic folds is steeper than the plunge of the major flexure which may be a reflection of the primary orientation, or the more irregular orientation and thickness of the ore lenses in contrast to the surrounding strata.

All types of dykes have been folded and show vergence relationships similar to bedding planes. Locally, apparent non-folded dykes cross-cut highly deformed iron formation, whereas in other areas, dykes of similar composition may be strongly folded. The nonfolded dykes may be affected by wavelength and amplitude of the folds, original orientation, or different fold patterns resulting from contrasting rheological properties between the dykes and intruded rock (Dube et. al., 1989).

Evidence of earlier isoclinal and refolded folds has been observed on the limbs of larger folds in the iron formation on the 1000 level and 1580 ramp. In the 1580 ramp, subhorizontal axial planes may be related to an earlier folding event, however, distribution of poles to bedding planes of these refolded folds is consistent with that of the whole mine (Fig. 3). Therefore, fold axes of earlier folds and the major flexure are probably coaxial. Alternatively, because iron formation is easily deformed, these earlier folds may have a primary soft-sedimentary origin, later reoriented by the event that established the fold pattern of the mine area.

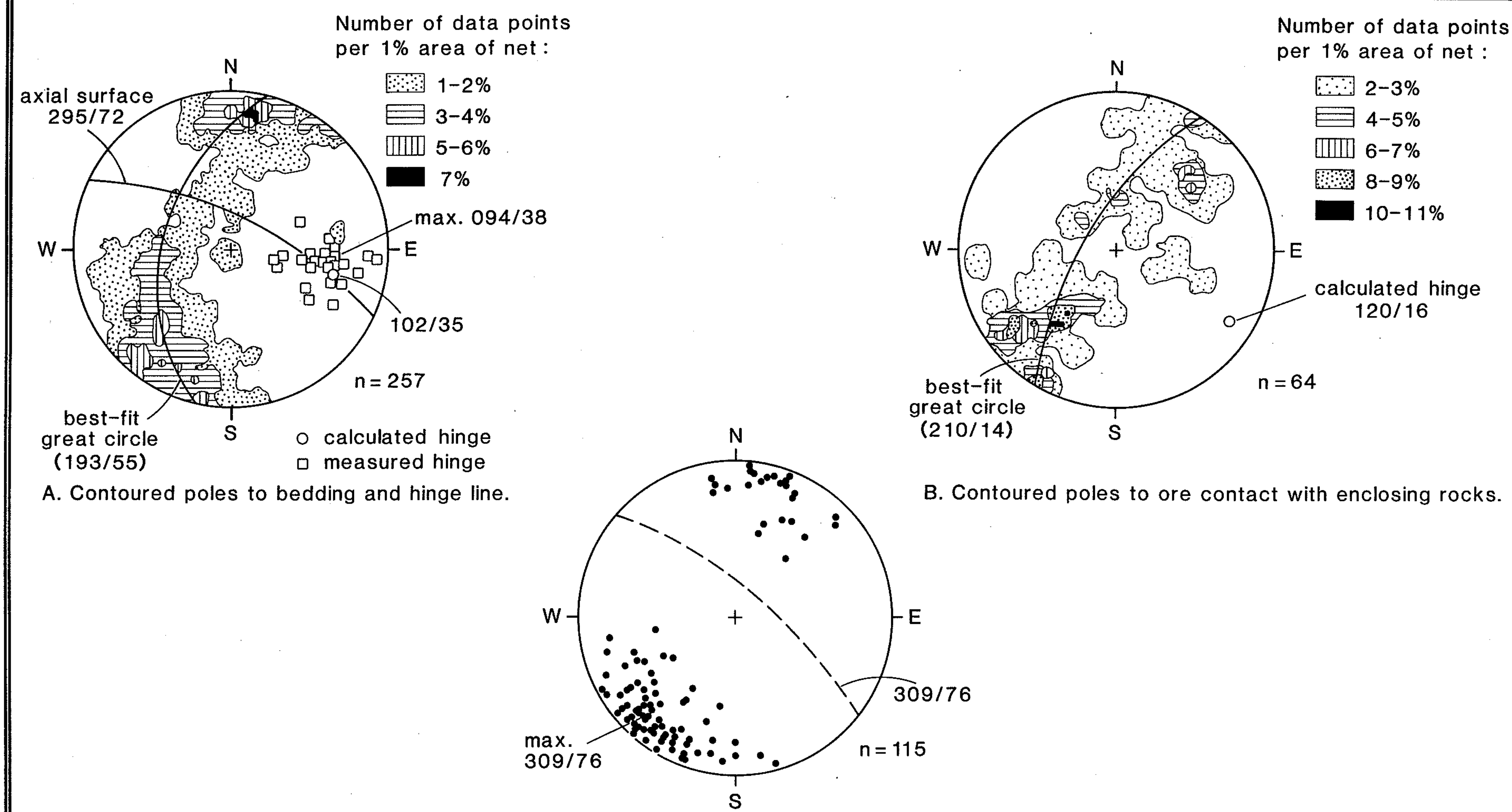
A schistosity is developed in the more ductile rocks such as the basalt the iron-silicate layers of the iron formation, and the matrix of the heterolithic fragmental unit, and a cleavage is developed in the felsic rocks. The average measured orientation of this foliation is 309°/76°, axial planar to the folds, striking sub-parallel to, but dipping more steeply than the overall bedding envelope (Dube et. al., 1989). Local striations on bedding and foliation planes plunge east, sub-parallel to the measured fold hinges of mesoscopic folds.

Subhorizontal quartz veins and en echelon patterns of quartz veins are common throughout the mine in zones of moderately to strongly foliated basalt, felsic rocks, and heterolithic fragmental and locally in felsic dykes. The average orientation of these veins is subperpendicular to the axial planar foliation, therefore, probably related to the folding event and indicating the direction of stretching in the strata.

Irregular, discontinuous, and lenticular zones of quartz breccia are located at or near the contact between the ore and footwall and/or hanging wall rocks. These zones are most abundant in the middle to lower parts of the mine (hinge area of the major flexure), locally containing up to 40% sulphide and probably representing dilatant zones formed during folding.

HANGINGWALL-FOOTWALL CONTACT

The orientation of the contact between the hanging wall basalt and footwall quartz-phyric rhyolite varies from easterly, north of the Mattabi mine area, to northwest in the Lyon Lake and Sturgeon Lake Mines area, to more easterly again, west of the Lyon Lake Mine area. Detailed underground mapping, core logging, and surface mapping indicates that this contact is characterized by a strong schistosity in the basalt, and a well-defined cleavage in the rhyolite, oriented subparallel to stratigraphy at an average of 286°, dipping steeply 50° to subvertical. This high strained zone has been observed to be more than 100 meters wide on the 600 level (Dube et. al., 1989). Most of the deformation seems to be taken up in the more ductile hanging wall basalt. Striations measured on cleavage surfaces in the rhyolite are subparallel to hinge lines of mesoscopic folds and may be related to the folding. Subhorizontal quartz veins observed in this high strain contact zone suggest vertical extension. Figure 2 shows that this contact is folded, and the asymmetry of the fold suggests relative motion in section is reverse. Striations and mineral lineations measured on foliation and bedding planes, and the orientation of hinge lines, are all subparallel to each other, suggesting the folding and faulting are related. Four possible timing relationships may occur. Firstly, if the shearing and folding are coeval, the reverse motion in section, coupled with the oblique orientation of linear structures suggests oblique dextral movement of the contact fault.



A. Contoured poles to bedding and hinge line.

B. Contoured poles to ore contact with enclosing rocks.

C. Poles to schistosity and cleavage.

Figure 3 Equal area projections (lower hemisphere) of folds and their related structural elements: A- Contoured poles to bedding. Crosses represent measured hinge lines and the open circles represent the location of the hinge line according to the best fit great circle; B- Contoured poles of the ore contact with enclosing rocks. The open circles showsthe location of the hinge line according to the best fit great circle; C- Poles to foliation.

Secondly, if the shearing and folding are coeval, local layer parallel shearing would develop on the limbs of the fold. Thirdly, if this shear is folded by a later deformation, than the asymmetry of the fold is a geometric consequence only, and gives no information about the fault sense. Fourthly, if shearing along this contact occurred after folding, the axial planar cleavage of the fold would refract into the shear foliation.

Figure 4 is a structural contour map of this contact, and illustrates a steep zone - flat zone - steep zone in a northward direction reflecting the limb-nose-limb of the major fold. The contours also show that the dip of this contact is shallowing in an eastward direction reflecting the plunge effect of the major open fold, and that Sturgeon Lake Mine possibly represents the surface expression of the fold nose of the major fold.

UNDERGROUND TOUR

Due to continuous production and development, specific stop locations will be decided upon in August 1990, in order to provide the best exposures for the field trip participants.

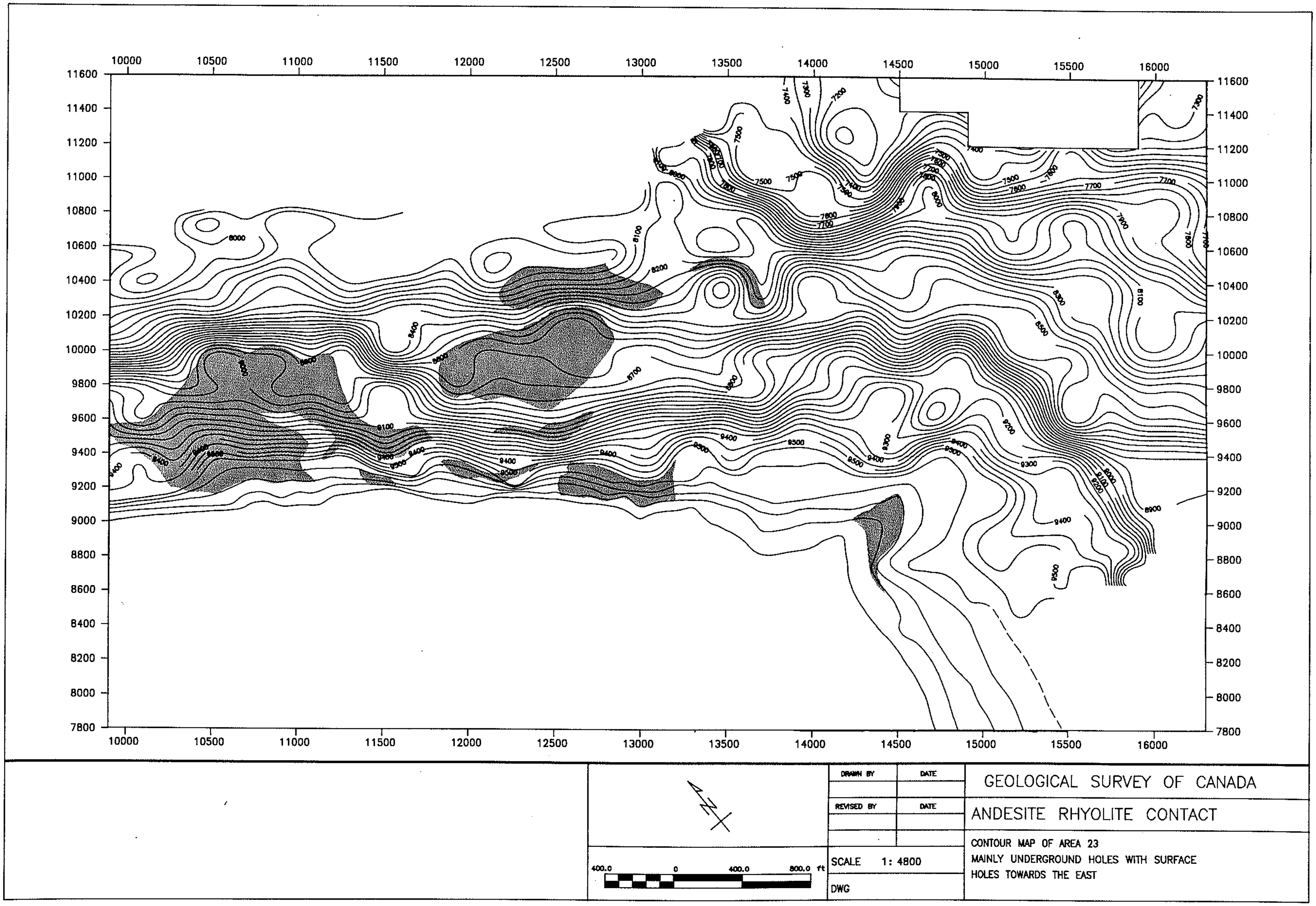


Figure 4 structural contour map of the contact of the hanging wall basalt and footwall felsic volcanics illustrating that the dip of this contact is shallowing in an eastward direction reflecting the plunge effect of the major open fold. Shaded areas are plan view of stoped ore zones.

PROTEROZOIC GEOLOGY OF THE THUNDER BAY TO MARATHON AREA

J.M. Franklin
Geological Survey of Canada,
601 Booth Street,
Ottawa, Ontario.
K1A 0E8

The Proterozoic rocks are shallow-dipping, and in the Thunder Bay area lie unconformably on the peneplained Archean surface of the Superior Province. The Archean rocks are part of several northeast-trending "belts" of metamorphosed and complexly deformed volcanic and sedimentary rocks, intruded by felsic and intermediate to ultramafic rocks.

The lithostratigraphy of the Proterozoic rocks is shown in Table 1.

EARLY PROTEROZOIC

North of Lake Superior, The Proterozoic sequences include the Gunflint and Rove formations. The Gunflint Formation has been studied by Goodwin (1956, 1960) and Moorhouse (1960). Its mineral assemblages are described in detail by Floran and Papike (1975). Detailed descriptions of fossils from the Gunflint Formation area recorded by Barghoorn and Tyler (1965), Barghoorn (1971) and Edhorn (1973). The Rove Formation has been described by Morey (1969) and Guel (1970, 1973). Much of the descriptive detail is taken from these authors.

GUNFLINT FORMATION

The Gunflint Formation extends continuously for 177 km northeasterly from Gunflint Lake to beyond Thunder Bay, from where it can be traced intermittently to the Slate Islands (Sage et al., 1975) southwest of Schreiber. It has an average thickness of 122m (Goodwin, 1956), although drilling near Thunder Bay indicates a much greater thickness in this more easterly area. Except for local faulting and brecciation the Gunflint Formation is structurally simple and uncomplicated, with an average southeast dip of five degrees.

Goodwin (1956) established the principal stratigraphic divisions in the western area of the Gunflint Formation, and determined that its deposition was in part cyclical. A basal conglomerate member is overlain by two members, each composed of chert, tuffaceous shale, and carbonate-taconite (siliceous grainstone) submembers. These members are in turn overlain by a discontinuous limestone member. Shegelski (1982 and this volume) noted that the detailed stratigraphic relations in many areas of the Gunflint Formation are more complex than suggested by Goodwin. The principal members, described below, are usually present, but the stratigraphic relationships between the various "facies" are in part the products of diagenetic changes, and in part reflect deposition in a series of semi-isolated basins. The principal members described by Goodwin (1956) are as follows:

(a) Basal Kakabeka Conglomerate Member: This member is polymictic conglomerate, to 1.4m thick. Clasts of Archean volcanic rocks, granite and quartz are cemented by a matrix of chlorite and quartz. The unit is highly irregular in thickness but laterally persistent.

(b) Lower Member: Algal chert, consisting of reef-like mounds of finely banded black, red, and white oolitic chert which are intergrown and cemented in dolomite, forms the western margin of Gunflint exposure. This chert contains abundant microfloral remains (Barghoorn and Tyler, 1965, Barghoorn, 1971; Edhorn, 1973).

The lower tuffaceous shale submember ranges to 6m thick. It overlies the lower algal chert submember in the area west of Kakabeka Falls and is composed of fissile black shale containing much volcanic ash.

Table 1 Proterozoic Stratigraphy
Thunder Bay District

Osler Group	Tholeiitic basalt, minor rhyolite and sedimentary strata Basal clastic unit
Local Disconformity	
Sibley Group	
Kama Hill Formation	Clay and K-feldspar-rich mudstone
Rossport Formation	Dolomite (locally sandy), chert, stromatolites
Pass Lake Formation	Quartz arenite, minor conglomerate
Disconformity	
Animikie Group	
Rove Formation	Shale, greywacke
Gunflint Formation	Basalt,conglomerate (Kakabeba member), algal chert, siliceous grainstone, ferruginous and dolomitic carbonate units, shale, tuff
Unconformity	
	Archean rocks

The uppermost part of the lower member is subdivided into three "facies". The "lower west taconite facies", extends northeast from Gunflint Lake to Kakabeka Falls and is composed of wavy-banded granular chert-grainstone, carbonate, and oxides. The lower half contains disseminated greenalite granules in pale grey chert; siderite forms local beds. The upper half contains upwardly increasing amounts of hematitie and magnetite. This facies grades upward into jaspilitic algal chert and laterally into the "lower banded chert-carbonate" facies. The latter facies extends from Kakabeka Falls to Thunder Bay city, and consists of 10 to 15 cm siderite beds, with interbedded 5 to 15 cm grey cherty beds. Carbonaceous material and pyrite are common in shale interbeds. This facies grades into grainstone (taconite) towards the northeast.

The "lower east granular taconite facies" extends from Thunder Bay to Loon Lake, and consists of basal interbedded granular chert and ankerite, and upper interbedded red to green mottled chert and dolomitic limestone.

(c) Upper Member: The "upper algal chert submember" extends west from Nolalu to Gunflint Lake and consists of basal granular chert overlain by algal chert and, in the Mink Mountain area, possible amygdular basalt flows. The flows and algal chert are overlain by granular chert and bedded jasper. Jasper beds grade into tuffaceous shale of the overlying submember.

The upper tuffaceous shale is the only continuous submember in the Gunflint Formation and forms a key stratigraphic marker. It ranges to 30 m thick and thins laterally in either direction from Kakebeka Falls. It consists of black tuffaceous shale and siltstone with interbedded siderite and pyrite and extensive beds of volcanic ash. The ash contains ellipsoidal accretionary lapilli composed of concentric layers of small angular tuff fragments, arranged about a larger central vitric fragment of crystal. Hassler and Simonson (1989) described a variety of textures, and interpreted the unit to have formed as a result of hydroclastic eruption, and deposited from turbidity currents. Similar lapilli have formed due to the accumulation of volcanic dust in water droplets and on water coated shards (Moore and Peck, 1962).

The upper tuffaceous shale submember grades into the upper "taconite" (grainstone) and chert-carbonate submember. The upper taconite facies extends from Gunflint Lake to Thunder Bay and is composed of wavy bands of granular greenalite-bearing chert. The greenalite-bearing granules are round to oval, evenly distributed throughout a layer, and appear to have formed "in situ". The unit exhibits rusty weathering, contains abundant hematite and magnetite in granules towards the top, and grades westward into chert-carbonate. The latter "facies" consists of interbedded grey chert and brown carbonate (siderite with lesser dolomite and ankerite). Brecciation and folding, apparently contemporaneous with deposition, are common.

(d) Upper Limestone Member: The upper limestone member marks the top of the Gunflint Formation. Minor chert beds, illite and volcanic shards are present, and tuffaceous shale is most prevalent in the eastern area of Gunflint exposures.

Goodwin (1956) concluded that the Gunflint was deposited in a shallow basin which had limited circulation with an open sea. After initial algal activity in the neritic zone, volcanic activity (tuffaceous shale) was accompanied by sinking of the basin. Silicate-bearing material was deposited in the deepest portions while in the neritic, or intertidal zone (between Kakebeka Falls and Thunder Bay) banded chert-carbonate formed. Farther to the northeast, the "lower east taconite facies" formed in agitated, oxygenated waters. As the basin filled, conditions of algal growth returned, initiating the "Upper Gunflint" cycle.

Volcanic activity, marked by local basalt flows, terminated deposition of the upper algal chert and resulted in widespread distribution of pyroclastic materials of the upper tuffaceous shale. Downwarping resulted in deposition of granular iron silicate rocks in the deeper, southwestern portion of the basin, while on the shallow northeast shore, chert-carbonate was deposited. As the basin filled, sporadic but violent volcanic activity was accompanied by the entry of sea water, resulting in formation of the upper limestone.

Goodwin (1956) in drawing an analogy with the Santorin volcano of the Aegean Sea, suggested that volcanism was the chief source of iron and silica. Alternatively, Hough (1958) suggested deposition in fresh water basin, with material derived through weathering of an adjacent landmass, and deposition controlled by limnic cycles.

ROVE FORMATION

The Rove Formation conformably overlies the Gunflint Formation. In the Thunder Bay and Pigeon River area, it attains a thickness of at least 380 m, and possibly more than 610 m (Guel, 1970). The formation consists of three lithologic units (Geul, 1970) which are, from the base upwards:

1. black pyritic shale and argillite (base)
2. interbedded argillite and greywacke (transition sequence of Morey, 1969)
3. quartzitic greywacke with minor argillite interbeds (top)

The lower argillite is the dominantly exposed unit in Ontario. It contains ellipsoidal carbonate concretions of up to 1 m in principal axis length; a few concretions have complex form, and all appear to "replace" the shale, as its laminations can commonly be followed through the concretions. The transition sequence consists of thin-bedded greywacke, is the thickest unit of the Rove Formation and is exposed mostly in northeastern Minnesota. Morey (1969, 1972) noted that the detrital material in the Rove Formation was derived from Archean terrane to the North.

The metamorphic age of the Rove Formation is considered to be 1.7Ga (Peterman, 1966; Franklin, 1978a), but Franklin et al. (1986) interpret both it and the Gunflint Formation to have been deposited at about 1.9 to 2.2 Ga.

MIDDLE AND LATE PROTEROZOIC

THE SIBLEY GROUP

The name "Sibley" was first assigned to this group by Tanton (1931). Franklin et al. (1980) divided the Sibley Group into three formations; the Pass Lake Formation, composed of quartz arenite, is overlain by the Rossport Formation, a carbonate-rich unit. The uppermost Kama Hill Formation is a mudstone. Each of these has been divided into members by Cheadle (1986).

The age of the Sibley Group is a key question with respect to the position of these rocks within the framework of the Keweenawan tectonic event, and the Late Precambrian polar wandering curve. An Rb/Sr whole rock isochron gives an age of 1339 $\pm$ 31 Ma (^{87}Rb = 1.42 x 10^{-11} yr $^{-1}$) The Sibley Group lies non conformably on quartz-feldspar porphyry exposed on the western side of Lake Nipigon ($1536.7^{+10}/_{-2.3}$ Ma), and is overlain by the Osler Volcanic rocks ($1097.3^{+4.8}/_{-4.6}$ Ma) and cut by Logan diabase intrusions (1109.7$\pm$ 2 Ma) (Davis and Sutcliffe, 1986).

SUMMARY OF THE DEPOSITIONAL HISTORY

The Sibley Group was deposited in an elongate, north/south basin which was initially probably deepest in the south. Lacustrine conditions prevailed throughout its history (Cheadle, 1986). The basin formed relatively rapidly and along its margins alluvial fans of conglomerate were formed from locally derived and rapidly deposited material. The initial period of rapid deposition gave way quickly to relatively slow deposition of the arenites of the Pass Lake Formation. The basin transgressed northwards towards the end of the period of deposition of the Pass Lake Formation and extended far north (at least to southwestern Lake Nipigon) during deposition of the lower member of the Rossport Formation. Moderately rapid regression marked the middle stage of Rossport deposition, accompanied by increased clastic deposition and stromatolite growth to the north of Nipigon and chert precipitation to the south. Transgression subsequently followed as the basin extended northward at least to Armstrong, north of Lake Nipigon. The basin was relatively constant in depth, and it slowly filled with clay-rich dolomite. The transition to the Kama Hill Formation marks a change from predominantly sub-aqueous to predominantly sub-aerial deposition; Cheadle (1986) noted a reversal in paleocurrent direction from predominantly southward during the deposition of the Pass Lake and Rossport formations, to north-westward during deposition of the Kama Hill Formation. Deposition of the Kama Hill Formation continued in a tectonically stable mud-flat environment. Primitive life flourished during quiescent periods of deposition of the Kama Hill sediments.

REGIONAL STRUCTURE OF THE THUNDER BAY AREA

The Proterozoic rocks of northern Lake Superior are only slightly deformed. Both the Early and Late Proterozoic sequences dip gently (3° to 10°) towards the axis of the Keweenawan synclinorium (Fig. 1). Faulting, most of which is relate to the Late Proterozoic mid-continent rift, is prominent, and some structures may be related to Early Proterozoic tectonism.

The earliest faults, suggested by Shegelski (1982) to be pre -or syn-Gunflint deposition, may be related to development of the Animikie basin, during early rifting.

Faults active during the Keweenawan tectonic event may be divided into three principal groups. the earliest set are poorly represented N-S faults that may have been active at the time of deposition of the Sibley Group, and locally contain uranium mineralization (Yule, 1979). In a few areas, these faults are occupied by diabase dykes. North of Nipigon, they may have been reactivated during late-Keweenawan tectonism.

The most prominent Keweenawan faults are three N 60 E sets (Fig. 1). The most westerly set corresponds approximately to the western margin of Early Proterozoic outcrop, from Gunflint Lake to Thunder Bay city. Another set corresponds to the major dyke swarm that forms the shoreline and islands near the U.S. border, and extends through Silver Islet to Edward Island. The most easterly set bounds the northwest side of Isle Royale. These three sets are members of Craddock's (1972) "strike-faults", which bound the axial horst of the Keweenawan basin. Most of these faults parallel the axis of the Lake Superior syncline, dip towards the centre of the lake, and have reverse movement. However, the major movement on the most westerly fault set near Thunder Bay is normal. The other sets may also have had early listing normal movement, and those that are occupied by gabbro dykes were obviously active at, or prior to, the major period of Keweenawan magmatism.

The youngest faults are radial to the major flexure in the Keweenawan rift, centered south of Nipigon. These form major lineaments in the basement. Some are occupied by late mafic dykes, and the Sibley Group is offset along a few which have major vertical movement.

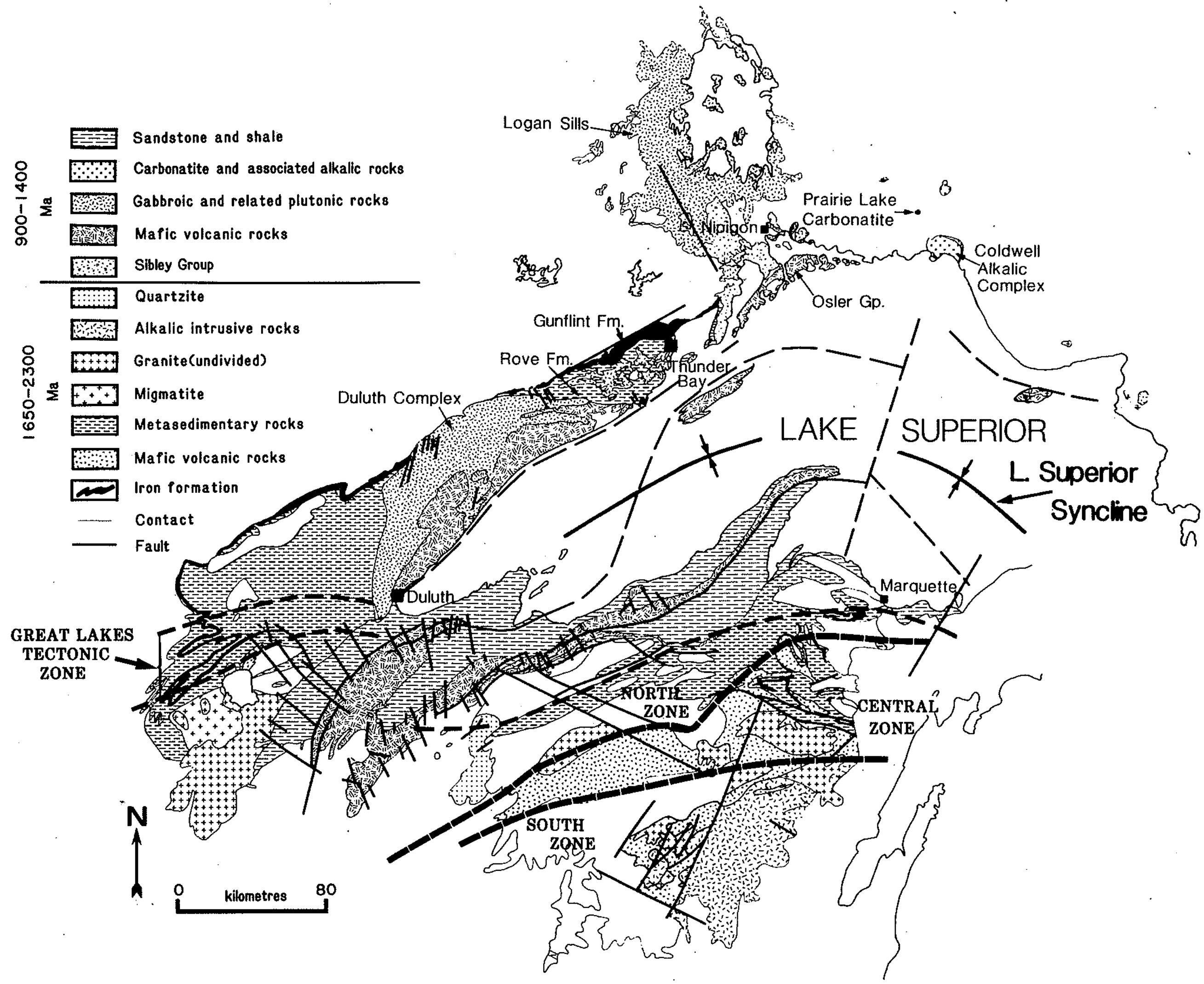


Figure 1 Geology of the Proterozoic rocks of the Lake Superior area, after Morey (1978b) and Ontario Geological Survey maps.

THE GUNFLINT FORMATION IN THE THUNDER BAY AREA

R.J. Shegelski
ESSO Resources Ltd.,
Calgary, Alberta.

INTRODUCTION

The Gunflint Formation is an early Proterozoic stable shelf which unconformably overlies a peneplained Archean basement (Fig. 2). Minor exposures of basal Gunflint Formation occur in the Rossport-Schreiber area and on the Slate Islands, but the bulk of the outcrop is found between Gunflint Lake and the northeast corner of Thunder Bay.

In discussing the Gunflint chemical sedimentary rocks, a general classification parallel to that used for limestones is employed to describe chert and carbonate rocks alike. There is a complete gradation in composition between chert and carbonate end-members and, secondly, identical textures exist in both the chert and the carbonate units. Replacement fronts between chert and carbonate are observed at all scales, making determination of the precursor difficult. In the present field guide I have assumed that calcium carbonate was the precursor to all chert-carbonates because of textural similarities to Phanerozoic carbonate sequences. Much, but not all, of the chert component of these rocks originated during diagenesis. The orthochemical component of the chert-carbonate rocks is either micrite (silicified equivalent would be chert) or spar (silicified equivalent is crystalline quartz). The allochemical component consists of granules, intraclasts, oolites, pisolites and shards in decreasing order of abundance. Granules of undetermined origin, but generally considered to be arenite-sized rounded fragments of previously deposited chemical sediment, are the most abundant allochem and form the chemical sedimentary rock grainstone (generally referred to as 'taconite' by previous workers).

A wide variety of carbonates occur in the Gunflint Formation and include ferroan dolomite (ankerite), calcite, dolomite and siderite in decreasing order of abundance. Chert beds within the Gunflint Formation contain a variety of minor minerals which result in diagnostic colors as follows - cryptocrystalline silica (white), hematite (red), limonite-goethite (brown), septechlorite-chamosite (green), anthraxolite-kerogen (black to grey).

This portion of the Guidebook is adapted from Shegelski (1982). A more extensive description is available in that Guidebook.

DESCRIPTION OF STOPS AND ITINERARY, GUNFLINT FORMATION

KM	Description
0.0	Intersection Highways 11-17 and 61
0.4	Entrance to Valhalla Inn
4.1	Mapleward Road
6.6	Railroad Overpass
8.7	Junction Highway 130
14.8	Railroad overpass
20.1	Junction Highway 588
25.0	Kakabeka Falls LCBO
25.7	Entrance to Kakabeka Falls Park
26.5	Junction Highway 590 - TURN LEFT and proceed west
32.3	Junction Highway 595
50.5	Junction Highway 588 - TURN RIGHT and proceed west
54.2	Bridge across Whitefish River Turn off the road on the right hand side into a small parking area just before the bridge. Cross the road and examine the riverside exposures along the north bank of the river.

STOP 1: WHITEFISH FALLS

This is an excellent exposure of the basal Gunflint Formation and includes all of the lithologies typical of the basal member. Here, a thin irregular cover of basal conglomerate (commonly referred to as "Kakabeka conglomerate") overlies an Archean diorite intrusion. The basal conglomerate consists of boulders derived from the diorite which are set in a chlorite

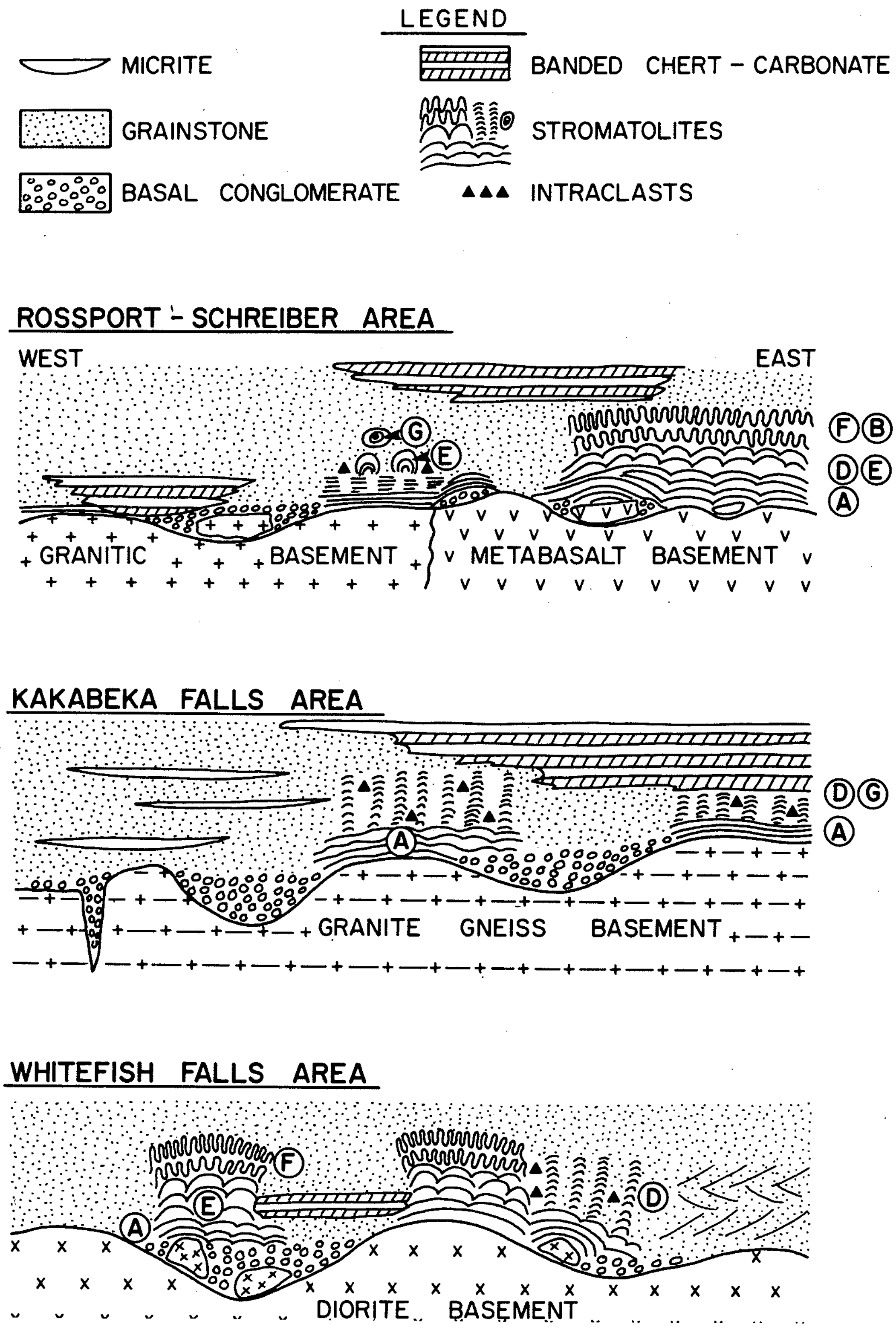


Figure 2 Facies association of the basal member of the Gunflint Formation

cement and sandy matrix. The basement and the basal conglomerate are overlain by (bio) chemical sedimentary rocks including large (up to 2 m diameter) stromatolite bioherms containing oolites and crudely banded or crossbedded granular chert-carbonate. The bioherms are cherty stromatolites which locally display color zonation from siliceous bleached cores or greenish iron silicate-rich cores to reddish hematitic rims. The bioherms are morphologically zoned (forms as defined by Hofmann, 1969) and generally consist of planar bases (Form a) grading through bulbous centers (Form E) to columnar tops (Forms B, D, F) (Fig. 2). Locally, oncolites (Form G) are found between bulbous and columnar forms. This upward increasing diversification of stromatolites is typical of the basal Gunflint and has also been observed at Kakabeka Falls and in the Rossport-Schreiber area.

The interface between stromatolite bioherms and chemical grainstone is defined by a distinctive cellular texture which has formed from the binding of granules by a cryptalgal cement. Minor layers, lenses and intraclasts of chemical mudrock occur in the troughs of columnar stromatolites as well as in the surrounding grainstones. Rare replacements of mudrock by magnetite occur near the base of the formation and, combined with chamosite-rich grainstone, represent the only occurrences of iron formation at this stop.

A reconstruction of facies relationships from several localities in the basal Gunflint, including Whitefish Falls, is presented in figure 3. Such a three-dimensional reconstruction is possible because both cross sections and bedding planes are readily observed at several localities in the Gunflint Formation. All other occurrences of Gunflint, chemical sedimentary rocks generally conform to this facies reconstruction although slight differences, particularly the absence of well developed bioherms and an increase in the proportions of grainstone and/or chert-carbonate, may be present.

0.0	Turn around and retrace route to proceed east along Highway 588
3.7	Junction Highway 590 - TURN LEFT and proceed East.
21.9	Junction Highway 595
27.7	Junction Highway 11-17

STOP 2: JUNCTION OF HIGHWAYS 590 AND 17-11

2A: LOWER BASAL MEMBER

Walking along the base of the cliff one progressively encounters - (i) basal conglomerate which is similar to that found in the Kaministikwia River to the north, (ii) crude interlayers of green chamositic chert and associated white chert, and brown ankeritic carbonate. These two sedimentary rocks unconformably overlie and are in fault contact with Archean basement, here a greenstone, but a gneissic complex to the west. Proceeding westward there are steep dilatant fractures in the gneissic basement. These fractures strike subparallel to highway and are filled with Gunflint basal conglomerate. The conglomerate contains rounded siliceous pebbles and cobbles and is cemented with marcasite and calcite. The outcrop contains two generations of faults; the earlier (Early Proterozoic) were developed prior to or during deposition of the Gunflint and were subsequently filled by basal conglomerate, whereas the later faults (Late Proterozoic) displace the Archean-Proterozoic unconformity.

2B: UPPER BASAL MEMBER

Return to the junction and walk uphill on the west side of Highway 590. Proceeding uphill one encounters laminated cryptalgal chert overlain by a wavy bedded grainstone-micrite facies. The latter is capped by chaotic, slumped, laminated chemical sedimentary rocks which are probably cryptalgal. A few meters further uphill, overlying the previous sequence is a brecciated and slumped pyritic black chert. This latter unit is identical to the basal chert in the hydro spillway (next stop) and, along with the underlying units, forms the basal member of the Gunflint Formation in this area.

0.0	Proceed east along High 11-17 towards the town of Kakabeka Falls.
0.7	Entrance to Kakabeka Falls Provincial Park. Turn left into Parking Lot of Kakabeka Falls Provincial Park (sampling with hammers in the park is not allowed).

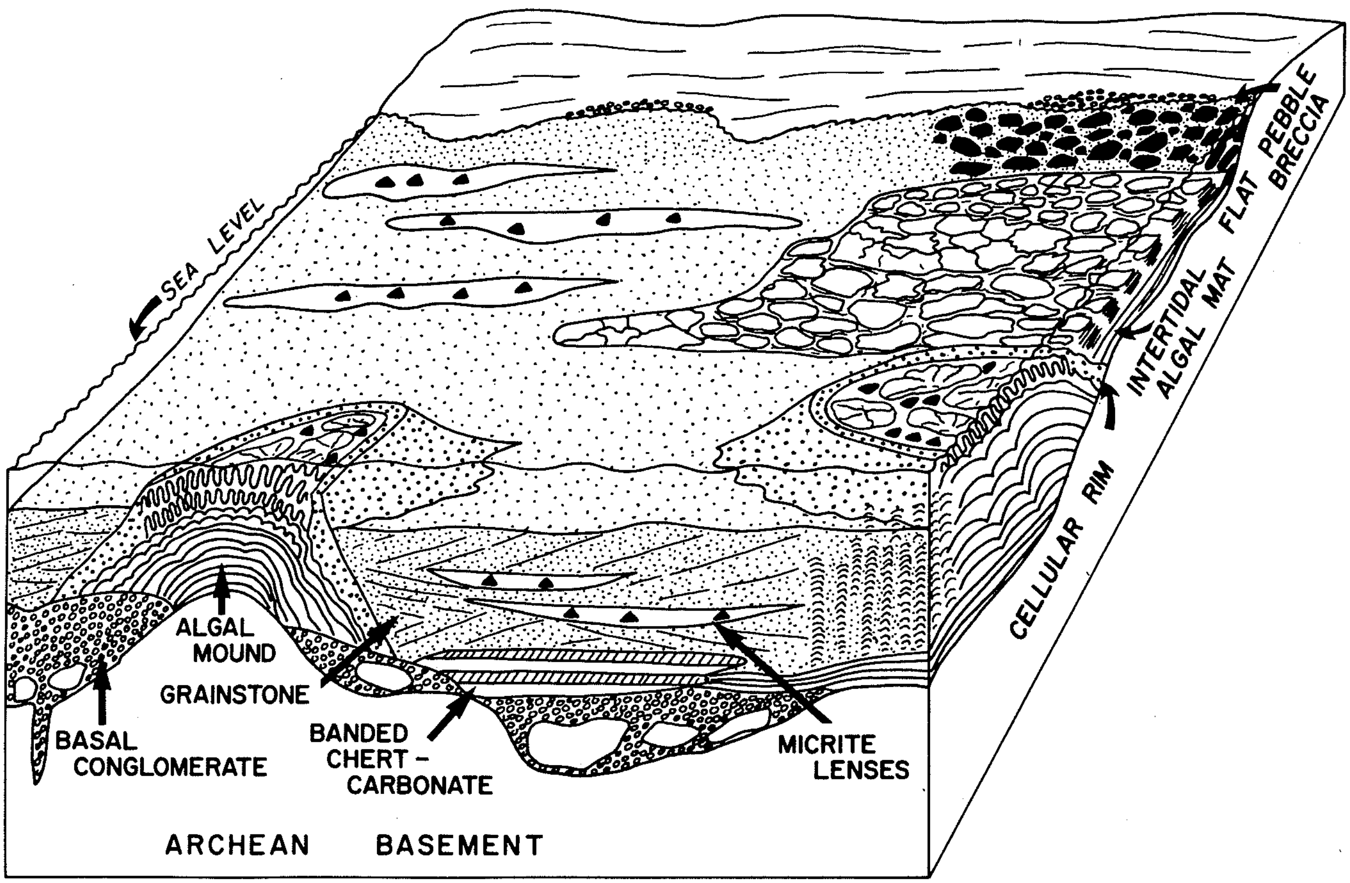


Figure 3 Reconstruction of intertidal-supratidal facies in Gunflint chemical-sedimentary rocks

STOP 3: KAKABEKA FALLS

The outcrop on the north edge of the parking lot contains excellent layers of banded chert-carbonate within black fissile shale. The alternating dark grey chert and brown siderite-ankerite layers display slump and soft sediment deformation features. Microscopic examination of banded chert-carbonates reveals delicate lamination in the chert which resembles the "ribbon texture" of algal mats. The interlayered carbonate bands contain complex microspherical structures which likely resulted by nucleation from a gel state. Local thick beds of carbonaceous siderite (2-3 wt% carbon) form carbonate iron formation; contemporaneous deposition of carbon and carbonate suggests biological activity during iron deposition.

Cross the parking lot to the observation decks and observe the Falls. The lip of the Falls is at the same stratigraphic level as the top of the spillway and is formed by the resistant cap of the lowermost chert-carbonate band. Looking down the gorge one can observe the lapilli-tuff member as a lighter grey unit near the base overlain by a thick sequence of black shales. The section measured at Kakabeka Falls is section B of figure 4. Note that shale is the predominant lithology in the Kaministikwia sections and this is, in fact, typical for the Gunflint Formation in general throughout the Thunder Bay region.

Following the river upstream within the park towards the swimming area, an excellent exposure of the lapilli tuff member, a marker horizon that occurs throughout the lateral extent of the Gunflint Formation. This unit is exposed in outcrops beneath the highway bridge. These beds, along with pyritic chert-carbonate, calc-arenite and shale, are also very well exposed in a spillway section about 300m below the falls. Access to this section is difficult and dangerous, as permission must be obtained from Ontario Hydro to enter the power plant.

The lapilli consist of a central shard, or less commonly a crystal of plagioclase or pyroxene, surrounded by a sphere of fine-grained volcanic dust.

Following the road further upstream to the swimming area, good outcrops of the basal algal chert are exposed in the river bed just beyond the change houses. This unit is correlative with that observed at Whitefish Falls.

0.0	Return to Highway 11-17 and TURN RIGHT to proceed East
0.7	Kakabeka Falls LCBO
1.3	Junction Oliver Road - TURN LEFT
1.4	Turn right after crossing tracks10.7
	Murillo Hotel - Country Tavern
18.1	Junction Highway 130
26.3	Junction Belrose Road (also called Central Avenue) - TURN LEFT
27.1	On the left or west side of the road is a gate and entrance to Dickenson's Quarry which is now private property of Great West Timber. Permission must be obtained to enter the quarry.

STOP 4: DICKENSON'S QUARRY

At the north edge or far side of the quarry, there is an excellent 11 meter thick cross-section of shale-grainstone facies. This facies consists of alternating black to brown fissile shale and cherty grainstone layers and lenses. Although the facies is crudely layered due to local predominance of either lithology, the section is predominantly composed of shale. There is widespread intercalation and lateral facies change between lithologies and wavy bedding is a common feature. Individual grainstone lenses contain either cherty of calcareous cement and display early diagenetic zonation defined by jaspilitic cores and chamositic margins. The shale-grainstone facies is capped by a planar jointed Logan diabase sill (Room) which has baked the Gunflint shale at its basal contact.

0.0	Return to Oliver road
0.8	Turn left at Oliver road
1.7	Intersection of Highway 11-17 proceed straight through
2.5	Golf Links Road
3.5	Lakehead University Main Entrance
4.2	Balmoral Avenue
5.1	Hill Street - TURN RIGHT and go 1 block (about 150m) to Queen St.

The corner lot contains a spectacular flat-dipping outcrops of "beach rock". This is on private property and permission must be obtained from the household before examining the exposure.

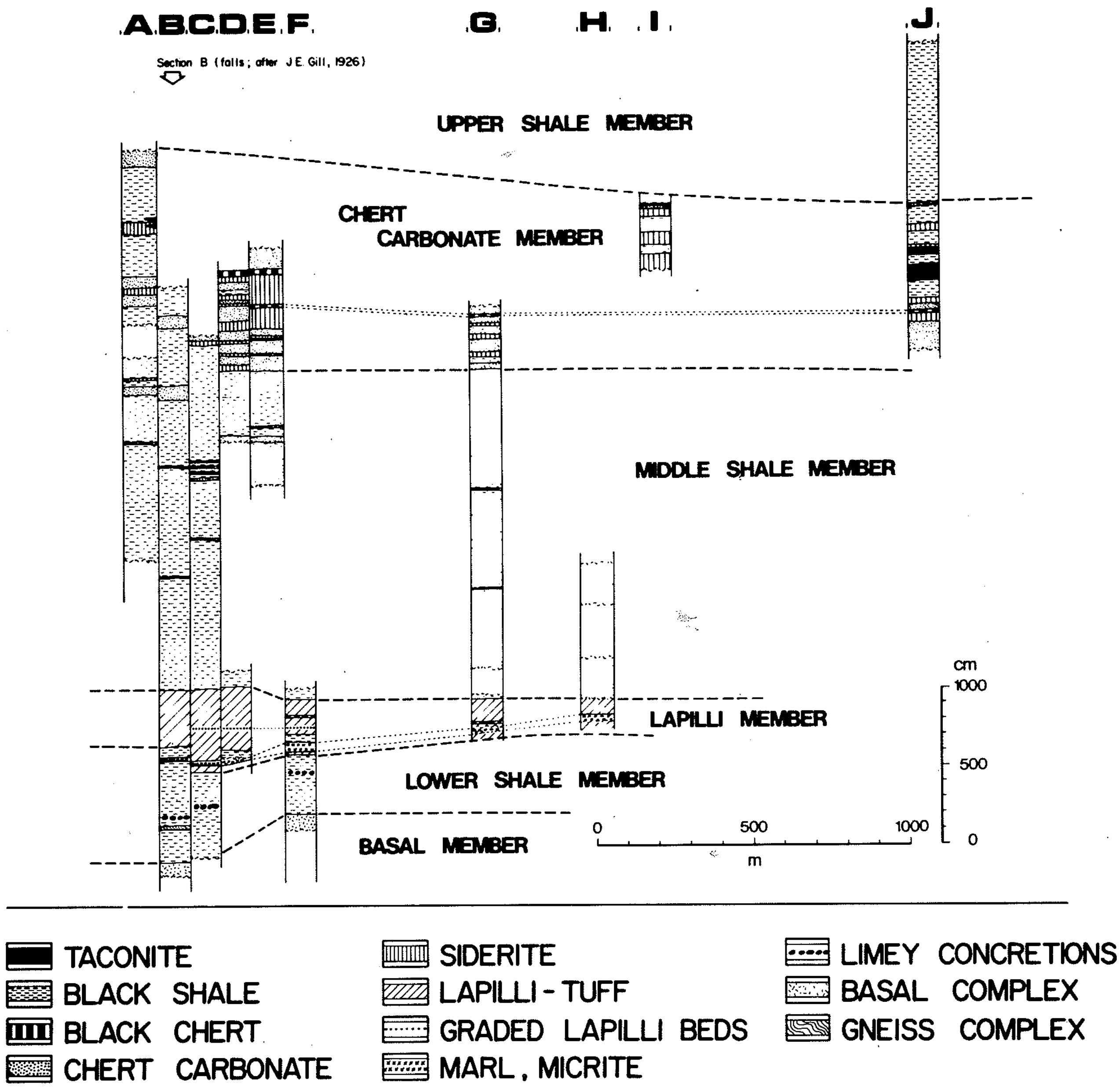


Figure 4 Stratigraphy of the Gunflint Formation, Kaministikwia River gorge.

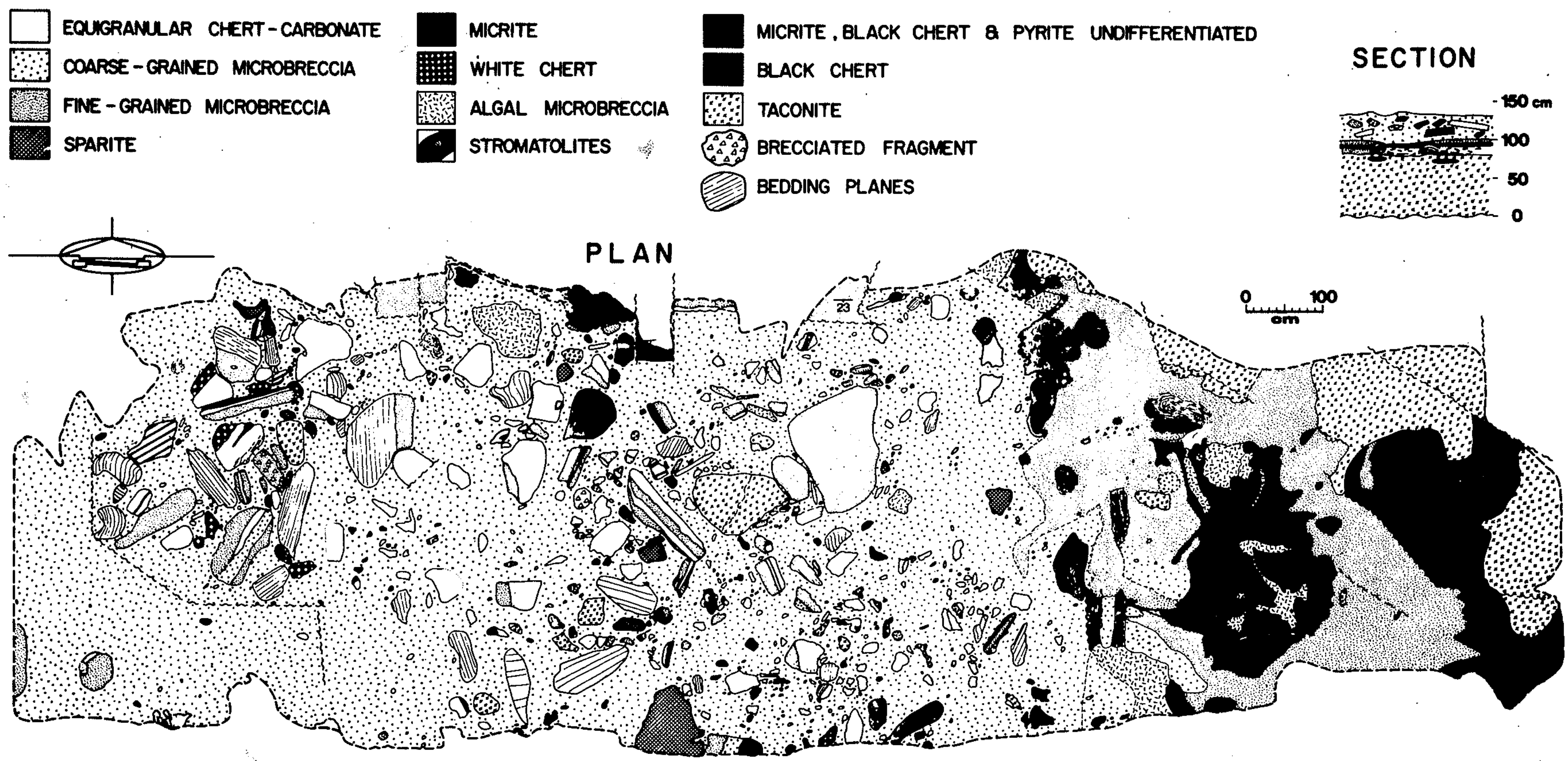


Figure 5 Intraformational breccia in the Gunflint Formation.

STOP 5: BEACH ROCK

This outcrop reveals a planar view of beach rock facies overlying a grainstone base (Fig. 5). In section, calcareous stromatolites, carbonate mudrock and intraformational breccia form a layer, 60 cm thick. The presence of intraclasts greater than 1 meter in diameter in plan view indicates that they are tablet shaped. Intraclasts are angular, range in size from micro-to macro-scale and fall into discrete size ranges. Sorting of the breccias is attributed to strong wind and wave action which probably periodically reached hurricane force. This beach rock exposure contains a rate occurrence of primary calcareous stromatolites and calcite cementation and is similar to the Upper Limestone member, the uppermost unit in the Gunflint Formation. The present elevation of beach rock with respect to the aforementioned member either necessitates severe block faulting (not impossible but improbable) or the occurrence of a lower stratigraphic unit which is similar to the Upper Limestone member. The author favours the latter interpretation based upon regional stratigraphy, particularly in the Centennial Park area which will be Stop 8.

Continue Down Queen Street for 2 Blocks

5.5	High Street - TURN LEFT and proceed north to Hillcrest Park.
5.6	Oliver road
6.3	Entrance to Hillcrest Park

STOP 6: HILLCREST PARK

On a clear day Sibley Peninsula (the Sleeping Giant) can be seen across Thunder Bay. The large cuesta to the southwest is Mount McKay, and is entirely composed of Rove Shale and yet is of similar elevation to the top of the Gunflint Formation. Block faulting may obscure the stratigraphic relations within the Gunflint, as well as between the Gunflint and Rove Formations. The wall built around the flagpole at the lookout is made of beach rock and the large bell rests upon the Upper Limestone member. Proceed down the concrete stairs, turn right and walk along the laneway. The member dips shallowly towards Thunder Bay and consists of calcareous beach rock which overlies a dark ankeritic chert-carbonate facies. Calcareous beach rock contains algal bioherms and basal cryptalgal laminites with fenestrae fabric which are overlain by coarse grained, poorly sorted intraformational breccia. A traverse up-section indicates general fining upwards as can be seen from comparison of this beach rock with finer material at the site of the bell. Note that Gunflint stromatolites, both calcareous and cherty, appear to have thrived in shallow water, high energy environments.

0.0	Leave Hillcrest Park and TURN RIGHT to proceed north along High Street
0.4	Red River Road
1.4	River Street
2.3	4 Way Stop at Margaret Street, TURN LEFT and proceed north along Balsam Street to Huron Ave.
3.1	3 Way Stop at Huron TURN RIGHT and proceed east to Hudson Ave.
3.2	TURN LEFT at Hudson Ave. and proceed to Bridge over Current River. Here Hudson has become Arundel St.
5.4	Bridge over Current River
5.6	TURN RIGHT into Boulevard Lake Park
5.9	Parking area on right at curve in road. This is the north end of Boulevard Lake.

STOP 7: NORTH BOULEVARD LAKE

The outcrop on the north shore of the lake contains gently dipping alternating beds of fine grained ankerite (micrite) and lenticular to wavy bedded cherty grainstone. This is typical grainstone-micrite facies. Crosscutting fractures contain marcasite with anomalously heavy sulphur isotopic ratios (+20 to $-28^{o}/_{oo}$) compared to a mean sulphur isotopic ratio of $+9^{o}/_{oo}$ for syn-sedimentary pyrites. Examination of planar surfaces and vertical exposures between this outcrop and the Boulevard Bridge reveals cycles of alternating grainstone-rich beds and micrite-rich beds. Lenticular bedding, wavy bedding and tabular crossbedding are evident in cross-section along the river bank. Near Boulevard Bridge a greenish carbonate contains planar lamination, intraclasts of chemical mudrock and excellent fenestrae fabric. The carbonate is capped by carbonate domes with nodular texture interpreted to be algal bioherms of an intertidal-supratidal environment. The cyclic occurrence of grainstone overlain by micrite and the algal cap on the uppermost cycle strongly resembles shallowing-upward sequences typical of shallow water carbonate platforms. Grainstone-micrite cycles have been correlated northward for 2.7 km to Trowbridge Falls.

Black shale covers the stromatolite zone of the upper cycle and pyrite is particularly concentrated at this contact. Dome-shaped and undulating beds of pyritic, stromatolitic black chert also occur at this contact and can be traced northward for 900 meters along the east side of Current River to Centennial Park. Sulphur isotopic ratios for 25 samples collected along this horizon are remarkably uniform at $+9^{o}/_{oo}$ ^{34}S. This horizon may represent another time line in the Gunflint, as it was deposited over a large area under uniform conditions. The heavy isotopic ratios for reduced sulphides in this organic-rich environment may indicate that the marine basin was evaporitic and restricted from open circulation with the ocean.

0.0	Return to Arundel St.
0.3	Turn Right at Arundel St.
0.5	Turn left onto Centennial Park road and proceed to Parking Lot
1.6	Parking lot for Centennial Park. Cross Current River on the foot-bridge, climb over the hill on the west side of the river and follow the path to the outcrops along the side of the river.

STOP 8: CENTENNIAL PARK

Riverside exposures consist of beach rock (upstream) and micrite-taconite facies (further downstream). A cross-sectional view of beach rock indicates tilted slabs of rock which forms a pseudo "Teepee" structure. This structure is presumably caused by breaking waves rather than being caused by vertical expansion of evaporite minerals. Well developed planar lamination of the upper flow regime type (Simons et al., 1965) is basal to the beach rock and overlies micrite-grainstone facies. Outcrops downstream contain planar surfaces of micrite and grainstone, synsedimentary sulphide layers and local stromatolite bioherms. This unit resembles the Upper Limestone member, but is overlain by an 11 - meter thick grainstone-shale facies which is partially exposed on the opposite bank of the river. This grainstone-shale facies which is partially exposed on the opposite bank of the river. This grainstone-shale facies has been traced for 1 km along Current River (Fig. 6); primary sedimentary structures within exposures along this section of the river indicate that this facies formed by alternating deposition of mud and chemical grainstone on tidal mudflats and on barrier bars, respectively. a thick black shale member overlies the grainstone-shale facies; the uppermost part of this shale is evident at Trowbridge Falls.

0.0	Return to Arundel Street
1.1	Turn left onto Arundel Street and proceed to Hodder Avenue
2.3	4 Way Stop at Hodder Avenue and the Hodder Avenue Hotel. Turn left and proceed north.
3.9	Intersection of Highway 11-17, turn right on Hwy. 17-11

STOP 9: TERRY FOX LOOKOUT

The typical cuestas of Sibley Peninsula and Pye Island are evident from this height of land. On the north side of the highway, a Logan diabase sill caps a cliff exposure of black shale (Rove Formation) which is underlain by the Upper Limestone member. This Upper Limestone member is a fining upward sequence of beach rock. The shard-like fragments in carbonate mudrock matrix near the top of the member are interpreted to be curled mudchips which were lifted from the desiccated intertidal mudflat and redeposited along these bedding planes. Similar mudchip breccias occur in the overlying Kama Hill Formation of the Sibley Group. (Franklin et al., 1980)

The Rove Formation consists of fissile black shale which is macroscopically indistinguishable from shale members in the Gunflint Formation. Further to the southwest the Rove shale gradually changes upwards into turbidites (Morey, 1972). The Rove Formation is therefore considered by the author to be more of a lateral, deep-basin, siliciclastic facies equivalent to the Gunflint as well as an overlying formation. According to this interpretation, periodic siliciclastic deposition of black (Rove) shale combined with (bio)chemical precipitation of (Gunflint) chert-carbonates best characterizes the Animikie Group in the Kaministikwia River-Current River Area.

SUMMARY

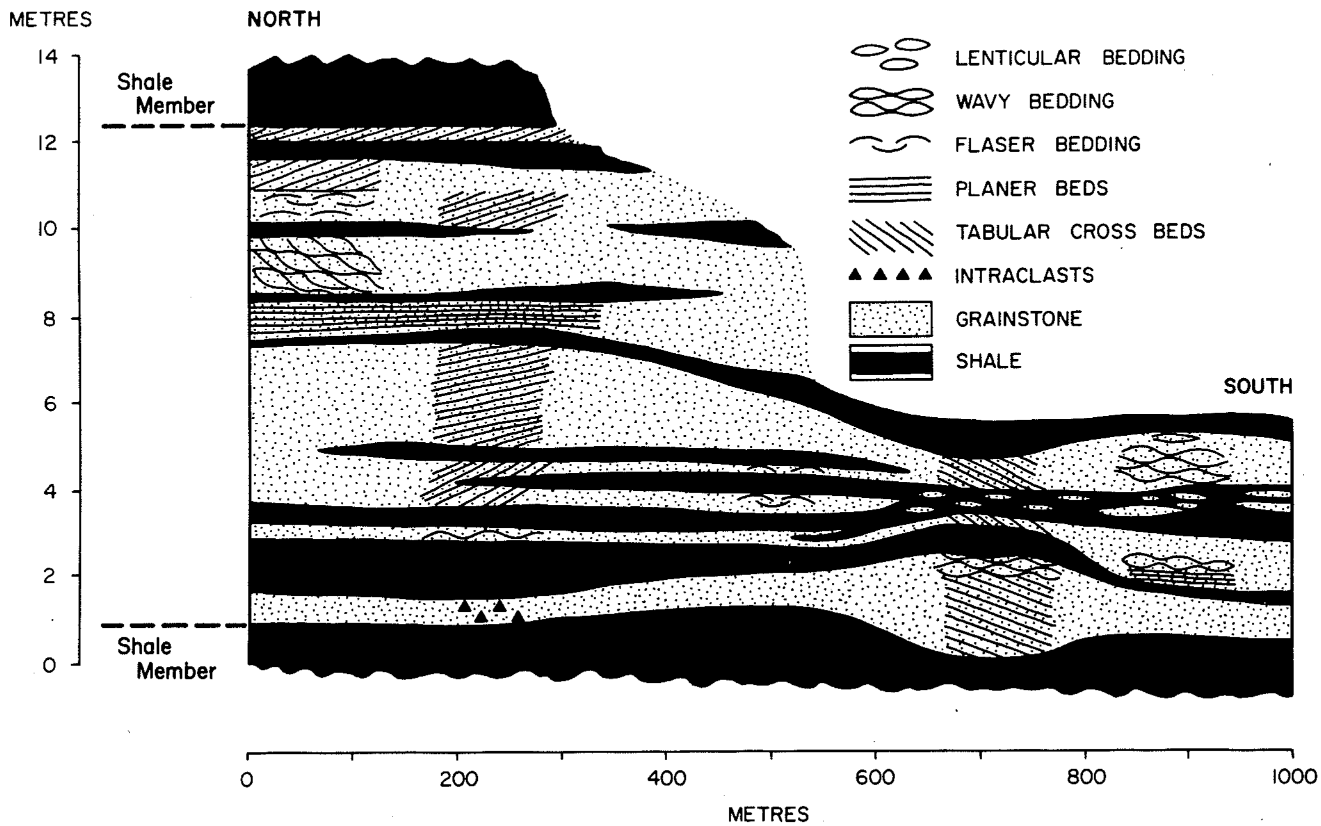


Figure 6 Grainstone-shale in the Current River section.

Attempts at lateral correlation for 27 kilometers along the shelf margin (Fig. 7) demonstrate the lenticular geometry of members on a regional scale. Lack of lateral continuity complicates stratigraphic nomenclature, hence, members are presently defined on the basis of their facies association. Grainstone-micrite members formed in areas of the shallow carbonate shelf and were possible barrier islands that were the sites of actively precipitating chert and/or carbonate and therefore escaped siliciclastic dilution. They probably represent barrier island complexes which migrated parallel to the shoreline. Shale members formed in lagoonal areas and mudflats along the shelf within shallow depressions. They received sufficient siliciclastic muds from river mouths and from longshore drift to gradually accumulate thick sequences of shallow water calcareous mudrock. Grainstone-shale members represent areas in which chemical precipitation and siliciclastic deposition occurred simultaneously and in which these two types of sediments were subsequently reworked and redistributed by tidal currents. The only other influences affecting sedimentation were a brief period of weathering on the Archean peneplain which produced the basal Kakabeka conglomerate and minor explosive intermediate volcanism which deposited lapilli-tuff in the lower Gunflint Formation.

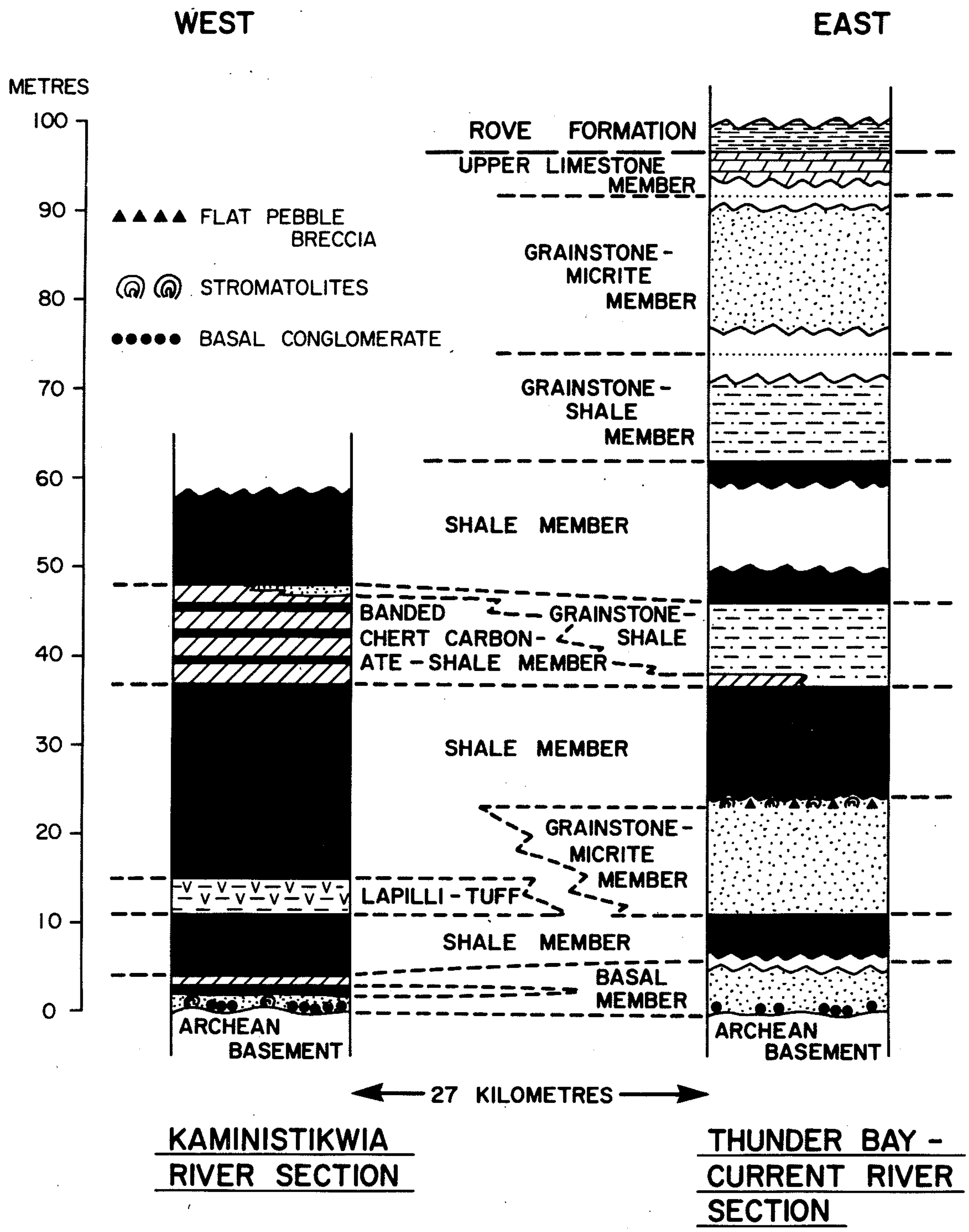


Figure 7 Generalized stratigraphy and lateral correlation of the Gunflint Formation, Kaministikwia River to Current River.

DESCRIPTION OF STOPS AND ITINERARY FOR THE SIBLEY GROUP

J.M. Franklin[1] and W.H. McIlwaine[2]

Distance measurements begin at the junction of Highways 11-17 and 61 near the Thunder Bay airport. All Stop locations are shown figure 8.

KM	Description
0.0	Proceed north along Highway 11-17
3.1	Harbour Expressway
4.7	Oliver road (Highway 130)
7.9	John Street
9.0	Red River Road (Highway 102)
11.4	Balsam Street
15.6	Hodder Avenue
16.6	Terry Fox Scenic Lookout
19.5	Spruce River Road (Highway 527 - formerly 800) to Armstrong
34.8	Lakeshore Drive
50.2	Highway 587 to Sibley Peninsula, Sibley Provincial Park and Silver Islet; turn right and proceed

STOP 1: PASS LAKE FORMATION

1A: This cliff is the type section for the Pass Lake Formation. Exposure is almost continuous for about 3.2 km along the tracks and gives a stratigraphic thickness of 50 meters.

At the western end of this outcrop, a quarry provides an excellent exposure of Sibley quartz arenite (Cheadle's Fork Bay Member). In the railway cut at the western edge of the quarry, Rove shale is altered to a reddish colour. This oxidative alteration affected the Rove for several meters below its contact with the Sibley Group. A thin lens of basal conglomerate is exposed here (Cheadle's Loon Member); a better exposure is available to the northeast, and is described below. The quartz arenite exposed there is thick bedded, very well sorted and poorly graded. Isolated boulders occur within the basal beds. Ripple marks are present, and most easily seen on broken slabs. A few low angle cross beds may also be observed.

1B

56.8 The conglomerate is better exposed behind the railway shed at the east end of Pass Lake.

Clasts in the basal polymictic conglomerate are composed of 93% Gunflint iron-formation, 6% quartz and 1% granite. Boulders are of variable size and angularity, and are cemented by a sandy matrix. The contact with overlying sandstone is sharp; only a few pebbles are found in the base of the overlying unit. The sandstone is moderately to poorly indurated, thick bedded at the bottom of the section, and composed of quartz, with minor chert and feldspar, in a calcite matrix.
Continue along Highway 587

57.5 Pass Lake East road: turn left and proceed east.

60.5	No. 4 Road: turn left and proceed north
61.0	Y Junction - bear left
62.1	Field on left side (west) of road: park here and proceed on foot across field and through bush to cliff exposure along the CNR tracks - about 400 meters.

[1] Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario. K1A 0E8

[2] Ministry of Northern Development and Mines, Box 5000, 435 S. James Street, Thunder Bay, Ontario. P7C 5G6

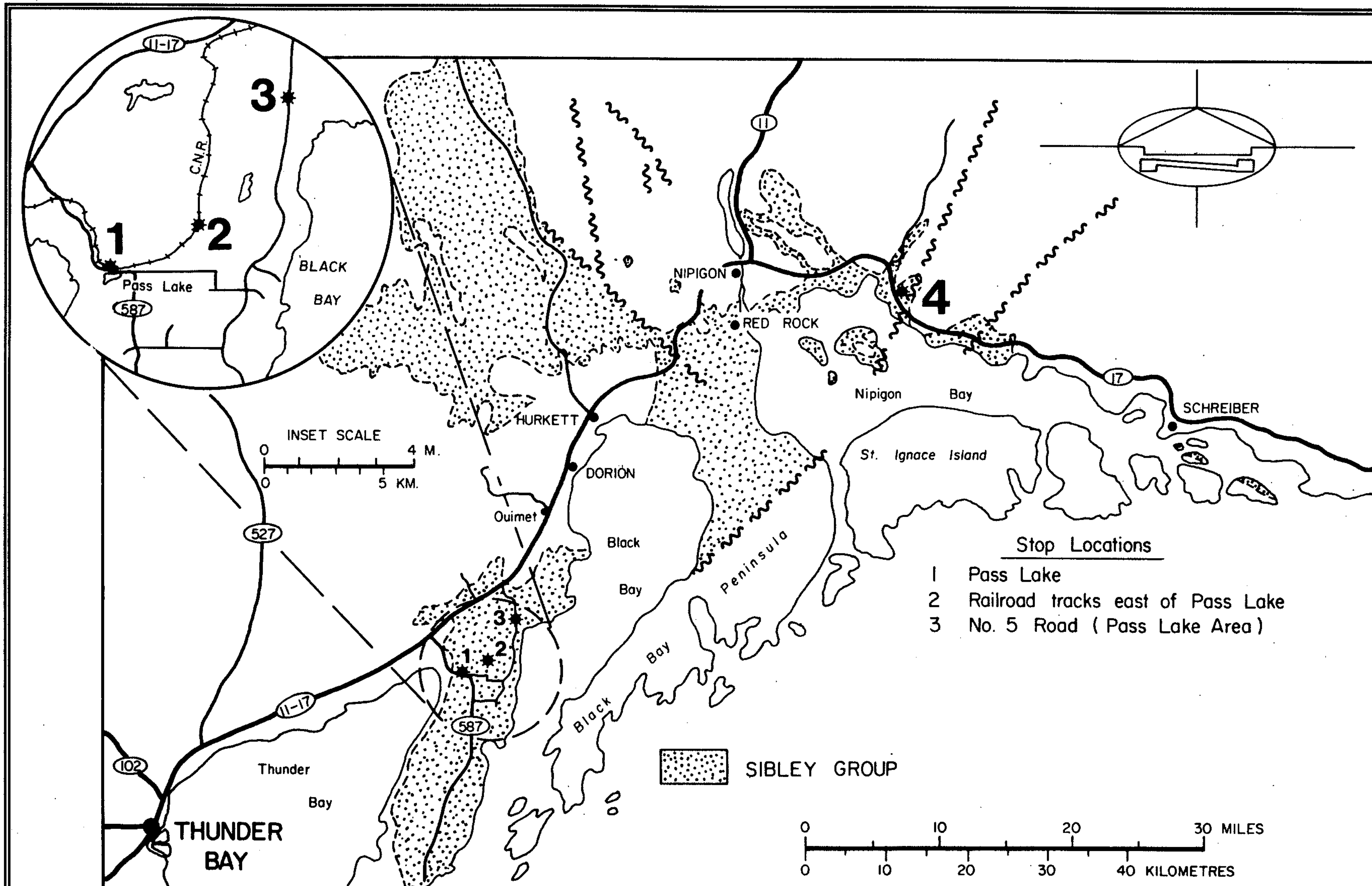


Figure 8 Stop locations and distribution of the Sibley Group, Ontario.

STOP 2: INTRAFORMATIONAL BRECCIA CUTTING ROSSPORT FORMATION

This is an excellent exposure of intraformational breccia cutting the Rossport Formation, containing clasts of both Rossport Formation, including chert from its central member, and Kama Hill Formation. Clasts range from 1 cm to 40 cm. The breccia zone has sharp, near vertical walls and is cut by sandstone dykes. Although the fragments have virtually all been derived from the stratigraphically higher sections of the Sibley Group, the sandstone dykes may have been derived from below.

Return to No. 4 Road and continue north

64.2 Gravel Pit - for turning vehicle around. Proceed back to Pass Lake East Road.

68.1 Pass Lake East Road - turn left.

Proceed along Pass Lake East Road from Junction with No. 4 road.

69.4 Right angle bend to left
70.5 No. 5 road - turn left

STOP 3: KAMA HILL SANDSTONE

This is a series of small outcrops of reddish-brown to reddish-purple sandstone of the Kama Hill Formation. It is fine to medium-grained, and although not evident at this location, ripple marks and some mudcracks are common in this unit. These occur along the shore of Lake Superior to the east (McIlwaine, 1971b). The main difference between this sandstone and sandstone of the Rossport Formation is the lack of carbonate in the Kama Hill Formation. The Kama Hill sandstone is a local facies of this formation. More generally the unit is composed of poorly fissile microlaminated mudstone. The sandy facies occurs only on the Sibley Peninsula, and probably indicates proximity to a source area and shoreline.

Continue North along No. 5 Road

84.0 Cross CNR Tracks
84.7 Highway 11-17 - Turn Right
84.7 Proceed northeast along Highway 11-17
97.2 Road to Ouimet Canyon
106.2 Bridge over Wolf River
109.5 Road to Stewart Lake
130.0 Highway 627 - proceed straight
131.4 Highway plaque describing Red Rock Cuesta. The Cuesta features a thick sequence of red sand limestone (lower member of the Rossport Formation) capped by a diabase sill.
133.5 Road cut here shows a diabase sill cutting Archean rocks and Sibley Group
138.2 Road to Moseau Mountain - Proceed through Nipigon on Highway 11-17
142.7 Junction of Highways 11 and 17. Proceed straight, along Highway 17
161.5 Junction of Domtar Road
163.8 First Lookout, Kama Hill

STOP 4: ROSSPORT FORMATION OVERLAIN BY KAMA HILL FORMATION WITH A CAPPING OF DIABASE

A broad anticline of sandy red carbonate is exposed in the prominent road cut to the north of this lookout. This represents the lowest member of the Rossport Formation, Cheadle's Channel Island Member. It contains rounded isolated quartz clasts set in a carbonate matrix. Clasts increase in abundance towards both the top and the base of the member; its uppermost part consists of interbedded quartz arenite and sandy dolomite, best seen below the chert-carbonate of the central member at the next part of this stop (4B) described below.

Soft-sediment deformation produced the anticlinal structure. Three thin diabase sheets follow bedding planes; the sills pinch out, and locally cut across bedding at a high angle.

Follow Highway 17 (south) to the distinctive red and white interbedded sandy dolostone. Sandstone beds (white) are lenticular in shape, and are interbedded with red, sandy dolostone; this area represents the uppermost part of the Channel Island Member.

164.2 Second Lookout, Kama Hill (Stop 4B).

In the roadcut to the north of the second lookout, the following features can be observed:

(1) Two thin Keweenawan diabase sills, partially replaced by carbonate, cut across the poorly developed bedding planes at a low angle.

(2) Finely laminated chert-carbonate of the chert-stromatolite unit (central member of the Rossport Formation; Cheadle's Middlebrun Bay Member) is exposed below the lower sill. Up to 15 cm of anthraxolitic carbonate has accumulated at the base of the chert. An oily smell may be detected when this anthraxolite is freshly broken. The remainder of the unit is composed of thinly interlaminated chert and carbonate. This distinctive unit is the key marker horizon in the Sibley Group, as it has been traced from Channel Island near Rossport, to near Lake Marie Louise on the Sibley Peninsula, near the Wolf River northwest of Dorion, and was intersected in drill core near Wolfpup Lake, 27 km north of Pearl, Ontario. No microfossils have been detected in it, but the characteristic thin laminations suggest a cryptalgal or stromatolitic origin (Cheadle, 1986).

(3) Red dolostone above this unit is the upper member of the Rossport Formation, Cheadle's Fire Hill Member. It is marked by many cream-coloured spherical spots (average diameter 10mm). Similar spots are evident throughout this unit, and commonly have a small amount of graphite or hydrocarbon at the centre. In thin section, the only apparent mineralogical change in the spots is the lack of hematite coatings on clay and carbonate grains. The upper member has very little clastic content, and contains authigenic K-feldspar, illite, and some expandable clay, in addition to calcite and dolomite; in drill core, it contains anhydrite.

(4) Irregular, flame-shaped, bleached zones follow structures and bedding plane cleavage in the red dolostone. Diagenetic leaching of hematite and destruction of clay minerals and feldspar has occurred along the fractures.

(5) Above the road cut and overlying the talus slope, the Kama Hill Formation is exposed. It is somewhat fissile, and contains approximately 4% hematite, which coats very fine grained expandable clay and microcline, and forms blades of specularite in tiny vugs. Bleaching along fractures is common in this rock. Purple mudstone contains abundant synaeresis cracks, and to the west, at Stewart Lake, contains thin stromatolite beds. The mudstone is finely laminated (visible only in thin section) and contains lenticular zones of mud-pellet micro-conglomerate and mud-chip breccia. A few gypsum casts have been found in this outcrop, as well as a few raindrop prints. These attest to the subaerial origin of this unit.

The expandable clay mineral consists of regularly interstratified vermiculite, chlorite and montmorillonite. This clay, together with microcline and minor illite, are probably authigenic minerals, and constitute the principal minerals of some laminae. Fine-grained clastic quartz is an important constituent of silty beds.

The type section of the Kama Hill Formation is at the top of Kama Hill. The best access is provided by following the Domtar Logging Road (1.3 km west of the first lookout at Kama Hill) for 0.5 km to the first power line. The section is at the top of the hill and is exposed on the power line. This site cannot be visited during this trip, due to time constraints.

AMETHYST IN THE THUNDER BAY AREA

M.C. Smyk and M.S. O'Brien
Resident Geologist's Office,
Ministry of Northern Development and Mines,
Box 5000, 435 S. James St.,
Thunder Bay, Ont.
P7C 5G6

The occurrence of amethyst in the Thunder Bay area has been known since the early 19th century (Logan 1846). Local amethyst deposits have most recently been described by Vos (1976) and Patterson (1985). Amethyst occurs in both Archean and Proterozoic rocks in an area extending from Thunder Bay, eastwards to near Rossport along the north shore of Lake Superior (Fig 9a, b).

Amethyst occurs in veins and vein breccias that occupy fault zones proximal, and perhaps related, to the margin of the two structural provinces. Many of the amethyst veins contain base metals and silver, and appear to be part of a spectrum of local vein types (described below) which include lead-zinc-barite veins (Franklin and Mitchell, 1977) and silver veins (Franklin et al., 1986). The best quality and most productive amethyst deposits occur in veins which are at or immediately below the unconformity of the Sibley Group with Archean granitic rocks. A few occurrences are in the lower part of the Rossport Formation. Most of the veins are zoned and vuggy, and were formed through deposition from low-temperature fluids which migrated along the unconformity, and precipitated because of rapid fluid expansion. Fluid inclusion studies on the closely related lead-zinc-barite veins, together with S-isotope determinations on these veins (Franklin and Mitchell, 1977), indicate that they formed at about 100°C. Lead isotope studies indicate that some of the vein constituents were locally derived, and that the fluids were poorly mixed, and probably highly saline.

The deposits in McTavish Township, including the Pearl Lake Mine, are typical. Franklin and Mitchell (1977) and Patterson (1985) suggested that the veins originated from dewatering of the Proterozoic basinal rocks. Geo-pressured fluids were driven by a major thermal anomaly in the Mid - Continent (Keweenawan) Rift, and moved up along basin-margin faults. Franklin and Mitchell (1977) suggested that Archean faults at the unconformity may have been reactivated; multi-stage dilation and brecciation is common in fault-hosted veins and vein breccias.

STOP 5: PEARL LAKE AMETHYST MINE

The local geology (Fig. 10) has been mapped by McIlwaine (1971) and described by Patterson (1985). The mine property is underlain by Sibley Group basal conglomerate and mudstone resting unconformably on Archean biotite-quartz monzonite that contains xenoliths of metasedimentary rocks. Generally, the granitic rocks form resistant hills and rocky knolls, while the flat-lying sedimentary rocks occur in the valleys. Local amethyst-bearing veins and vein breccias occur along a strike length of almost one kilometer and are spatially associated with a persistent fault trending 035°, and related splays. The original discovery was made by tracing amethyst-bearing granitic boulders to their bedrock source.

STOP 5A: BEAVER POND NARROWS

A 3m - wide quartz breccia zone trending 035° is exposed along the (faulted?) unconformable contact between Archean granitic rocks on the west and basal conglomerates of the Proterozoic Sibley Group on the east side of the ravine. The granitic rocks are red, massive, coarse-grained and equigranular, and consist of approximately 70% red K-feldspar, 25% yellowish plagioclase and 5% quartz. There are two conglomeratic units exposed on the east face. The upper unit consists of 70% to 75% well-rounded granitic clasts in a white, sandy matrix. The lower unit contains reworked clasts of Sibley Group (Pass Lake Formation) quartz arenite (30%), rounded clasts of Archean granite, and Animikie

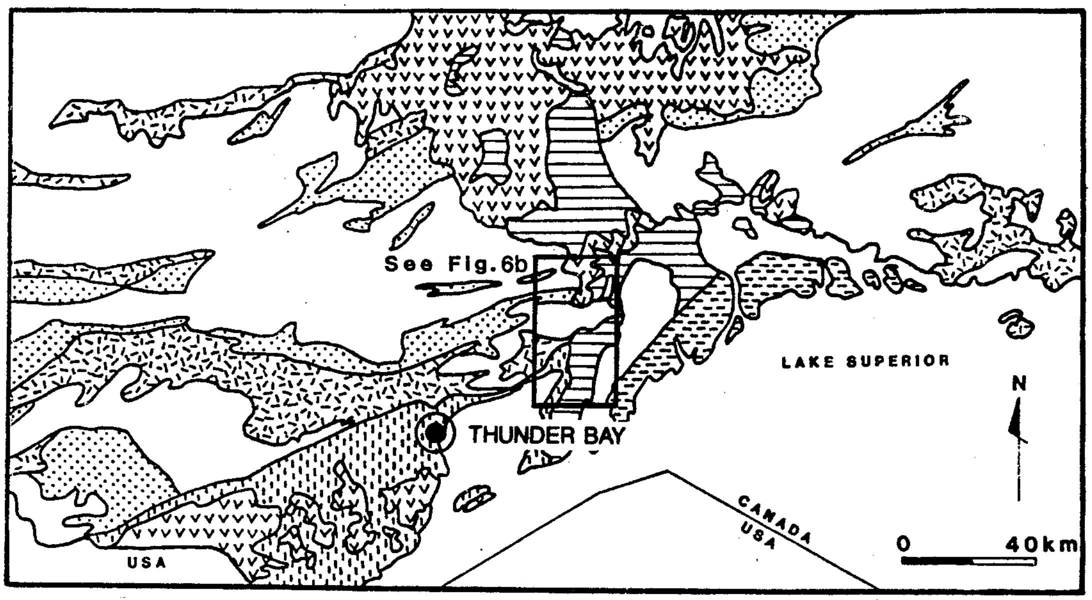


Figure 9a General geology of the Thunder Bay Area

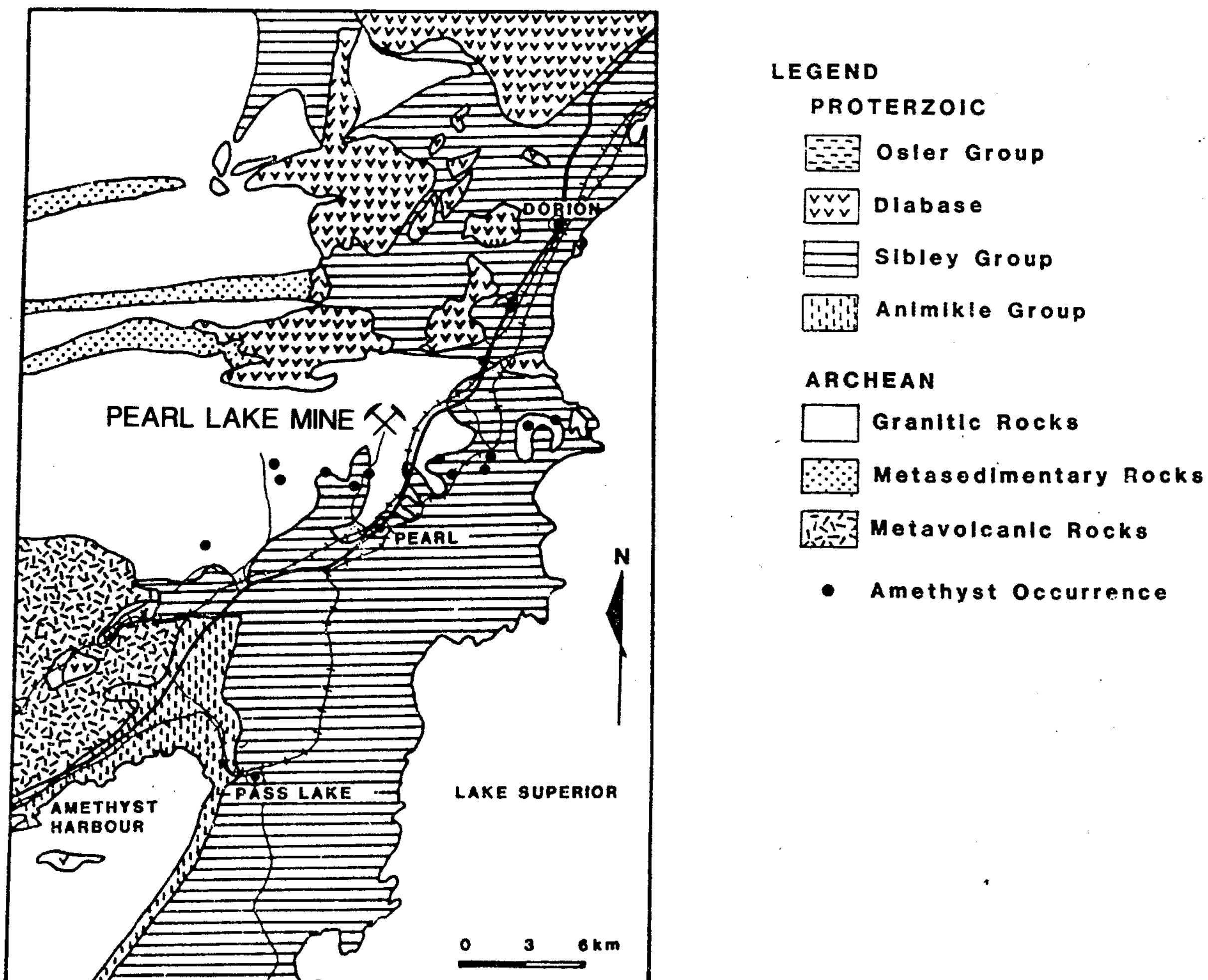


Figure 9b Amethyst occurences near Thunder Bay (after Patterson 1985).

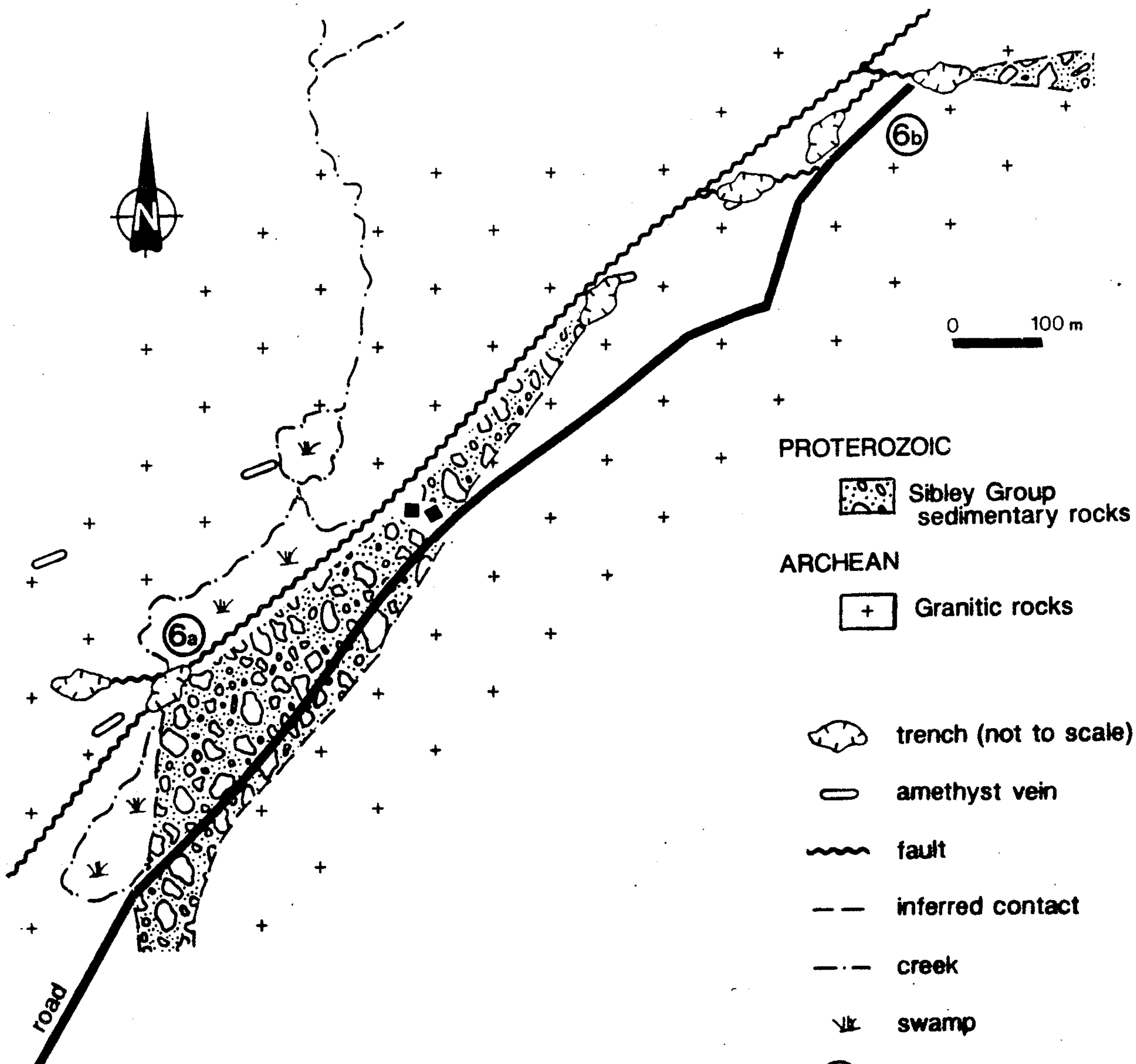


Figure 10 Geological sketch map of the Pearl Lake Amethyst Mine

Group (Gun flint Formation) taconite (20% to 30%), in a reddish sandstone matrix. The sharp contact between the two units dips shallowly to the east. The breccia zone has been pitted and excavated for mining purposes. It consists of 10% to 20% white quartz, 5% to 10% vugs up to 60 cm in size, lined with hematite-coated, pale to dark purple amethyst, and 70% to 85% silicified clasts of conglomerate and granite.

Along the trail leading from the gift shop to Stop 5a, an outcrop displays the unconformity. At this site an anastomosing quartz vein $\pm$ breccia system occurs at the (faulted?) contact between Sibley conglomerate and Archean granite.

STOP 5B: NORTH BRECCIA ZONE

The northeastern extension of the main amethyst-bearing zone here consists of a number of granite-hosted veins and vein breccias containing silicified granite fragments. The attitude of the vein-bearing structures ranges from approximately 035^{o} to 090^{o}. The veins and breccias are exposed in a number of open cuts, stripped areas and trenches. In vugs, amethyst occurs as crystals typically up to 1.5 cm in size, ranging in colour from green and smoky black, to pale to dark purple. White, pink and light brown barite is a major accessory mineral that commonly cements brecciated fragments. Minor pyrite, sphalerite and galena also occur locally.

REFERENCES

Anglin, C.D., Franklin, J.M., Loveridge, W.D., Hunt, P.A., and Osterberg, S.A., 1988: Use of zircon U-Pb ages of felsic intrusive and extrusive rocks in eastern Wabigoon Subprovince, Ontario, to place constraints on base metal and gold mineralization; in Ragiogenic age and isotopic studies: Report 2, Geological Survey of Canada, Paper 88-2, p.109-115.

Armbrust, G., 1969: Hydrothermal alteration of a breccia pipe deposit, Tribag mine, Batchawana Bay, Ontario; Economic Geology, v. 64, p. 551-563.

Bakker, F.J., 1979: Petrology and Geochemistry of the Quartz Pebble Conglomerate of the Geco Massive Sulphide Deposit at Manitouwadge, Ontario; B.Sc. Thesis, McMaster University, Hamilton, Ontario.

Balint, F. and Salm, E.C., 1983: Brief Description of Winston Lake Stratigraphy; internal Corporation Falconbridge Copper report.

Balint, F., 1982: Proposed Diamond Drilling-Zenmac Option; internal Corporation Falconbridge Copper report.

Barghoorn, Elso S., Tyler, Stanley, A. 1965: Microorganisms of Middle Precambrian Age from the Animikie Series, Ontario, Canada; in Current aspects of exobiology, California Institute of Technology, Jet Propulsion Laboratory, Pasadena, Chapter 3, p. 93-118.

Barghoorn, E.S. 1971: The Oldest Fossils; Scientific American, v. 224, No. 5, p. 30-42.

Bartley, M.W. and Page T.W., 1957: Unpublished Geological Report on the Hemlo area, Thunder Bay District, Ontario, for Canadian Pacific Railway Limited; Ministry of Northern Development and Mines, Resident Geologist's Files, Thunder Bay, 17p.

Bartley, M.W., 1940: Geology of the Big Duck-Aguasabon Lakes Area; in Forty-Ninth Annual Report of the Ontario Department of Mines, v. XLIX, part VII, llp.

Bell, A.M., 1981: Vergence: an evaluation; Journal of Structural Geology, v. 3, p. 197-202.

Bitler, K.A., 1988: Who Owns Hemlo: World Investment News, February 1988, p. 14-16.

Blecha, M., 1974: Batchawana area, a possible Precambrian porphyry copper district; Canadian Institute of Mining and Metallurgy, Transactions, v. 77, p. 393-398.

Brown, P., Carson, D., Cooper, P., Farr, E., Gibson, J., Harvey, J., Hogg, W., MacIsaac, N., Mackie, B., Nash, W., and Pierce, G., 1985: Golden Giant joint venture properties-Hemlo; in R.H. McMillan and D.J. Robinson,(eds), Gold and Copper-zinc Metallogeny, Hemlo-Manitouwadge-Winston Lake, Ontario, Canada; Geological Association of Canada and Canadian Institute of Mining and Metallurgy, A Compendium, p.58-65.

Brown, A.C., 1971: Zoning in the White Pine copper deposit, Ontonagon County, Michigan; Economic Geology, v. 66, p. 543-573.

Brown, W.L. and Bray, R.C.E., 1960: The Geology of the Geco Mine; Canadian Mining and Metallurgical Bulletin, v. 58, p. 3-11.

Buchan, R., 1983a: Mineralogical Examination of Thirty Sulphide Samples from Winston Lake Project; internal Falconbridge Metallurgical Laboratories Report.

Buchan, R., 1983b: Mineralogical Examination of Host Rocks from Winston Lake Project; internal Falconbridge Metallurgical Laboratories Report.

Burk, R., Hodgson, C.J., and Quartermain, R.A., 1986: The Geological Setting of the Teck-Corona Au-Mo-Ba Deposit, Hemlo, Ontario, Canada; In A.J. MacDonald (ed), Proceedings of Gold'86, an International Symposium on the Geology of Gold, Toronto, Ontario, p. 311-326.

Burnie, S.W., 1970: Sulfur isotopes in the White Pine mine, Ontonagon County, Michigan; Unpublished Miscellaneous thesis, McMaster University, 132 p.

Busby-Spera, C., 1984: Large volume rhyolite ash flow tuff eruptions and submarine caldera collapse in the lower Mesozoic, Sierra Nevada, California; Journal of Geophysical Research, v. 89, p. 8417-8427.

Cannon, W.F. and Klasner, J.S., 1976: Phosphorite and other apatite-bearing sedimentary rocks in the Precambrian of northern Michigan; United States Geological Survey Circular 746, 6 p.

Card, K.D., Poulsen, K.H., and Franklin, J.M., 1990: Regional Metallogeny of Superior Province; in Precambrian Metallogeny related to plate tectonics, Precambrian Research, in press.

Card, K.D. and Ciesielski, A., 1986: DNAG No. 1. Subdivisions of the Superior Province of the Canadian Shield; Geoscience Canada, v. 13, p. 5-13.

Cas, R.A.F. and Wright, J.V., 1987: Volcanic Successions, Modern and Ancient; Allen and Unwin Publishing, London, 528 p.

Cheadle, B.A., 1986: Alluvial-playa sedimentation in the lwere Keweenawan Sibley Group, Thunder Bay District, Ontario; Canadian Journal of Earth Sciences, v. 23, p. 527-542.

Cheadle, B.A. 1981: The stratigraphy and depositional history of the Rossport Formation stratotype, Channel Island, Thunder Bay district, Ontario; Unpublished B.Sc. thesis, Lakehead University, Thunder Bay, 46 p.

Corfu, F. and Muir, T.L., 1989: The Hemlo-Heron Bay Greenstone Belt and Hemlo Au-Mo Deposit, Superior Province, Ontario, Canada: 1. Sequence of Igneous Activity Determined by Zircon U-Pb Geochronology; Chemical Geology (Isotope Geoscience Section), v. 79, p. 183-200.

Craddock, C. 1972: Late Precambrian regional geological setting in Geology of Minnesota; in P.R. Sims and G.B. Morey, (eds), a Centennial volume, Minnesota Geological Survey, University of Minnesota, St. Paul, p. 281-291.

Dahl, R., McGoran, J., and Watkinson, D.H., 1986: The Coldwell Complex Platinum-Group Element Deposit: 1. Geological Relationships of Layered Gabbroic Host Rocks; in Program with Abstracts, Geological Association of Canada-Mineralogical Association of Canada-Canadian Geophysical Union, Joint Annual Meeting, Ottawa, May, 1986, v. 11, p. 61.

Dahl, R., Watkinson, D.H., and McGoran, J.W., 1987a: Two Duck Lake Intrusion, Coldwell Alkaline Complex, Ontario: 1. Geology and Structure; in Program with Abstracts, Institute on Lake Superior Geology, 33rd Annual Meeting, Wawa, Ontario, May, 1987.

Dahl, R., Watkinson, D.H., and McGoran, J.W., 1987b: Two Duck Lake Intrusion, Coldwell Alkaline Complex, Ontario: 2. Petrology and Base-Metal/ PGE Geochemistry; in Program with Abstracts, Institute on Lake Superior Geology, 33rd Annual Meeting, Wawa, Ontario, May, 1987.

Davis, D., Krogh, T.E., Hinzer, J., and Nakamura, E., 1985: Zircon dating of polycyclic volcanism at Sturgeon Lake and implications for base metal mineralization; Economic Geology, v. 80, p. 1942-1952.

Davis, D.W. and Jackson, M.C., 1988: Geochronology of the Lumby Lake Greenstone Belt: A 3 Ga complex within the Wabigoon Subprovince, Northwestern Ontario; Geological Association of America Bulletin, v. 100, p. 818-824.

Davis, D.W. and Sutcliffe, R.H., 1985: U-Pb zircon ages from the Nipigon Plate and Northern Lake Superior, Geological Association of America Bulletin, v. 96, p. 1572-1579.

Davis, D.W., Sutcliffe, R.H., and Trowell, N.F., 1988: Geochronological constraints on the tectonic evolution of a late Archean greenstone belt, Wabigoon Subprovince, northwest Ontario, Canada; Precambrian Research, v. 39, no. 3, p. 171-191.

Davis, D.W. and Trowell, N.F., 1982: U-Pb zircon ages from the eastern Savant Lake-Crow Lake metavolcnicmetasedimentary belt, northwest Ontario; Canadian Journal of Earth Sciences, v. 19, p. 868-877.

Dube, B., Koopman, E.R., Franklin, J.M., Poulsen, K.H., and Patterson, M.R., 1989: Preliminary study of the stratigraphic and structural controls of the Lyon Lake massive sulphide deposit, Wabigoon Subprovince, northwest Ontario; in Current Research, Part C, Geological Survey of Canada, Paper 89-1C, p. 275-284.

Edhorn, Anna-Stina 1973: Further investigations of fossils from the Animikie, Thunder Bay, Ontario; Proceedings Geological Association of Canada, v. 25, p. 37-66.

Floran, R.J. and Papike, J.J. 1975: Petrology of the Low-Grade Rocks of the Gunflint Iron Formation, Ontario-Minnesota; Geological Society of America Bulletin, v. 86, p. 1169-1190.

Fowler, A.D., Jensen, L.S., and Peloquin, S.A., 1987: Varioles in Archean Basalts: Products of Spherulitic Crystallization; Canadian Mineralogist, v. 25, p. 275-289.

Fralick, P.W., Barrett, T.J., Jarvis, K.E., Jarvis, I., Schnieders, B.R., and Vande Kemp, R., 1989: Sulfide-Facies Iron Formation at the Archean Morley Occurrence, Northwestern Ontario: Contrasts with Oceanic Hydrothermal Deposits; Canadian Mineralogist, v. 27, p. 601-616.

Franklin, J.M., and Mitchell, R.H., 1977: Lead-Zinc-Barite Veins of the Dorion Area, Thunder Bay District, Ontario; Canadian Journal of Earth Sciences, v. 14, p. 1963-1979.

Franklin, J.M., Gibb, W., Poulsen, K.H., and Severin, P., 1977: Archean metallogeny and stratigraphy of the South Sturgeon Lake area; Institute of Lake Superior Geology, 23rd annual meeting, Mattabi field trip guidebook 75 p.

Franklin. J.M. 1978a: Rubidium-strontium age determination of the Rove Formation shale; in R.K. Wanless and W.D. Loveridge (eds), Rubidium-strontium isochron age studies, report 2, Geological Survey of Canada Paper 77-14, p. 35-39.

Franklin, J.M. 1978b: The Sibley Group, Ontario, in R.K. Wanless and W.D. Loveridge (eds), Rubidium-strontium isochron age studies, report 2, Geological Survey of Canada Paper 77-14, p. 31-34.

Franklin, J.M., McIlwaine, W.H., Poulsen, K. H., and Wanless, R.K. 1980: Stratigraphy and depositional setting of the Sibley Group, Thunder Bay district, Ontario, Canada; Canadian Journal of Earth Sciences v. 17, No. 5, p. 633-651.

Franklin, J.M., Lydon, J.W. and Sangster, D.F., 1981: Volcanic-associated massive sulfide deposits; Economic Geology, 75th Anniversary Volume, p. 485-627.

Franklin, J.M., Mcllwaine, W.H., Shegelski, R.J., Mitchell, R.H., and Platt, R.G., 1982: Proterozoic geology of the Northern Lake Superior area; in J.M. Franklin (ed), Geological Association of Canada- Mineralogical Association of Canada, Field Trip Guidebook Trip 4, 71 p.

Franklin, J.M., Kissin, S.A., Smyk, M.C., and Scott, S.D., 1986: Silver Deposits Associated with the Proterozoic Rocks of the Thunder Bay District, Ontario; Canadian Journal of Earth Sciences, v. 23, p. 1576-1591.

Franklin, J.M. and Thorpe, R.I., 1982: Comparative metallogeny of the Superior, Slave and Churchill Provinces; in R.W. Hutchinson, C.D. Spence, and J.M. Franklin, (eds), Precambrian sulphide deposits, H.S. Robinson Memorial Volume, Geological Association of Canada, Special Paper Number 25, p. 3-90.

Friesen, R.G., Pierce, G.A., and Weeks, R.M., 1982: Geology of the Geco Base Metal Deposit; Geological Association of Canada, Special Paper 25, p. 343-363.

Goodwin, A.M. 1956: Facies Relations in the Gunflint Iron Formation; Economic Geology, v. 51, p. 565-595.

Goodwin, A.M. 1960: Gunflint Iron Formation of the Whitefish Lake Area; Ontario Department of Mines, v. 69, part 7, p. 41-63.

Gross, G.A., 1965: Geology of iron deposits in Canada. 1. General geology and evaluation of iron deposits; Geological Survey of Canada, Economic Geology Report 22, 181 p.

Groves, D.A., 1984: Stratigraphy and alteration of the footwall rocks beneath the Archean Mattabi massive sulphide deposit, Sturgeon Lake, Ontario; Unpublished M. Sc. Thesis, University of Minnesota-Duluth, Duluth, MN, 115 p.

Groves, D. A., Morton, R. L., and Franklin, J. M., 1988, Physical volcanology of the footwall rocks near the Mattabi massive sulphide deposit, Sturgeon Lake, Ontario: Canadian Journal of Earth Sciences, v. 25, p. 280-291.

Guel, J.J.C. 1970: Geology of Devon and Pardee Townships and the Stuart Location; Ontario Department of Mines and Northern Affairs, GR 87, 52 p., Accompanied by Map 2207, scale 1 inch to 1/2 mile.

Guel, J.J.C. 1973: Geology of Crooks Township, Jarvis and Prince Locations, and Offshore Islands, District of Thunder Bay; Ontario Division of Mines, Geology Report 102, 46 p., Accompanied by Map 2250, scale 1 inch to 1/2 mile.

Halet, R.A., 1965: The Zenith Zinc Mine; Western Miner, October, 1965.

Harcourt, G.A., and Bartley, M.W., 1939: The Schreiber area; Ontario Department of Mines, Annual Report XLVII, Part IX, Accompanied by Map 47j, one inch to 1/2 mile.

Harris, D.C., 1986: The Hemlo Gold Deposits, Ontario; in J. Pirie and M.J. Downes, (eds), Proceedings fo Gold '86, an International Symposium on the Geology of Gold Deposits, Excursion Guidebook, p.11-165.

Harris, D.C., 1987: Geology of the Hemlo Deposit, Ontario; Institute on Lake Superior Geology, Thirty-third Annual Meeting, Wawa, Ontario, May, 1987, Field Trip Guidebook, v. 33, part 4, 57 p.

Harris, D.C., 1989: The Mineralogy and Geochemistry of the Hemlo Gold Deposit, Ontario; Geological Survey of Canada, Economic Geology Report 38, 88 p.

Hassler, S.W. and Simonson, B.M., 1989: Deposition and alteration of volcaniclastic strata in two large, early Proterozoic iron-formations in Canada: Canadian Journal of Earth Sciences, v. 26, p. 1574-1585.

Heaman, L.M. and Michado, N., 1987: Isotope Geochemistry of the Coldwell Alkaline Complex: I. U-Pb Studies on Accessory Minerals; in, in Program with Abstracts, Geological Association of Canada, Annual Meeting, Saskatoon, Saskatchewan, p. 54.

Heslop, J., 1970: Geology, mineralogy and textural relationships of the Coppercorp deposit, Mamainse Point, Ontario; unpublished MSc thesis, Carleton University, Ottawa, 103 p.

Hough, J.L. 1958: Fresh-water Environment of Deposition of Precambrian Banded Iron Formations; Journal Sedimentary Petrology, v. 28, No. 4, p. 414-430.

Hudak, G.J.,III, 1989: The physical volcanology and hydrothermal alteration associated with the F-Group Archean volcanogenic massive sulphide deposit, Sturgeon Lake, northwestern Ontario; Unpublished M.Sc. Thesis, University of Minnesota-Duluth, Duluth, MN, 172 p.

Hugon, H., 1984: The Hemlo Deposit: Gold Mineralization within a Dextral Shear Zone: in Summary of Field Work, Ontario Geological Survey, Miscellaneous Paper 119, p. 212-217.

Hugon, H., 1986: The Hemlo Gold Deposit, Ontario, Canada: A Central Portion of a Large-Scale, Wide Zone of Heterogeneous Ductile Shear; in A.J. MacDonald, (ed), Proceedings of Gold '86, An International Symposium on the Geology of Gold Deposits, Toronto, Ontario, p. 379-387.

Jongewaard, P.K., 1989: Physical volcanology and hydrothermal alteration of the footwall rocks to the Archean Sturgeon Lake massive sulphide deposit, northwestern Ontario; Unpublished M.Sc. Thesis, University of Minnesota-Duluth, Duluth, MN, 140 p.

Jost, M., 1968: The White Pine copper deposit, Michigan, U.S.A., Mineralogical investigation and genetic interpretation; Unpublished Ph.D. dissertation, Johannes Gutenberg Universitat, Mainz, Germany.

Kite, B.T., 1981: The Geology of the Zenith Zinc Deposit near Schreiber, Ontario; unpublished B.Sc. Thesis, Lakehead University, Thunder Bay, Ontario.

Koopman, E.R., Dube, B., Franklin, J.M., Poulsen, and Patterson, M.R., 1989a: Deformation of the Lyon Lake massive sulphide deposit, Wabigoon Subprovince, northwestern Ontario, Canada; in Program with Abstracts, Institute on Lake Superior Geology, Duluth, Minnesota.

Koopman, E.R., Franklin, J.M., and Poulsen, K.H., 1989b: Structural and stratigraphic controls of ore distribution in the Lyon Lake Zn-Cu-Pb-Ag deposit, Sturgeon Lake, northwestern Ontario; Ontario Mines and Minerals Symposium Mineral Development and Lands Branch Mineral Development Forum, December 1989.

Koopman, E.R., Franklin, J.M., and Poulsen, K.H., 1990: Structural and stratigraphic controls of ore distribution of the Lyon Lake Zn-Cu-Pb-Ag deposit, Sturgeon Lake, Ontario; in Program with Abstracts, Geological Survey of Canada Minerals Colloquium, January 1990.

Kuhns, R.J., 1988: The Golden Giant deposit, Hemlo, Ontario: Geological and geochemical relationships between mineralization, alteration, metamorphism, magmatism, and tectonism; unpublished Ph.D. thesis, University of Minnesota, Minneapolis, Minnesota, 458 p.

Kuhns, R.J., Kennedy, P., Cooper, P., Brown, P., Mackie, B., Kusins, R., and Friesen, R., 1986: Geology and mineralization associated with the Golden Giant deposit, Hemlo, Ontario, Canada; in A.J. Macdonald, (ed), Proceedings of Gold '86, An International Symposium on the Geology of Gold Deposits, Toronto, Ontario, p. 327-339.

Lefolii, K.P., 1987: Claims: Adventures in the gold trade; Key Porter Books Limited, Toronto, Canada, 264 p.

Leone, R.J., Seasor, R.W., and Rohrbacher, T.J., 1971: Geology of the White Pine area; in Society of Economic Geologists, Guidebook for Field Conference, Michigan Copper District, Michigan Technological University press, p. 44-66.

Lilley, F.E.M., 1964: An Analysis of the Magnetic Features of the Port Coldwell Intrusive; unpublished M.Sc. thesis, University of Western Ontario, London, Ontario.

Lipman, P.W., 1976: Caldera-collapse breccias in the western San Juan Mountains, Colorado; Geological Society of America Bulletin, v. 87, p. 1397-1410.

Lipman, P.W., 1989: Excursion 16B: Oligocene-Miocene San Juan volcanic field, Colorado; in C.E. Chapin and J. Zidek, (eds), Field excursions to volcanic terrains in the western United States, Volume I: Southern Rocky Mountain region; New Mexico Bureau of Mines and Mineral Resources Memoir 46, p. 303-380.

Logan, W.E., 1846: On the Geology and Economic Minerals of Lake Superior; Geological Survey of Canada, Report of Progress, 1846-47, p. 8-34.

Lum, H.K., 1973: Petrology of the Eastern Gabbro and Associated Sulphide Mineralization of the Coldwell Alkalic Complex, Ontario; unpublished B.Sc. thesis, Carleton University, Ottawa, Ontario, 68p.

Mainwaring, D.R. and Naldrett, A.J., 1977: Country-rock assimilation and the genesis of Cu-Ni sulfides is the Water Hen intrusion, Duluth complex, Minnesota; Economic Geology, v. 72, p. 1264-1284.

Marmont, S. and Carter, M.W., 1983: Terrace Bay - Zenmac Area Has a Long History of Mineral Exploration; Northern Miner, March 10, 1983.

Mattinen, P.R., 1978: Exploration Possibilities in the Zenith Mine (Big Duck Lake) Area, Schreiber, Ontario; internal Corporation Falconbridge Copper report.

McIlwaine, W.H., 1971a: McTavish Township (West Part of North Half), District of Thunder Bay; Ontario Department of Mines and Northern Affairs, Preliminary Map P.720, scale 1 inch to 1/4 mile.

McIlwaine, W.H., 1971b: McTavish Township (East Part of North Half) District of Thunder Bay; Ontario Department of Mines and Northern Affairs, Preliminary Geological Series, Map P. 721, scale 1 inch to 1/4 mile.

Milne, V.G., 1967: Geology of Cirrus Lake-Bamoos Lake Area; Ontario Department of Mines, Geological Report 43, Accompanied by Maps 2098 and 2099, scale 1 inch to 1/2 mile.

Milne, V.G., 1968. Geology of the Black River Area, District of Thunder Bay. Ontario Department of Mines, Geological Report 72, Accompanied by Maps 2143, 2144, 2145, 2146, and 2147, scale 1 inch to 1/2 mile.

Milne, V.G., 1974: Mapledoram-Gemmel Townships; Ontario Department of Mines, Geological Map 2280, scale 1 inch to 1000 feet.

Mitchell, R.H., and Platt, R.G., 1982: The Coldwell Alkaline Complex; in J.M. Franklin, (ed), Proterozoic Geology of the Northern Lake Superior Area, Field Trip Guidebook, Geological Association of Canada-Mineralogical Association of Canada, Joint Annual Meeting, Winnipeg, Manitoba, p. 42-61.

Moore, J.G. and Peck, D.L. 1962: Accretionary Lapilli in Volcanic Rocks of the Western Continental U.S.; Journal of Geology, v. 10, No. 2, p. 182-193.

Moorhouse, W.W. 1960: Gunflint Iron Range in the Vicinity of Port Arthur; Ontario Department of Mines, v. 69, part

7, p. 1-40, Accompanied by 7 maps, Scale 1 inch to 1/2 mile.

Morey, G.B. 1969: The Geology of the Middle Precambrian Rove Formation in Northeastern Minnesota; Minnesota Geological Survey Sp-7, Special Publication Series, University of Minnesota, Minneapolis, 62 p.

Morey, G.B. 1972: Gunflint Range; in P.K. Sims, and G.B. Morey, (eds), Geology of Minnesota - A Centennial Volume (Schwarcz Volume), Minnesota Geological Survey, p. 218-225.

Morton, R.L. and Franklin, J.M., 1987: Two-fold classification of Archean volcanogenic massive sulphide deposits; Economic Geology, v. 82, p. 1057-1063.

Morton, R.L., Hudak, G.J., and Franklin, J.M., 1989: The Mattabi ash flow tuff and its relationship to the Mattabi massive sulphide deposit; IAVCEI abstracts, New Mexico Bureau of Mines and Mineral Resources Bulletin 131, p. 196.

Morton, R.L., Hudak, G.J., Walker, J.S., and Franklin, J.M., 1990: The volcanology of the south Sturgeon Lake area; in Program with Abstracts, Geological Survey of Canada Minerals Colloquium, January 1990, p. 27.

Morton, R.L., Walker, J.S., and Hudak, G.J., 1988: Geology and metallogeny of the South Sturgeon Lake Greenstone Belt; in Program with Abstracts, Geological Survey of Canada Current Activities Forum, Ottawa, January, 1988.

Morton, R.L., 1984: Lithology, Stratigraphy and Hydrothermal Alteration of Volcanic Rocks in the Winston-Cleaver Lakes Area, Ontario; Summary of work for Corporation Falconbridge Copper, January 1984.

Muir, T.L. and Elliott, C.G., 1987: in Summary of Field Work and Other Activities, Ontario Geological Survey, Miscellaneous Paper 137, p.117-129.

Muir, T.L., 1982a: Geology of the Heron Bay Area, District of Thunder Bay; Ontario Geological Survey, Geological Report 218, 89 p., Accompanied by Map 2439, scale 1 inch to 1/2 mile.

Muir, T.L., 1982b: Geology of the Hemlo Area, District of Thunder Bay; Ontario Geological Survey, Geological Report 217, 65 p. Accompanied by Map 2452, scale 1 inch to 1/2 mile.

Muir, T.L., 1983: Geology of the Hemlo-Heron Bay Area; in A.C. Colvine, (ed), The Geology of Gold in Ontario; Ontario Geological Survey, Miscellaneous Paper 110, p. 230-239.

Muir, T.L., 1986: Hemlo Tectono-Stratigraphic Study; in Summary of Field Work and Other Activities, Ontario Geological Survey, Miscellaneous Paper 132, p. 95-106.

Muir, T.L., Elliott, C.G. and Corfu, F., 1988: The Tectono-Stratigraphic Setting of the Hemlo Au-Mo Deposit, Ontario, Canada; in Extended Abstracts, Poster Program, Bicentennial Gold '88, Melbourne, Australia, p. 95-97.

Mumin, A.H., 1988: Tectonic and structural controls on massive sulphide deposition in the south Sturgeon Lake volcanic pile, northwestern Ontario, and Hydrothermally altered rocks associated with the Lyon Lake massive sulphide ore deposits, Sturgeon Lake, northwestern Ontario; Unpublished M.A.S. Thesis, University of Toronto, Toronto, Ontario, 166 p.

Nebel, M., 1982: Stratigraphy, depositional environment, and alteration of Archean felsic volcanics, Wawa, Ontario: Unpublished M.Sc. Thesis, University of Minnesota-Duluth, Duluth, MN, 114 p.

Page, T.W., 1948: A Report on the Property of Lake Superior Mining Corporation Ltd., Hemlo Area, Ontario; unpublished report, Resident Geologist's Files, Ministry of Northern Development and Mines, Thunder Bay, Ontario, 23 p.

Patterson, G.C., 1984: Field Trip Guidebook to the Hemlo Area; Ontario Geological Survey, Miscellaneous Paper 118, 31 p.

Patterson, G.C., 1985: Amethyst in the Thunder Bay Area; Ontario Ministry of Natural Resources, unpublished report.

Pearson, W.N., 1980: Copper metallogeny, North Shore region of Lake Huron, Ontario; Unpublished Ph.D. dissertation, Queens University, Kingston, 403 p.

Peterman, Zell E. 1966: Rb-Sr Dating of Middle Precambrian Metasedimentary Rocks of Minnesota; Geological Society of America Bulletin, v. 77, p. 1031-1044.

Petersen, E.U., 1984: Metamorphism and Geochemistry of the Geco Massive Sulphide Deposit and its Enclosing Wall-Rocks; unpublished Ph.D. Thesis, University of Michigan, Ann Arbor, Michigan.

Pirie, I.D., 1979: Report on the Geology of the Winston Lake Area near Schreiber, Ontario; internal Corporation Falconbridge Copper field report.

Poulsen, K.H. and Franklin, J.M., 1981: Copper and gold mineralization in an Archean trondhjemitic intrusion, Sturgeon Lake, Ontario; in Current Research, Part A, Geological Survey of Canada Paper 81-1A, p. 9-14.

Pye, E.G., 1957: Geology of the Manitouwadge Area; Ontario Department of Mines, v. LXVI, part 8, 114 p.

Pye, E.G., 1964: Mineral Deposits of the Big Duck Lake Area; Ontario Department of Mines, Geological Report 27.

Ramsay, J.G., 1967: Folding and fracturing of rocks; McGraw-Hill, New York, 568 p.

Rigg, D.M., 1983: Geology and Geochemistry of the North Lake Area, Northwestern, Ontario; internal Corporation Falconbridge Copper report.

Ripley, E.M., 1981: Sulfur isotopic studies of the Dunka Road Cu-Ni deposit, Duluth complex, Minnesota; Economic Geology, v. 76 p. 610-620.

Riverin, G. and Hodgson, C.J., 1980: Wall-rock alteration at the Millenbach Cu-Zn mine, Noranda, Quebec; Economic Geology, v. 75, p. 424-444.

Roach, S.N., 1982a: Summary of 1981 Exploration Program Winston Lake Project; internal Corporation Falconbridge Copper report.

Roach, S.N., 1982b: Summary of 1981 Exploration Program, Zenmac Option; internal Corporation Falconbridge Copper report.

Robinson, P.C., 1979: Geology and Evolution of the Manitouwadge Migmatite Belt, Ontario, Canada; unpublished Ph.D. Thesis, University of Western Ontario, London, Ontario.

Rohrbacher, T.J., 1968: Selected trace elements, copper and iron distribution in the Coper Harbour conglomerate, White Pine, Michigan; Unpublished M.Sc. thesis, University of Toledo, 111 p.

Rowe, R., 1982: Stratigraphic relations of the Crandon massive sulfide deposit, Wisconsin; in Program and Abstracts, Geological Association of Canada, Annual Meeting, Winnipeg.

St-Onge, M.R., Lucas, S.B., Scott, D.J., and Begin, N.J., 1989: Evidence for the development of oceanic crust and for continental rifting in the tectonostratigraphy of the Early Proterozoic Cape Smith Belt; Geoscience Canada, v.

16, no. 3, p. 119-122.

Sage, R.P., Treacher, K., Meloche, D., and Bathe, D. 1975: Slate Islands, District of Thunder Bay; Ontario Division of Mines, Preliminary Map P. 997, Geological Series, Scale 1 inch to 660 feet or 1:7,920, Geology and compilation, 1974.

Sangster, D.F., 1972: Precambrian Volcanogenic Massive Sulphide Deposits in Canada: A Review; Geological Survey of Canada, Paper 72-22, 44 p.

Schnieders, B.R., 1987: The geology of Sulphide-facies iron formation and associated rocks int he Lower Steel River-Little Steel Lake area, Terrace Bay, Ontario; Unpublished M.Sc. Thesis, Lakehead University, Thunder Bay, Ontario, 1968.

Severin, P.W.A., 1982: The geology of the Sturgeon Lake Cu-Zn-Pb-Ag-Au deposit; Canadian Institute of Mining and Metallurgy Bulletin, v. 75, p. 107-123.

Severin, P.W.A., 1979: Proposed Exploration-Big Duck Lake Area, Schreiber, Ontario; internal Corporation Falconbridge Copper report.

Severin, P.W.A., 1980: Summary Report and Recommendations, Winston Lake Area; internal Corporation Falconbridge Copper report.

Shegelski, R.J., 1982: Field Guide: The Gunflint Formation in the Thunder Bay area; in J.M. Franklin, (ed), Field Trip Guidebook Trip 4, Proterozoic geology of the Northern Lake Superior area, Geological Association of Canada-Mineralogical Association of Canada, p. 15-31.

Sibson, R.H., Robert, F., and Poulsen, K.H., 1988: High angle reverse faults, fluid pressure cycling, and mesothermal gold-quartz deposits; Geology, v. 16, p. 551-555.

Simmons, B.D., 1983: Winston Lake Discovery; Presentation to the Prospectors and Developers Convention in Toronto, Ontario.

Siragusa, G.M. and Chivers, K.M., 1985: White Lake Area, District of Thunder Bay; in Summary of Field Work and Other Activities, Ontario Geological Survey, Miscellaneous Paper 126, p.75-77.

Siragusa, G.M., 1985: Precambrian Geology of the White Lake (Hemlo) Area, Theresa Lake Section, District of Thunder Bay, Ontario; Ontario Geological Survey, Preliminary Map P.2738, scale 1 inch to 1/4 mile.

Stockwell, C.H., 1964: Part II. Geological Studies: Fourth report on structural provinces, orogenies, and time-classification of rocks of the Canadian Precambrian Shield; Geological Survey of Canada, Paper 64-17, part II, 29 p.

Suffel, G.G., Hutchinson, R.W., and Ridler, R.H., 1971: Metamorphism of Massive Sulphides at Manitouwadge, Ontario, Canada; Society of Mining Geologists of Japan, Special Issue 3, p. 235-240.

Tanton, T.L., 1930: Zenith Zinc Mine Rossport: Zinc and Lead Deposits of Canada; Geological Survey of Canada, Economic Geology Series No.8.

Tanton, T.L. 1931: Fort William and Port Arthur and Thunder Cap Map Areas: Thunder Bay District, Ontario; Geological Survey of Canada, Memoir 167, 222 p.

Thomson, J.E., 1931: Geology of the Heron Bay Area, District of Thunder Bay; Ontario Department of Mines, v. 40, part.2, p.21-39, Accompanied by Map 40d, scale 1 inch to 1 1/2 miles.

Thomson, J.E., 1932: Geology of the Heron Bay-White Lake Area, District of Thunder Bay; Ontario Department of Mines, v. 41, part.6, p. 34-52, Accompanied by Map 41j, scale 1 inch to 2 miles.

Timms, P.D. and Marshall, D., 1959: The Geology of the Willroy Mines Base Metal Deposits; Geological Association of Canada Proceedings, v. 11, pp. 55-65.

Trowell, N.F., and Johns, G.W., 1986: Stratigraphic correlation of the western Wabigoon subprovince, northwestern Ontario; in J. Wood and H. Wallace, (eds), Volcanology and Mineral Deposits; Ontario Geological Survey Miscellaneous Paper 129, p. 50-61

Trowell, N.F., 1974: Geology of the Bell Lake-Sturgeon Lake area, districts of Thunder Bay and Kenora; Ontario Geological Survey Report 114, 67 p.

Trowell, N.F., 1983: Geology of the Sturgeon Lake area, districts of Thunder Bay and Kenora; Ontario Geological Survey Report 221, 97 p.

Unger, D., 1980: Report on the 1980 Geological Mapping of the Zenmac Option; internal Corporation Falconbridge Copper report.

Vos, M.A., 1976: Amethyst Deposits of Ontario; Ontario Division of Mines, Ministry of Natural Resources, Geological Guidebook No. 5, 99 p.

Walford, P., Stephens, J., Skrecky, G., and Barnett, R., 1986: The geology of the 'A' zone, Page-Williams mine, Hemlo, Ontario, Canada; in A.J. Macdonald, (ed), Proceedings of Gold '86, An International Symposium on the Geology of Gold Deposits, Toronto, Ontario, p. 362-378.

Watkinson, D.H., Dahl, R., and McGoran, J.W., 1986: The Coldwell Complex Platinum-Group Element Deposit: 2. Relationships of Platinum-Group Elements to Pegmatitic, Biotite-Bearing Gabbro and the Role of a Fluid Phase; in Program with Abstracts, Geological Association of Canada-Mineralogical Association of Canada- Canadian-Geophysical-Union, Joint Annual Meeting, Ottawa, Ontario, v. 11, p. 142.

Watkinson, D.H., Mainwaring, P.R., and Lum, H.K., 1973: Petrology and Copper Mineralization of the Coldwell Complex, Ontario: in Program with Abstracts, Geological Society of America, Annual Meeting 5, p. 856.

Weege, R.J. and Pollock, J.P., 1971: Recent developments in the native-copper district of Michigan; in Society of Economic Geologists, Guidebook for field conference, Michigan Copper District, Michigan Technological University Press, p. 18-43.

Wells, G.S., 1985: Summary of 1980 Exploration Program, Zenmac Option; internal Corporation Falconbridge Copper report.

Williams, H., Turner, F.J., and Gilbert, C.M., 1954: Petrography; W.H. Freeman and Co., Publishers, San Francisco, 406 p.

Williams, H.R., and Breaks, F.W., 1989: Geological Studies in the Manitouwadge-Hornepayne Area; in Summary of Field Work and Other Activities, Ontario Geological Survey, Miscellaneous Paper 146, p. 79-96.

Wolfson, I., 1988: Petrographic Report, Lyon Lake ore samples; Unpublished company report.

Wood, J., and Wallace, H., (eds.) Volcanology and Mineral Deposits, Ontario Geological Survey Misc. Paper 129. p.

50-61.

Yule, G., 1979: Investigations of the Good Morning Lake radioactive fault breccia, Innes Lake, Dorion Township, Northwestern Ontario; H.B.Sc. Thesis, Lakehead University, Thunder Bay, Ontario, 93 p.