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GEOLOGY OF THE LAKEHEAD REGION

by

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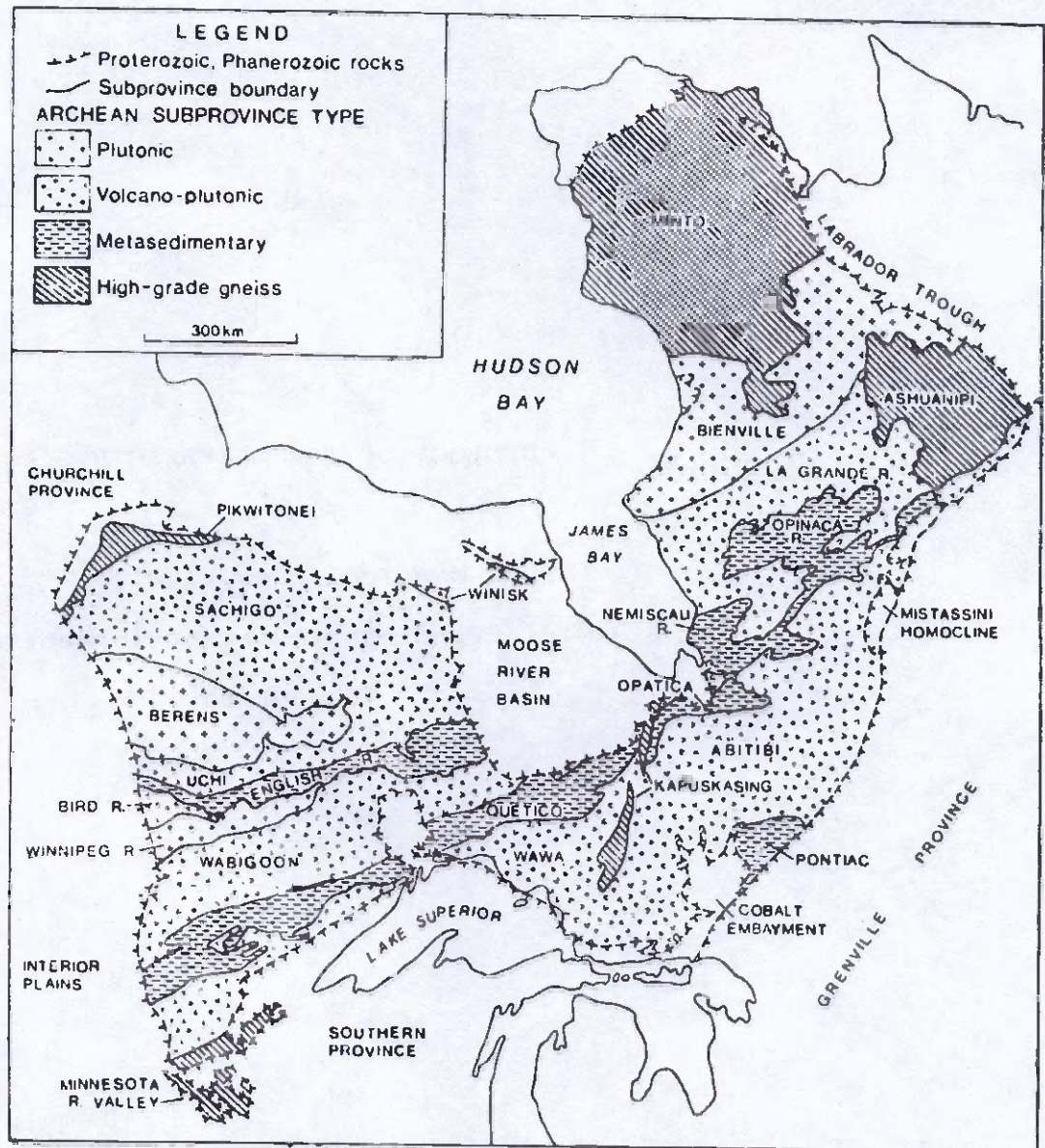


Figure 1: Lithotectonic Subprovinces of the Superior Province (after Card and Ciesielski, 1986)

INTRODUCTION

The Lakehead Region is an area bound to the east by Lake Superior, to the south by the international border, includes Lake Nipigon, and extends to the west as far as the town of Atikokan. This region is underlain by precambrian rocks of the Canadian Shield, in particular, Archean rocks the Superior Province and Proterozoic rocks of the Southern Province (Fig 1). Three subprovinces of the Superior Province are represented; Wabigoon, Quetico, and the Abitibi-Wawa. The Southern Province is subdivided into the Nipigon Plate and the Port Arthur Homocline, the northern fringe of the Animikie basin (enclosed map).

Stockwell (1964,1970) and Card and Ciesielski (1986) subdivided the Superior Province on the basis of structure and lithology. In this region, the subprovinces are supracrustal-intrusive complexes, the predominance of either volcanic or sedimentary rock being a key distinguishing characteristic. The Abitibi-Wawa and the Wabigoon Subprovinces are arcuate assemblages of volcanic rock dominated belts, "greenstone belts", and tonalitic to granodioritic, massive to gneissic intrusions. The Quetico Subprovince is a linear metasedimentary terrane cored by granitic intrusions and migmatites. The subprovince boundaries are broad schist zones and/or distinct faults, the Seine River-Quetico fault couple, which separate the Quetico from the Wabigoon, being good examples of both of these types of strain.

The Nipigon Plate and the Port Arthur Homocline consist of flat lying sedimentary rock unconformably overlying the Superior Province. These also include diabase sills and dykes, late mafic intrusions and volcanic rocks.

ARCHEAN STRATIGRAPHY

High precision U-Pb zircon geochronology, carried out at the Royal Ontario Museum, has resulted in a dramatic change in the rate of increased understanding of the archean. Absolute dates combined with structural interpretations of well documented lithological distribution has resolved the stratigraphy within some greenstone belts. It has also helped resolve the relationships between plutonism, volcanism, sedimentation, metamorphism, deformation and mineralization. Many stratigraphic relationships within subprovinces, and in between subprovinces have also been resolved. In the Lakehead Region some fundamental distinctions, and similarities, can now be drawn in between the Wabigoon and the Abitibi-Wawa tonalite-greenstone terranes and their relationship to the intervening paragneiss of the Quetico.

In Atikokan, the Marmion Lake Batholith, a tonalite complex, has been dated at 3004 ma (Davis et al, 1988). This batholith is unconformably overlain by the Steeprock Group sedimentary rocks and subsequent volcanic rocks. This unconformity is well exposed on the east wall of the Roberts pit. Stone et al.(in press) describe the unconformity and the overlying Wagita Formation as follows; "The unconformity at the base of the Wagita Formation is marked by a transition from unaltered tonalite to sericitized tonalite and a sericite-carbonate-quartz 'grit', which becomes well bedded and clast bearing up section". The Wagita formation is overlain by the stromatolitic Mosher Carbonate Formation and altered iron formation (Jolliffe Ore Zone Formation). This sedimentary sequence is overlain in ascending order by; ultramafic pyroclastics, mafic pillow lavas and intermediate to felsic flows, tuffs and breccias. This sequence is then overlain by clastic sediments.

Felsic tuffs in the Lumby Lake Greenstone Belt, on the

north side of the Marmion Lake Batholith, have been dated at 2950 ma and 2999 ma (Davis and Jackson, 1988). These are considered to be the stratigraphic equivalents to the felsic volcanics in the Steeprock mine area. This older, carbonate sediment bearing, volcanic-sedimentary sequence has been recognized in many greenstone belts in the northern parts of the superior province (Thurston and Chivers, 1990). These sequences, as well as quartz arenites-iron formation-ultramafic lava sequences recognized in other belts, have shallow water sedimentary features. This is the basis for the recognition of early archean platformal sequences which developed at about 3000 ma (Wood et al, 1986).

These shallow water quartz-arenite and carbonate bearing sequences (>2.85 Ga) are generally overlain by mafic-ultramafic volcanic sequences (2.740 to 2.700 Ga) and mafic to felsic volcanic cycles (2.775 to 2.700 Ga) and comprise what is stratigraphically referred to as "Keewatin". Keewatin rocks volumetrically dominate all greenstone belts in the Superior province. The younger of these Keewatin volcanic sequences are not recognized in the Atikokan area. They are however, the only recognizable volcanic sequence older than 2.700 Ga in the Shebandowan greenstone belt of the Abitibi-Wawa subprovince. Thurston (1986) interprets two mafic to felsic volcanic cycles. Corfu and Stott (1986) have generated a date of 2.733 Ga for felsic volcanics in the Middle Shebandowan Lake area. Stratigraphic relationships in this belt have been masked by intense deformation.

A common thread for both the Abitibi-Wawa and the Wabigoon is that they both contain younger "Temiskaming"-type sequences, found in many greenstone belts in the Canadian Shield. These volumetrically insignificant sequences consist of alluvial-fluvial sedimentary rock, mainly conglomerate and arkose with or without alkalic and shoshinitic volcanic rock (Shegelski, 1980). In the

Atikokan-Mine Centre area, a distinctive conglomerate unit, referred to as the Seine Conglomerate, is found along the Quetico-Wabigoon subprovince boundary. The contact relationship with adjoining Quetico sediments and Wabigoon volcanics is unclear because of poor exposure and shearing. Poulsen (1984), deduced that the Seine Conglomerates were younger on the basis of preserved evidence of deformation history. This is supported by U-Pb dates of 2685 ma for a clast in the conglomerate and an age range of 2700 to 2795 ma for detrital zircons in the Quetico Metasediments (Davis et al, 1990).

In the northern Shebandowan greenstone belt, interdigitated "Temiskaming" alkaline volcanics, conglomerates and sandstones are dated at 2689 ma and the underlying "Keewatin" volcanics at 2733 ma (Corfu and Stott, 1986). The contact relationship is obscured by well developed foliation leading to speculation that the "Temiskaming" rocks developed in a fault bounded through. A second belt of Temiskaming rocks a few miles to the south has a clearly disconformable stratigraphic relationship with underlying "Keewatin" volcanic rocks.

In summary, the stratigraphy within the volcanic-tonalite terrane consist of 3 b.y. old tonalite basement, overlain by a 3 b.y. old platformal sequence, 2.74 b.y. sub-alkalic volcanics and 2.68 b.y. old calc-alkalic volcanics and sedimentary rocks. Recent age dates on detrital zircons from Quetico greywacke range from 2700 to 2795 ma (Davis et al, 1990) strongly suggesting a coeval relationship with the sub-alkalic volcanics. The suggestion by Williams (1987), that the Quetico represents an accretionary prism is well accepted. A host to the 3 b.y. tonalite intrusive has yet to be recognized.

The last major geological event in the tonalite-greenstone terrane is represented by the voluminous tonalites which intruded from 2690 to 2677 Ma (Colvine, et

al 1988). The granitic rocks in the Quetico intruded from 2670 to 2650 Ma (Percival, 1989). Both of these subprovinces have widespread mafic intrusions which are not well constrained geochronologically.

The isoclinally folded, steeply dipping Archean rocks of the Superior Province, are overlain by the shallow dipping Proterozoic rocks of the Southern Province. The Port Arthur Homocline consists of the Gunflint Formation (taconite, algal chert, limestone, and tuffaceous shale) and the overlying Rove Formation (shale, greywacke). The age of these rocks is poorly constrained but is considered to be older than 1850 Ma. (Morey, 1983). The Nipigon Plate consists of the Pass Lake Formation (quartz arenite), the Rossport Formation ("redbeds" - dolomite, sandstone, chert, mudstone) and the Kama Hill Formation (mudstone). These three formations comprise the Sibley Group which unconformably overlies Archean rocks and the Rove Formation. These sedimentary rocks are overlain by the Keweenaw volcanic rocks of the Osler Group. They have also been intruded by early Keweenaw diabase, the dominant lithology in the Nipigon Plate, and late Keweenaw gabbroic stocks and dykes.

METALLOGENY

The Lakehead Region is host to a wide variety of types of mineral deposits. The greenstone-tonalite terrane contain volcanogenic Cu-Zn deposits associated with felsic volcanic rocks, komatiite (intrusive ? extrusive ?) hosted Ni-Cu deposits, mafic intrusive hosted platinum metal group deposits, iron deposits, and lode gold deposits. The

Quetico subprovince contains subeconomic rare element deposits in pegmatites and platinum metal group and Cu-Ni deposits in mafic intrusives. Proterozoic rocks contain numerous veins in sedimentary rocks with varied amounts of silver, zinc, lead, barite and fluorite in; native copper deposits in mafic volcanic rocks; and platinum metal group-Cu-Ni deposits in mafic intrusives.

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FIELD TRIP 1

MAFIC INTRUSIONS, PLATINUM-GROUP-ELEMENTS MINERALIZATION AND
GRANITOID ROCKS IN THE LAC DES ILES AREA

Introductory Discussion and Field Guide
36th Annual Institute on Lake Superior Geology
Thunder Bay, Ontario

by

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INTRODUCTION

The Lac des Iles region provides an opportunity to examine a wide range of Late Archean plutonic rock types. This field trip will emphasize: the role of mafic magmatism in the evolution of this Late Archean plutonic terrane; the relationship between mafic magmas and granitoid plutonism; and aspects of platinum-group-elements mineralization.

The field trip is based on surveys done in the Lac des Iles area by the Ontario Geological Survey in 1985 and 1986. This work included detailed mapping of the Lac des Iles Complex (Sutcliffe and Sweeny, 1986) and Tib Gabbro (Smith and Sutcliffe, 1987), regional mapping of the granitoid rocks (Sutcliffe and Smith, 1988), and a gravity survey of the area (Gupta and Sutcliffe, in press). The discussion presented here makes considerable use of excerpts from Sutcliffe (1989) and Sutcliffe *et al* (1989) to which the reader is referred for further details on aspects of the geochemistry of the plutonic rocks. Previous mapping in the area is primarily the reconnaissance mapping and compilation of Sage *et al* (1974).

OVERVIEW

Mafic to ultramafic intrusive rocks in the Lac des Iles area form part of an east-northeast trending linear zone of Late Archean mafic plutons which extends over 200 km from Atikokan to Lake Nipigon. This zone parallels the boundary between the Wabigoon and Quetico Subprovinces (Stockwell *et al* 1972). The rocks examined on this field trip are all within the Wabigoon Subprovince.

In the immediate region of the field trip, the mafic to ultramafic plutons are intruded into gneissic tonalite host rocks and are distributed in a circular pattern, the diameter of which is approximately 30 km (Figure 1). These plutons are composed of serpentinite, wehrlite, minor lherzolite, clinopyroxenite and websterite to magnesian gabbro and ferrogabbro (Sutcliffe and Smith 1988). The Lac des Iles Complex (LDIC) is the largest of the mafic to ultramafic plutons and displays the most complete spectrum of lithologies. The Tib Gabbro is the second largest intrusion and consists of predominantly gabbro. The marginal zones of larger intrusions and some of the smaller intrusions consist of hornblende, and hornblende gabbro to diorite. Most of the mafic to ultramafic intrusions have well preserved igneous mineralogy and are not significantly deformed. The form and tectonic setting of the mafic-ultramafic intrusions, particularly the LDIC, is similar to mafic intrusions associated with orogenic terrains such as the Mesozoic Alaskan complexes (Taylor 1967, Findlay 1969). Preliminary U/Pb zircon data indicate that the gabbroic rocks of the Lac des Iles Complex and the Tib Gabbro are 2.69 Ga (D.W. Davis, Royal Ontario Museum, unpublished data), however, an older Sm/Nd age of 2,738 $\pm$ 27 Ma is reported for the ultramafic rocks of the LDIC by Brugmann and Naldrett (1987).

A major zone of platinum-group-elements (PGE) and Cu-Ni sulphide mineralization, known as the Roby Zone, occurs in gabbroic rocks of the LDIC. The property is being developed by Madelaine Mines Limited and this may become the first primary producer of platinum group elements (PGE) in Canada (Northern Miner Press 1988). The PGE mineralization in the Roby Zone of the LDIC is associated with sulphides and altered silicates in gabbroic rocks. This mineralization exhibits characteristics suggesting the control of both magmatic mixing processes (Naldrett and Campbell 1979; Todd *et al* 1982; Sharpe 1985) and volatile dominated processes (Talkington and Watkinson 1984; Boudreau *et al* 1986; Stumpfl and Ballhaus 1986). PGE mineralization also occurs in ultramafic rocks of the LDIC and in other intrusions in the area and these occurrences indicate that PGE's were concentrated at several stages in the fractionation of the mafic magmas.

Two granitoid plutons consisting of hornblende tonalite and biotite tonalite intrude older gneissic tonalite. The gneissic tonalite is dated at approximately 2.77 Ga and the younger hornblende tonalite at 2.73 Ga by the U/Pb zircon method (D.W. Davis, unpublished data). The younger granitoid plutons contain numerous net-veined mafic dikes and other textures indicating the coexistence of mafic and felsic magmas. Although the association of mafic magmas with late Archean granitoids has not been commonly recognized, there are other examples of associated Late Archean mafic and felsic magmatism in the Wabigoon Subprovince (e.g. Morrison *et al* 1986). Based on field relationships, Sutcliffe (1989) considered that hornblende tonalite and the mafic-ultramafic intrusions were contemporaneous. The geochronology does not support this conclusion and the mixing textures therefore are probably associated with an older phase of mafic magmatism than that which generated the major mafic-ultramafic plutons. Further studies will test this hypothesis.

In this guide, plutonic rocks are named using the classification of Streckeisen (1976).

LAC DES ILES COMPLEX

Geology

Mapping by Pye (1968) indicated that the Lac des Iles Complex (LDIC) has an area of approximately 30 km². The LDIC (Figure 2) consists of a predominantly ultramafic sequence centred on Lac des Iles and predominantly gabbroic sequences south of the lake (Pye, 1968; Watkinson and Dunning, 1979). The description given here is largely based on mapping by Sutcliffe and Sweeny (1986) and the work of Sutcliffe *et al* (1989). Details of the mineral assemblages and textures for rock types are summarized in Table 1.

A gravity study by Gupta and Sutcliffe (in press) indicates that the LDIC is associated with a 30 mgal gravity

high and is modelled as a funnel shaped body part of which extends to a depth of 4.5 km.

The ultramafic sequence is composed of two coalescing centers which are defined by the different distributions of ultramafic lithologies and by the attitudes of igneous layering (Figure 2). Hornblendite to pyroxene hornblendite and hornblende gabbro occur along the western margin of the LDIC and, with increasing modal pyroxene, grade into pyroxenite. The hornblende-rich lithologies are invaded and brecciated by veins of hornblende diorite to quartz diorite.

The northern ultramafic center is nearly circular in plan with a diameter of approximately 4 km (Figure 2) and consists of several cycles of serpentinite and wehrlite, olivine clinopyroxenite and clinopyroxenite, and websterite to gabbro norite. Lindhardt and Bues (1987) have shown that these cycles are typically 100 to 500 m thick in plan and individual layers can be traced along strike for up to 1.5 km. Olivine cumulates are most abundant around the southern and eastern perimeter of the center and the rocks become more pyroxene-rich towards the northwest. The absence of distinct marker horizons and extent of outcrop exposure places significant limitations on the interpretation of the lateral continuity and thickness of the cycles.

Modal layering due to variation in proportions of plagioclase and pyroxenes, olivine and clinopyroxene, and orthopyroxene and clinopyroxene is observed. Generally the modal layers are 1 to 20 cm thick and are continuous for meters. Minor disseminated euhedral grains of chromite are enclosed in cumulus olivine and clinopyroxene and are locally concentrated in layers up to 1 cm wide in the northern center. Igneous lamination defined by planar alignment of pyroxenes or plagioclase is also observed at several localities. Layer attitudes indicate that the northern center has an upright funnel-shaped form. Minor discordant serpentinite dikes, which contain angular fragments of pyroxenite intrude the northern ultramafic center.

The southern ultramafic center is elliptical in plan with an irregularly shaped wehrlite core centered on Southeast Angle Bay of Lac des Iles. The wehrlite core is surrounded by websterite which in some areas contains inclusions of wehrlite. Along the eastern margin of the center the websterite grades into gabbro norite. The southern center is predominantly composed of massive rocks and lacks well defined igneous layering.

The ultramafic rocks are partially separated from the gabbroic rocks south of Lac des Iles by a septum of tonalite. Watkinson and Dunning (1979) considered that the ultramafics were emplaced later than the gabbroic rocks based on the presence of gabbroic inclusions in the ultramafics. The only clearly defined gabbroic inclusions observed during the current study are troctolitic and not readily related to the gabbroic suite.

A 50 m wide, composite dike consisting of olivine gabbronorite, olivine websterite to lherzolite with hornblende diorite margins outcrops on the south-west shore of Lac des Iles and along the west side of the gabbro. This dike is olivine-rich (picritic) and is potentially a feeder for the southern ultramafic center

Gabbroic rocks occur largely in the southern part of the LDIC (Figure 2) and are host to the Pd, Pt mineralization of the Roby Zone. Based largely on detailed studies of drill core in the immediate vicinity of the Roby Zone, Watkinson and Dunning (1979) indicated that the gabbro could be divided into two major units known as the Eastern Gabbro (EG) and Western Gabbro (WG). As defined by Watkinson and Dunning (1979), the WG consists of interlayered gabbroic, noritic, pyroxenitic, and anorthositic rocks and the EG consists of oxide-rich gabbroic and noritic rocks. This distinction is significant because the Roby Zone is interpreted to occur along the contact of the two units. Macdonald (1985), Sutcliffe and Sweeny (1986), and Sweeny and Edgar (1987) suggested that the relationships between the units are complex and that many of the layers identified in drill core by Watkinson and Dunning (1979) are separate intrusive phases.

The gabbroic part of the LDIC consists of a leucogabbro to anorthositic gabbro unit and a unit of predominantly gabbronorite. In the vicinity of the Roby Zone, these units correspond with the EG and WG respectively, although in detail there is not a simple geographic subdivision (Figure 3). As defined by Sutcliffe et al (1989), both the EG and WG contain cumulus plagioclase, however, the WG contains cumulus orthopyroxene or alteration products pseudomorphing orthopyroxene. This represents a different definition of the units than that of Watkinson and Dunning (1979), however, because there is an important geographic connotation to the terms EG and WG in the Roby Zone, these names have been retained. The EG displays igneous lamination and weakly developed wispy modal layering and is pervasively altered obscuring the primary igneous assemblages. The WG typically consists of gabbronorite to norite and is fresh to altered. Locally, in the vicinity of the "D Zone", the WG contains cumulus olivine. Within the WG near the EG/WG contact, pegmatitic gabbro, gabbro breccia, gabbronorite, and clinopyroxenite form an arc-shaped zone of variable width (Figure 3). In this part of the WG, pegmatitic gabbro phases occur primarily as inclusions, pegmatoidal segregations and discordant pegmatite dikes. Pegmatitic phases are rare in the EG. The WG truncates the igneous lamination and weakly developed modal layering of the EG, indicating that the WG is younger than the EG.

An intrusion of medium-grained hornblende gabbro to leucogabbro with minor coarse-grained hornblende melagabbro occurs south of the EG and WG. The age of the hornblende

gabbro relative to other units of the LDIC has not been determined.

Late dikes of microdiorite-gabbro intrude the gabbroic rocks and locally result in intrusion breccia consisting of angular gabbroic and less common tonalitic fragments in a fine-grained igneous matrix. Late leucocratic tonalitic dikes and veins also cross cut the gabbroic rocks.

Petrological and geochemical studies by Sutcliffe *et al* (1989) indicate that distinct magma sequences can be recognized on a basis of crystallization order, differences in cumulus minerals, intrusive relations and geochemical differences. These sequences are 1) hornblende gabbro, 2) gabbro and gabbro-norite and 3) ultramafic. Where there is evidence of intrusive relations the most primitive sequences are emplaced late and toward the north end of the complex. In the absence of rocks representative of liquids, parental magma compositions can only be constrained by the compositions of cumulus phases. Mineral chemistry and crystallization sequences indicate that the parental magmas of the gabbroic and ultramafic sequences had tholeiitic basalt parental magmas with high alumina and picritic affinities respectively.

Mineralization

Exploration by Texasgulf Canada Limited and Boston Bay Mines Limited in 1975 and 1976 outlined 20.4 million tonnes of mineralization grading 6 ppm Pd and Pt and 0.2% Cu + Ni in the 600 m long Roby Zone along the EG/WG contact (Northern Miner Press 1976). Pt/(Pt+Pd) is in the range of 0.1 (Watkinson and Dunning 1979) to 0.2 (Sutcliffe *et al* 1989). Drill core sections show that the highest grade mineralization occurs in the WG immediately adjacent to the contact with the EG (Figure 4).

PGE mineralization in the gabbroic rocks is associated with 1) finely disseminated sulphide mineralization (1 to 2 modal % sulphide) in altered pyroxenite sheets along the WG/EG contact 2) disseminated, bleb-like and net-textured sulphide mineralization in gabbro-norite (1 to 4 modal % sulphide) and 3) coarse interstitial sulphide in pegmatitic gabbro. In addition to mineral assemblages noted in Table 1, rare euhedral apatite is observed in mineralized gabbros in the Roby Zone.

A detailed study of drill core from the Roby Zone (Dunning 1979; Watkinson and Dunning 1979; Cabri and Laflamme 1979) showed that vysotskite and braggite are the most abundant platinum group minerals (PGM) along with minor isomertieite, merenskyite, kotulskite, sperrylite and moncheite. In addition, electrum and Pt-Fe alloys have been identified by Sweeny (1989). The PGE mineralization is associated with chalcopyrite, pyrrhotite, pentlandite and pyrite interpreted by Watkinson and Dunning (1979) as being slightly metamorphosed equivalents of primary exsolution from monosulphide solid solution. Millerite and violarite

are present in secondary assemblages (Watkinson and Dunning 1979).

PGM in the Roby Zone are most commonly found with postcumulus hydrosilicates near silicate-sulphide interfaces (Talkington and Watkinson 1984). In this association, the PGM occur both within the hydrosilicates and at the sulphide-silicate interface.

PGE's also occur in the ultramafic rocks with sparsely disseminated to net-textured chalcopyrite, pyrrhotite, and pentlandite (<1 modal % sulphide). This type of mineralization is most commonly associated with websterite and gabbro-noritic phases of the northern and southern ultramafic centers.

Field and geochemical relations indicate that the EG and WG cumulates crystallized from a common magma and were subsequently emplaced as crystal mushes with the more fractionated and hydrous EG emplaced first. Lower concentrations of PGE, Au and S in the EG relative to WG suggests that sulphides were removed from the magma prior to the crystallization of the EG cumulates or that the parental magma was not saturated in sulphides at the time of cumulate formation. Magma mixing may have localized some sulphide mineralization along the EG/WG interface. However, because the phases are interpreted to have been emplaced as mushes, there was probably limited opportunity for sulphides to interact with magma. This would suggest a low R-factor and that this process may not be the explanation for the high grade mineralization.

Fe-rich pyroxene cumulate sheets intruded into the EG and WG are another possible source of PGE in the Roby Zone. The close spatial association of the sheets and mineralization suggest that the pyroxene cumulates are a likely source of PGE. As previously noted, the altered pyroxene cumulate sheet at the EG/WG contact contains particularly high-grade mineralization. Limited sampling of this group of rocks suggests that, in general, they have higher PGE contents than other samples of the ultramafic suite. Mixing of magmas which formed the pyroxene cumulates and the gabbro cumulates may have played a role in causing sulphide saturation in the pyroxene cumulates.

The association of PGM with secondary sulphides and altered silicates lead Watkinson and Talkington (1984) to suggest that fluid processes were active in mobilization and concentration of PGE in the Roby Zone. The role of fluids in PGE mineralized zones in the Bushveld and Stillwater Complexes is indicated by the occurrence of pegmatoids, volatile-rich phases and hortonolite replacement pipes (Todd et al 1982; Stumpfl and Ballhaus 1986; Schiffries 1982; Stumpfl and Rucklidge 1982). Fluid-bearing phases and fluid inclusions in both complexes suggest fluids that may be rich in Cl (Boudreau et al 1986; Stumpfl and Ballhaus 1986).

Several features of the Roby Zone indicate the importance of a volatile rich magma. These include: the association of PGM with postcumulus hydrosilicates near

silicate-sulphide interfaces; the association of PGM with pegmatitic phases and the mineralized breccia zones cored by pegmatitic gabbro which appear to have been fluid channels; and the highly fractionated Pd/Ir. Cl-rich apatite in mineralized gabbro suggest that the fluids associated with the mineralizing process were Cl-rich.

Brugmann et al (1989) have recently proposed a modification of this hypothesis and suggested that the volatile content of the residual magma triggered partial remelting of the gabbro cumulates in a process referred to as "constitutional zone refining".

Occurrences of PGE-bearing sulphides in the ultramafic sequence are hosted mainly by websterite at the top of cyclic units and by peridotites at the base of overlying units. These occurrences have not been studied in detail but may represent stratiform mineralization related to the interaction of fractionated magma with a new magma pulse in the ultramafic chamber (Sutcliffe and Sweeny 1986; Brugmann and Naldrett 1987). The presence of this mineralization, however, emphasizes that PGE's were concentrated at several stages in the fractionation of the LDIC magmas.

TIB GABBRO

Geology

The Tib Gabbro (Figure 5) is a layered intrusion with an area of approximately 25 km² located approximately 15 km northwest of the Lac des Iles Complex. The intrusion was mapped by Kaye (1966) and Smith and Sutcliffe (1987). The information summarized here is largely from Smith (in press). Petrological and geochemical studies are in progress and most of this summary is based in field observations. The gabbro is emplaced into an older suite of foliated to gneissic biotite tonalite which outcrops around the eastern, northern and southern margins of the intrusion. A younger suite of megacrystic granodiorite and granite intrudes the Tib Gabbro and older gneissic tonalites.

The Tib Gabbro is characterized by well preserved primary igneous mineralogy and moderately well developed layering. Layer attitudes indicate that the intrusion is funnel shaped and has been tilted to the northeast. The southwest part of the gabbro therefore represents the base of the exposed section. A gravity survey by Gupta and Sutcliffe (in press) supports this interpretation and models the 16 mgal gravity high associated with the intrusion as a tilted basin-shaped body.

The intrusion is divided into four zones by Smith and Sutcliffe (1987) which from the southwest to northeast are the Border, Lower, Middle and Upper Zones. The 50 to 150 m thick Border Zone is defined on a basis of the abundance of coarse grained to pegmatoidal gabbro which occurs around the northwestern, western and southern margins of the intrusion. The Lower and Middle zones probably represent distinct magma pulses which crystallized from the base upwards. These

zones record the crystallization of evolving cumulus assemblages which results in a mappable igneous stratigraphy with pyroxene rich cumulates near the base and plagioclase- and magnetite-rich cumulates near the top. The Upper Zone consists of amphibole-rich gabbro and is not well layered.

The 1.5 km wide Lower Zone consists of a sequence of layered gabbro with layers of orthopyroxenite and melagabbro (pyroxene cumulate) near the base. Above the orthopyroxenite the lithologies become more leucocratic and anorthosite appears as distinct layers. The top of the zone is marked by magnetite-rich gabbro with leucocratic layers. The 1.6 km wide Middle Zone is predominantly gabbro, however, leucocratic units are more abundant than in the Lower Zone. Conformable pegmatite layers and discordant dikes are common in the upper part of the zone. The top of the zone is again marked by magnetite-rich gabbro.

The Upper Zone is characterized by hornblende-rich, massive to weakly layered gabbro with cumulus apatite. The layers form a concentric pattern with a shallow inward dip. Leucocratic quartz-bearing granulite forms a discontinuous septa between the Upper and Middle Zones and the unit is interpreted to be metamorphosed tonalite but may alternatively be a final product of fractionation.

Layering in the Tib Gabbro is comparable to that developed in the Mulcahy Gabbro (Morrison *et al* 1986). Planar lamination of feldspars and pyroxene is found throughout the intrusion. Modal layering defined by changing compositions of cumulus minerals is best developed in the upper part of the Lower Zone and occurs in both cyclic and intermittent layer sequences. Distinctive modal layers include: orthopyroxenite (orthopyroxene cumulate); websterite (clinopyroxene - orthopyroxene cumulate); anorthosite (plagioclase cumulate) and magnetite-rich cumulate. The modal layers are usually tens of centimetres in width and persist for tens of metres along strike.

Mineralization

The Tib Gabbro is the second largest of the mafic intrusions in the lac des Iles area and is an exploration target for PGE and Cu-Ni mineralization. Intermittent exploration on the intrusion since 1966 has outlined a number of areas of disseminated Cu-Ni sulphide mineralization in the gabbro, some of which contain associated PGE. To date, the most significant mineralization that has been reported is at the Kuhner Occurrence in the Border Zone along the south margin of the intrusion. At this location, net textured sulphide mineralization is present in coarse grained gabbro and assay values up to 190 ppb Pt, 390 ppb Pd, 120 ppb Au, 1140 ppm Ni and 710 ppm Cu are reported by Smith (in press)

GRANITOIDS AND RELATED ROCKS

Tonalite

Gneissic to strongly foliated biotite tonalite gneiss is the oldest intrusive phase in the Lac des Iles area and is the host rock into which most of the younger plutons are emplaced. Discordant intrusive relations between the mafic-ultramafic intrusions and tonalitic gneiss are locally developed on the margins of intrusions such as the Tib Gabbro, where gabbro apophyses cross-cut the fabric of the gneiss and tonalitic gneiss enclaves occur in the gabbro.

Two late granitoid plutons occupy the center of the circular structure defined by the mafic intrusions. These plutons also discordantly intrude the tonalitic gneiss and contain tonalitic gneiss enclaves. The plutons include a hornblende-bearing tonalite with an area of 150 km² and a younger, foliated to massive biotite tonalite to microcline megacrystic granodiorite pluton which locally contains enclaves of the hornblende-bearing tonalite. The hornblende tonalite is intruded by numerous mafic to intermediate dikes and exhibits textures indicating the coexistence of mafic and felsic magmas (Sutcliffe 1989). These textures are the focus of the field trip stops in the granitoid rocks.

The association of early tonalitic gneisses and later massive plutons is common in granitoid terranes of the Western Superior Province (Schwerdtner et al. 1979).

Modal analyses of representative tonalites and other samples are shown in figure 6. The tonalite has a medium- to coarse-grained, hypidiomorphic texture and is massive to weakly foliated. The presence of coarse (1 to 2 cm), blocky to prismatic hornblende is a conspicuous feature of the pluton.

The major mineral phases are: 2 to 4 mm subhedral, tabular plagioclase of composition An₂₈₋₃₄ (40-58%); anhedral, weakly strained quartz (29-38%); 0.5 to 2 mm brown-green pleochroic biotite (7-16%) and subhedral to prismatic, 2 mm to 2 cm, green pleochroic hornblende (1-15%). Accessory phases are sphene, apatite and zircon. Plagioclase locally displays weak to moderate oscillatory zoning and minor alteration to sericite and epidote. Although hornblende is locally subordinate to biotite, this lithological unit is referred to as hornblende tonalite throughout the paper to distinguish it from the younger biotite tonalite in which hornblende is absent.

Mela-tonalite containing up to 60% hornblende occurs as irregular patches and intrusive dike-like bodies up to several meters in extent which grade into normal tonalite. Mela-tonalite is the major constituent in 2 to 3% of outcrops in the hornblende tonalite pluton. Hornblende in mela-tonalite is typically coarse, occurring as grains up to 2 cm of euhedral, prismatic to skeletal habit. Textural gradations between skeletal hornblende and coarse blocky hornblende are present. The mela-tonalites have complex

internal structures characterized by comb-layering and colloform structures of mafic material in felsic host.

Skeletal hornblende is best developed in mela-tonalite adjacent to mafic-felsic interfaces. Euhedral skeletal hornblende grains also occur as wispy schlieren in hornblende tonalite and in hornblendite units of mafic intrusions. The skeletal grains have hollow cores containing biotite or biotite + plagioclase. These assemblages probably reflect reaction between amphibole and trapped liquid.

Large rounded hornblende aggregates as large as 2 cm in diameter locally display a proto-orbicular texture according to the classification of Leveson (1966). These are present in tonalite adjacent to mela-tonalite and mafic dikes. The composite grains have central areas of pale-green pleochroic hornblende intergrown with quartz and feldspar producing a sieve texture. Their rims are composed of dark-green pleochroic grains of interlocking prismatic hornblende.

A breccia pipe, approximately 300 m in diameter, is emplaced into the western end of the hornblende-bearing tonalite pluton. The pipe consists of sub-angular hornblende-tonalite fragments and minor hornblendite fragments set in a matrix of fine-grained, comminuted and silicified tonalite. Tonalite fragments within the breccia are of the same composition and texture as the host rock but contain 2 to 3 cm thick silicified, sericitized and epidotized rinds.

Mafic dikes and hybrid textures

A wide range of textures indicative of contemporaneous mafic and felsic magmatism is developed associated with the emplacement of mafic to intermediate dikes in hornblende tonalite. The dikes range from parallel-sided intrusions, 1 to 2 m wide, with sharp contacts which cross cut foliation in the tonalite, to disaggregated linear inclusion swarms in which the mafic magma appears to have lost internal contiguity. These structures are similar to textures indicative of contemporaneous mafic-felsic magmatism and mixing reported by Blake (1981), Marshall and Sparks (1984), Furman and Spera (1985) and Hyndman and Foster (1988). Mafic dikes are observed in approximately 15% of the outcrops of the hornblende tonalite pluton.

The dikes are generally fine-grained, lack chilled margins and are typically back-veined by tonalite which forms wispy nets separating rounded centimeter- to decimeter-sized mafic globules. This texture is described as net-veining by Marshall and Sparks (1984).

With increased veining the net-textured dikes grade into disaggregated dikes in which the mafic to intermediate rock forms rounded and elongate bodies which resemble pillows of extrusive rocks in the host tonalite. In contrast to other descriptions of similar structures in Tertiary intrusives (Vogel 1982; Brown and Becker 1986;

Marshall and Sparks 1984), chilled rinds on "pillows" at Lac des Iles are generally absent.

Mineralogically the dikes consist of moderately fresh 0.5 mm laths of plagioclase (andesine An₄₀₋₄₆) and 1 mm grains of prismatic hornblende to pale green fibrous amphibole, and minor biotite, quartz, apatite and opaques. The mafic to intermediate dikes locally display discontinuous hornblende-rich reaction rims adjacent to host tonalite. In addition, mela-tonalite to hornblendite phases are generally associated with zones of dike emplacement.

Magma Mixing

Field evidence suggests that mixing between magmas has taken place. Geochemical data does not readily constrain the mixing process due to complications of fractionation of mafic end members and the presence of multiple components. The most likely magma chamber configuration during the emplacement of the late tectonic plutons is that of a zoned chamber with mafic magma underplating felsic magma (Figure 7). This schematic configuration has been supported by gravity studies (Gupta and Sutcliffe in press).

An important aspect of the interaction between the mafic and tonalitic magmas at Lac des Iles is the development of hornblende-rich cumulus phases. The skeletal texture of these phases suggests they may have formed in a supercooled magma during mixing.

Field relations and geochemical considerations suggest that the open system magmatic processes at Lac des Iles resulted in a continuum of disequilibrium textures, ranging from hybrid tonalite magmas to late net-veined mafic and intermediate dikes. These textures reflect emplacement and mixing of basic magma with varying degrees of fractionation into felsic magma throughout a large interval of the cooling history of the felsic magma. Furman and Spera (1985) have suggested that features of mixing associated with injection of mafic magma into a Sierran pluton represent a continuum of mixing states which can largely be related to host crystallinity. When the host magma has low crystallinity (<30%) convective velocities are high and mixing features occur on a scale of centimeters or less. At approximately 30-70% host crystallinity, mesoscopic schlieren, mafic inclusions and inclusion swarms are preserved. At higher crystallinity, outcrop scale textures such as dike swarms and large inclusion swarms are evident. These stages correspond to the observed textures at Lac des Iles. Net-veined dikes and other conspicuous mixing features at Lac des Iles probably formed at the latest stages of mixing when the efficiency of mixing is reduced but outcrop evidence for the coexistence of two magmas is enhanced. Breccia zones within the tonalite containing fragments of tonalite and mafic rocks with thick alteration selvages provide evidence of explosive degassing of the chamber.

In an evaluation of the situation of a zoned mafic-felsic chamber, Rice (1985) has shown that mixing of a

convecting, density stratified column is possible given large dimensions and a reasonable thermal gradient. An important aspect of a zoned chamber is the probability of roll over and explosive degassing if the magmas contain volatiles (Rice 1985). Evidence for degassing, mixing and violent emplacement of mixed chambers has also been documented by Marshall and Sparks (1984). Density instability of mafic magmas at the base of a chamber may be established by vesiculation of mafic magma and can lead to rapid transfer of mafic magma across a formerly stable mafic/felsic magma interface (Eichelberger 1980; Rice 1985).

Several of the associations reported in this study are common to other granitoid terranes in the Superior Province and suggest the model presented here may have wide applicability to late granitoids, particularly those containing hornblende. Xenoliths ranging in composition from hornblendite to microdiorite are a particularly widespread feature in late Archean plutons (eg. Davis and Edwards, 1985). These inclusions have conventionally been interpreted as pre-existing mafic rocks. However, as this and other studies (eg. Didier 1987) have suggested the inclusions may be chilled mantle-derived magma. Other features of contemporaneous mafic and felsic magmatism reported elsewhere include the presence of net-veined dike rocks in tonalitic to granodioritic plutons (Davis and Edwards 1985). Recent precise U-Pb zircon geochronology also indicates a close temporal association of late Archean gabbro plutons with granitoid rocks in other parts of the Wabigoon Subprovince (Morrison et al. 1985). These features suggest that mantle-derived mafic magmas are an important aspect of late Archean granitoid magmatism.

FIELD TRIP GUIDE

The field trip is designed to be completed in one day, including travel from and return to Thunder Bay. The trip includes 1 stop on gneissic tonalite host rock, 3 stops on the Tib Gabbro, 4 stops to examine the granitoids, and one stop at the PGE and Cu-Ni mineralization at the Roby Zone of Madelaine Mines Limited.

Due to logging and mineral exploration activity in the field trip area, the road conditions are continually changing. If you are doing this trip on your own you will need an up-to-date map of logging roads in the area.

Proceed west from Thunder Bay along highway 11/17. At Shabagua, turn north on highway 17 towards Dryden. Turn north on the Dog River Road (also signposted to Great Lakes Pulp and Paper Company Camp 234) which is approximately 10 km west of the townsite of Raith and 1 km west of the Central Time Zone/Eastern Time Zone marker. Begin recording mileage at the start of the Dog River Road. Drive carefully as loaded trucks are using this road! Proceed north on the Dog River Road for 21 km, at which point there is a major junction with a branch road from the east. The first outcrop is the large exposure on the southeast side of branch road approximately 50 metres from the junction.

Stop 1 - Gneissic tonalite

This outcrop has not been studied in detail but is included as a stop to illustrate the host lithology into which the mafic-ultramafic and granitoid intrusions are emplaced. As is typical of other areas of gneissic tonalite in northwestern Ontario, the outcrop is complex and several granitoid phases are present.

Continue north on the Dog River Road, pass through camp 234, at 40 km (distance from highway 17) turn northeast (right) at the major fork. Proceed northeast for 6.2 km and then turn east along branch road to Tib Lake. Proceed east along road to Tib Lake for 2.7 km at which point there is a small track that leads to the north. Stop 2 is on the west side of the track approximately 100 metres north of the Tib lake Road.

Stop 2 - Tib Gabbro Border Zone

The Border Zone of the Tib Gabbro is defined based on the abundant pegmatitic gabbro that occurs around the base of the intrusion. This well-exposed outcrop displays the typical textural variation in this zone. Proceeding upward from the lowermost exposed part of the outcrop textures range from coarse-grained and pegmatitic gabbro to medium grained gabbro and gabbro-norite with conformable pegmatitic layers.

Return to Tib Lake Road and proceed east for another 1.2 km. Stop 3 is on the north side of the road and is in the bush approximately 10 metres from the road.

Stop 3 - Tib Gabbro layering

Layered gabbro norite in the upper part of the Lower Zone of the Tib Gabbro. Well developed intermittent layering locally shows modal grading with pyroxene-rich bases and plagioclase-rich tops. Uralitic alteration (green coloured) occurs along fractures which locally cross cut the primary layering.

Return to Tib lake Road and proceed east for another 0.3 km. Stop 4 is on the south side of the Tib Lake Road, approximately 20 metres from the road, and is on the east side of a culvert.

Stop 4 - Tib Gabbro Lower Zone

This outcrop displays wispy layered gabbro norite near the top of the Lower Zone. The wispy layers consist of plagioclase or pyroxene rich bands in gabbro norite. Thin pyroxenite dikelets of various orientations cut the layering and contain inclusions of host gabbro norite. The dikelets are interpreted to be injections of cumulate material. A small hornblende-feldspar dikelet is also present. Green uraltic alteration of pyroxenes is associated with fractures and is a late feature.

Return to the Tib lake Road and continue east for another 3.2 km, past the outlet from Tib Lake, to a road which runs down the east side of Tib Lake. Turn south along this road and proceed for 11 km to the junction with another major logging road which is referred to here as the "Lac des Iles Road". Take this road to the southwest (in the direction of the Dog River Road) for 6.5 km, until the junction with the Roenicke Lake Road is reached. Take the Roenicke road north for 1.8 km and turn west along an unnamed bush road. Proceed along this road, taking the left fork after 2.1 km and reaching the outcrop for stop 5 after 5.0 km. Note: The bush road into this stop is becoming overgrown and will probably be difficult to recognize if you are following this after the spring of 1990.

Stop 5 - Tonalite breccia

This distinctive breccia zone with dimensions of approximately 100 by 300 metres cross-cuts the hornblende tonalite east of the Dog River. The breccia consists of subrounded fragments of tonalite with 2 to 3 cm thick sericite-epidote-silicified alteration rinds. The matrix consists of altered, fine grained and locally comminuted tonalite. Rounded irregular mafic clots up to 0.5 metres long are locally present interstitial to the tonalite

fragments. Several quartz-feldspar porphyry dikes intrude the breccia.

The breccia is interpreted to be the product of an explosive degassing of a magma chamber. The process is thought to provide further evidence for underplating of granitic magma by mafic magma. Explosive devolatilization is an expected consequence of the cooling of such a system if the mafic magma contains volatiles (Rice 1985).

Return to Roenicke Road and then proceed south to the "Lac des Iles Road". Turn southwest along the "Lac des Iles Road" (toward Dog River Road) and proceed for 6.5 km to the Garden Road. Turn east on the Garden Road and proceed for 1.8 km. Stop 6 is an outcrop on the south side of the road.

Stop 6 - Net-veined dikes

Mafic dikes are abundant in this area of the tonalite and range from intrusions with sharp contacts to dikes which are extensively veined by tonalite, such as the dike in this outcrop. The back veining texture is known as net-veined since the tonalite forms a wispy net which surrounds rounded centimetre- to decimetre-sized mafic globules. The structure is indicative of emplacement of the mafic magma prior to the consolidation of the felsic host.

Proceed for another 2.2 km east on the Garden Road. Stop 7 is a group of outcrops on either side of the road.

Stop 7 - Mixing textures

In this series of outcrops, mixing of mafic and tonalitic magmas has resulted in a variety of igneous textures. Net-veined and disaggregated mafic dike material is present in the lower part of the outcrop. The disaggregated mafic material could be readily confused with rounded mafic inclusions. Adjacent to the net veined dike, the foliation of the tonalite is destroyed by remobilization of the felsic material. At the top of the outcrop, the hornblende-rich mela-tonalite is interpreted to be a hybrid composition resulting from more complete interaction of the mafic and felsic magmas. Abundant skeletal textures in this rock may result from supercooling due to magma mixing.

Turn around and proceed west for 0.7 km. Stop 8 is an outcrop on the north side of the road.

Stop 8 - Hornblendite

Hornblendite occurs as outcrop scale pods within the tonalite and are gradational with the mela-tonalite. The hornblendite and mela-tonalite have complex internal structures such as comb-layering and colloform structures. Comb-layering is defined by the alignment of coarse acicular to skeletal hornblende which is oriented perpendicular to

layering. These rocks have geochemical characteristics which suggest that they are cumulates derived from the mafic dike suite. The hornblendites also resemble a suite of rocks known as "appinites" which are described in some Phanerozoic shoshonitic plutonic suites.

Return to the "Lac des Iles Road" and turn back to the northeast towards Lac des Iles. Proceed for 11 km to the road on the right which provides access to the Madelaine Mines property. Turn east (right) towards the Madelaine Mines Property, taking a left fork after 3.2 km and another left fork after 6.1 km. At the Madelaine Mines Property there is a gate and we will be met by staff from the company.

Stop 9 - PGE mineralization and geology in the vicinity of the Roby Zone - Madelaine Mines Limited

Note: Due to the ongoing exploration and development work on the Roby Zone, which includes extensive stripping and blasting, it is not possible to plan the precise features that can be examined at this location prior to arrival at the site.

Maps of the geology of the gabbroic units in the vicinity of the Roby Zone (Figure 8) show the complex relations between intrusive units. Figure 3 shows that the Roby Zone occurs along the contact between uniform EG and the more complex WG. Sheets and minor discordant dikes of pyroxene cumulate intrude the EG and the EG/WG contact. The sheets and dikes which are gabbro to websterite in composition, contain cumulus orthopyroxene + clinopyroxene and intercumulus plagioclase and vary from altered to fresh, even where hosted by altered EG. The highly uraltized pyroxenite is termed amphibolite in drill logs. The altered pyroxenite sheet is approximately 5 to 10 m thick and intrudes the EG but appears to grade into gabbro of the WG and is approximately conformable with the EG/WG contact.

Figure 8 is a detailed outcrop map from the "C-Zone". This zone is within the WG at the southern end of the Roby Zone and illustrates the complexity of the WG unit. The outcrop is predominantly fresh to moderately altered gabbro of the WG with several units of altered leucogabbro which form inclusions and layers within the gabbro. Some of the inclusions have cusped surfaces and gradational contacts suggesting that they were not solid at the time of incorporation. Also present within the gabbro are inclusions of pegmatitic gabbro, altered pyroxenite, and fine-grained, recrystallized, clinopyroxene amphibolite. The gabbro and leucogabbro are discordantly intruded and disrupted by mineralized gabbro pegmatite dikes and a 1 to 4 m wide mineralized intrusion breccia. The breccia is defined by an abundance of gabbroic

inclusions and contains a core of gabbro pegmatite. (Note: Unfortunately this outcrop is now partially destroyed by blasting.)

Textures in the WG, such as those shown in Figure 8, suggest that some type of incomplete mixing or contamination of the WG by the EG took place. The presence of rounded inclusions, particularly of EG in WG and complex interfingering of the two gabbro types on a decimeter scale suggests that the EG was not completely solidified when the WG was intruded. Contacts between the two gabbro phases are commonly cusped, net-veined and disaggregated. These textures are similar to those described in mixing zones between mafic and felsic magmas by Marshall and Sparks (1984) and would be reflected in the chemistry of the rocks only by more detailed sampling than that done in this study. The limited development of modal layering, the presence of cumulate textured dikes of WG intruding EG and the abundance of inclusion-rich outcrops suggests that the cumulates did not form in-situ but were emplaced as crystal mushes. This situation contrasts with layered intrusions that crystallized by upward and inward growth of a cumulate-liquid interface (eg. Irvine et al 1983).

End of trip. Return route to Thunder Bay will be by the recently opened access route to Lac des Iles from highway 527. This is a shorter trip than returning via the Dog River road.

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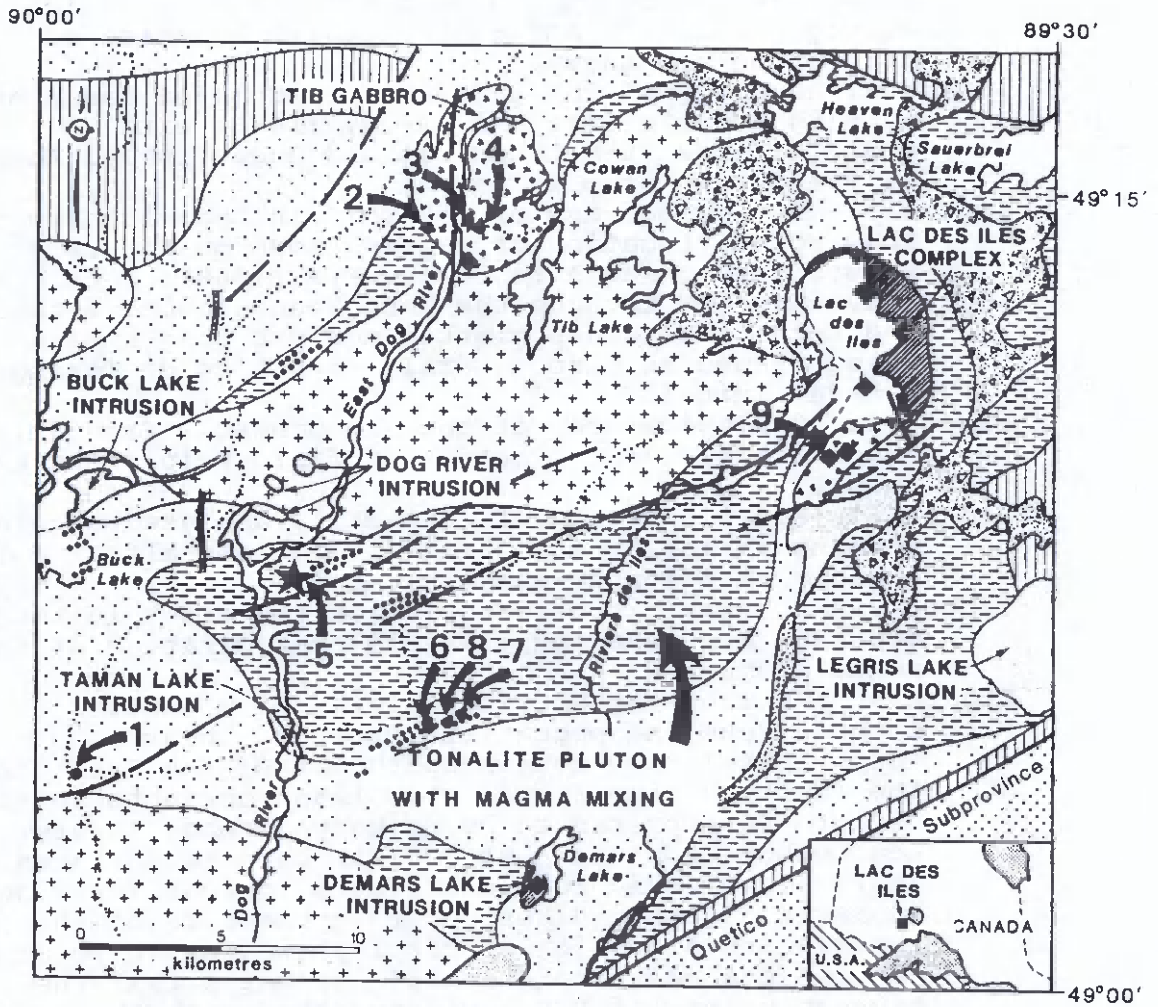
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PROTEROZOIC

diabase

ARCHEAN

Late Granitoids

biotite tonalite

hornblende tonalite

Late Mafic to Ultramafic Rocks

mafic to intermediate dikes

ultramafic

gabbro to gabbro-norite

hornblende gabbro, hornblendite

hornblende diorite

Early Granitoid Rocks

tonalitic gneiss

Supracrustal Rocks

mafic metavolcanics

metasediments

fault

PGE occurrence

contact

breccia zone

road

Figure 1.

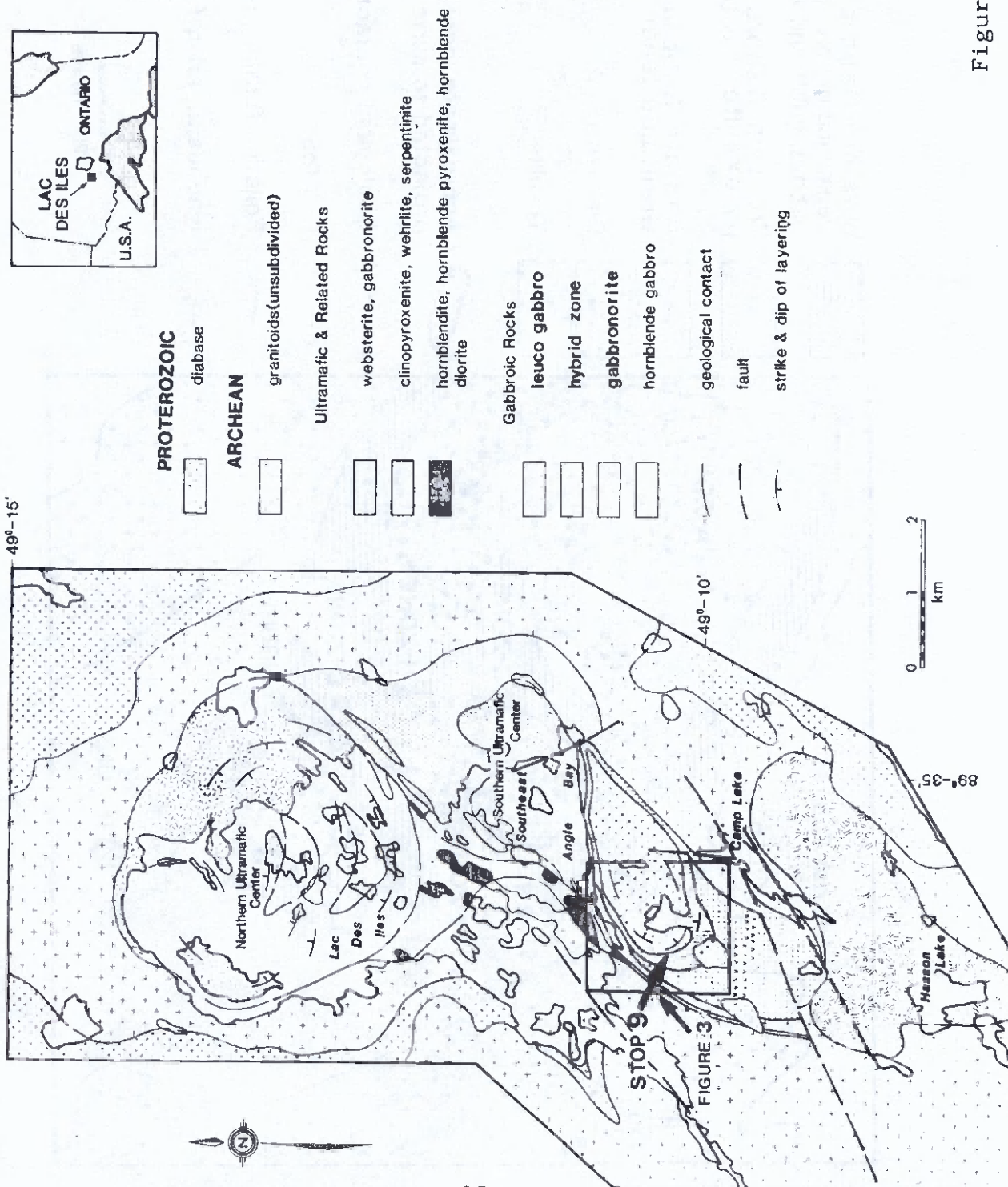


Figure 2.

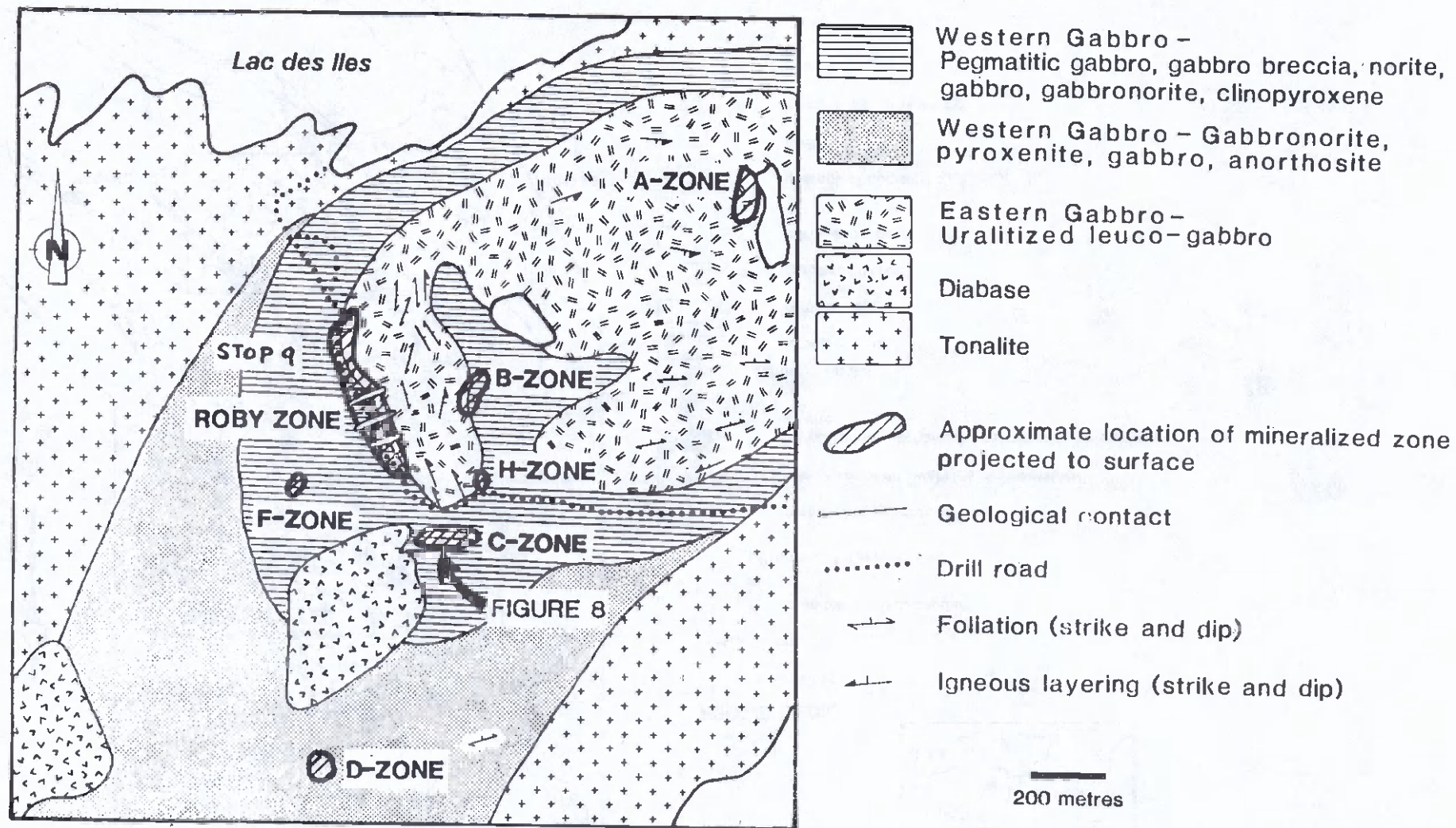


Figure 3.

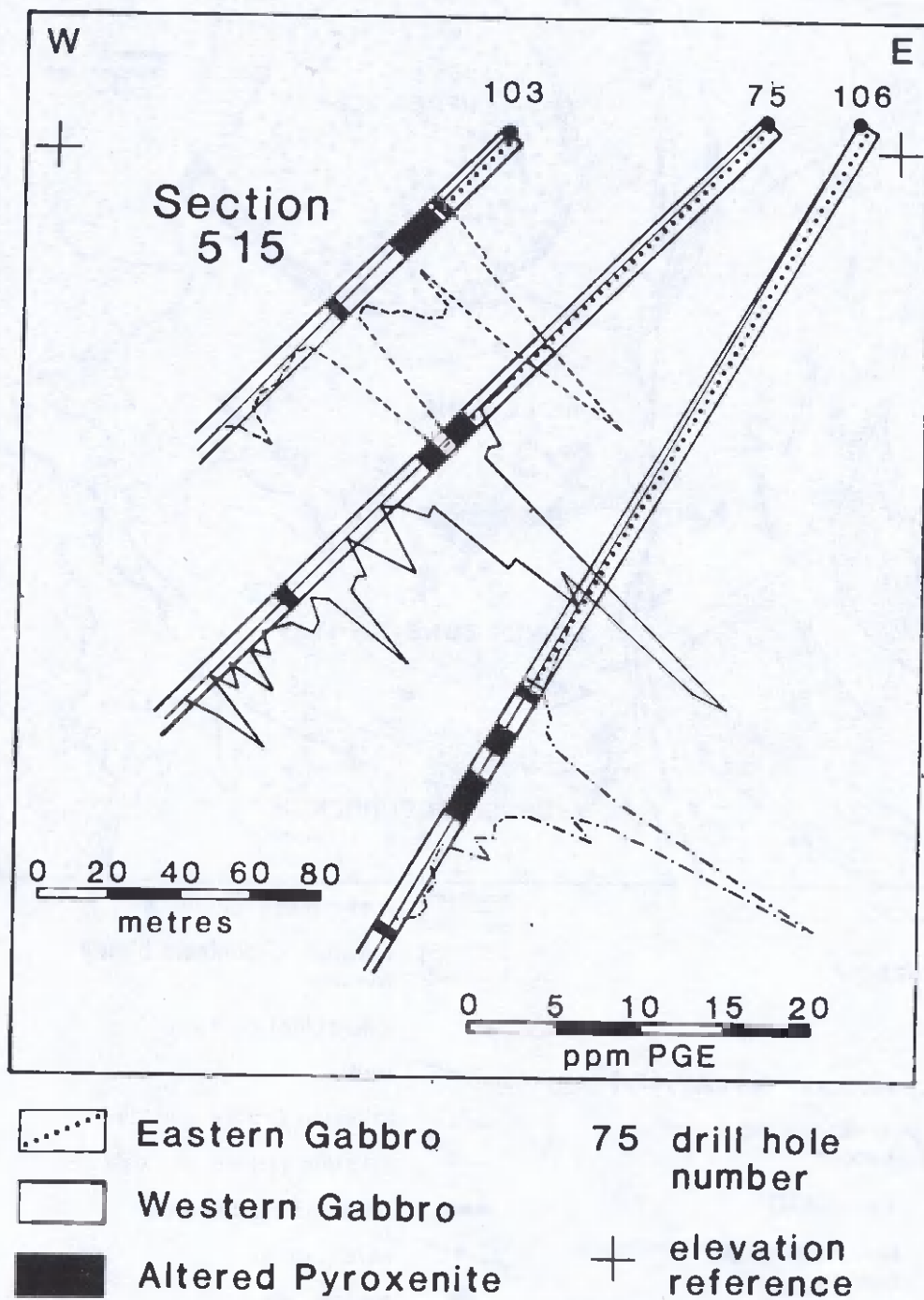
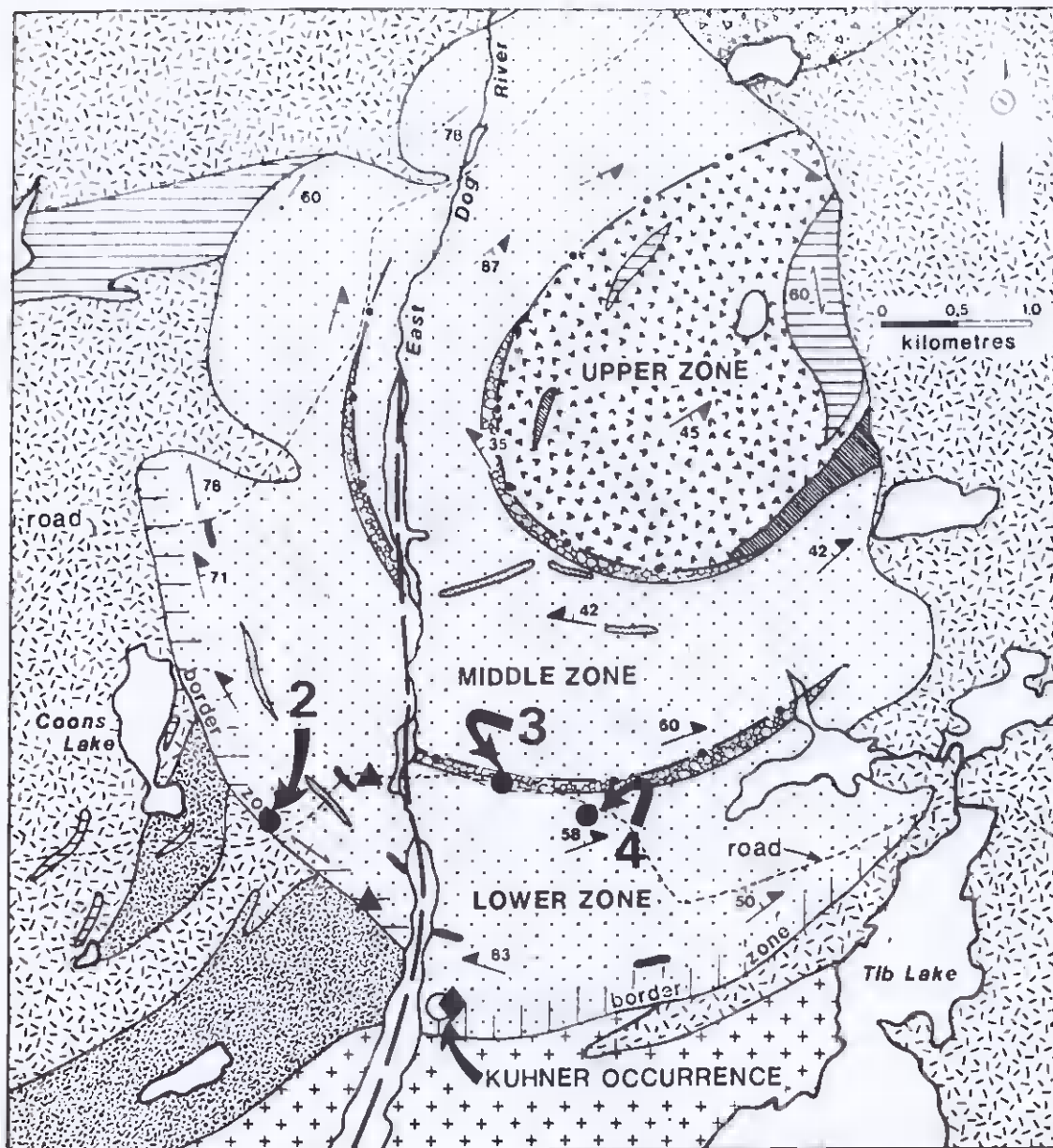


Figure 4.



PROTEROZOIC

diabase

ARCHEAN

granodiorite (megacrystic)

granodiorite, Hb tonalite (foliated)

TIB GABBRO

hornblende + quartz ± biotite gabbro

magnetite-rich gabbronorite

pegmatitic gabbro

quartz granulite

gabbro norite, norite

mafic metavolcanics

foliated to gneissic biotite tonalite

lithological contact

fault

layering (strike and dip)

foliation (strike and dip)

ultramafic cumulates

zone contact

trench site

sulphide mineralization with > 50ppb PGE

sulphide mineralization with > 500ppb PGE

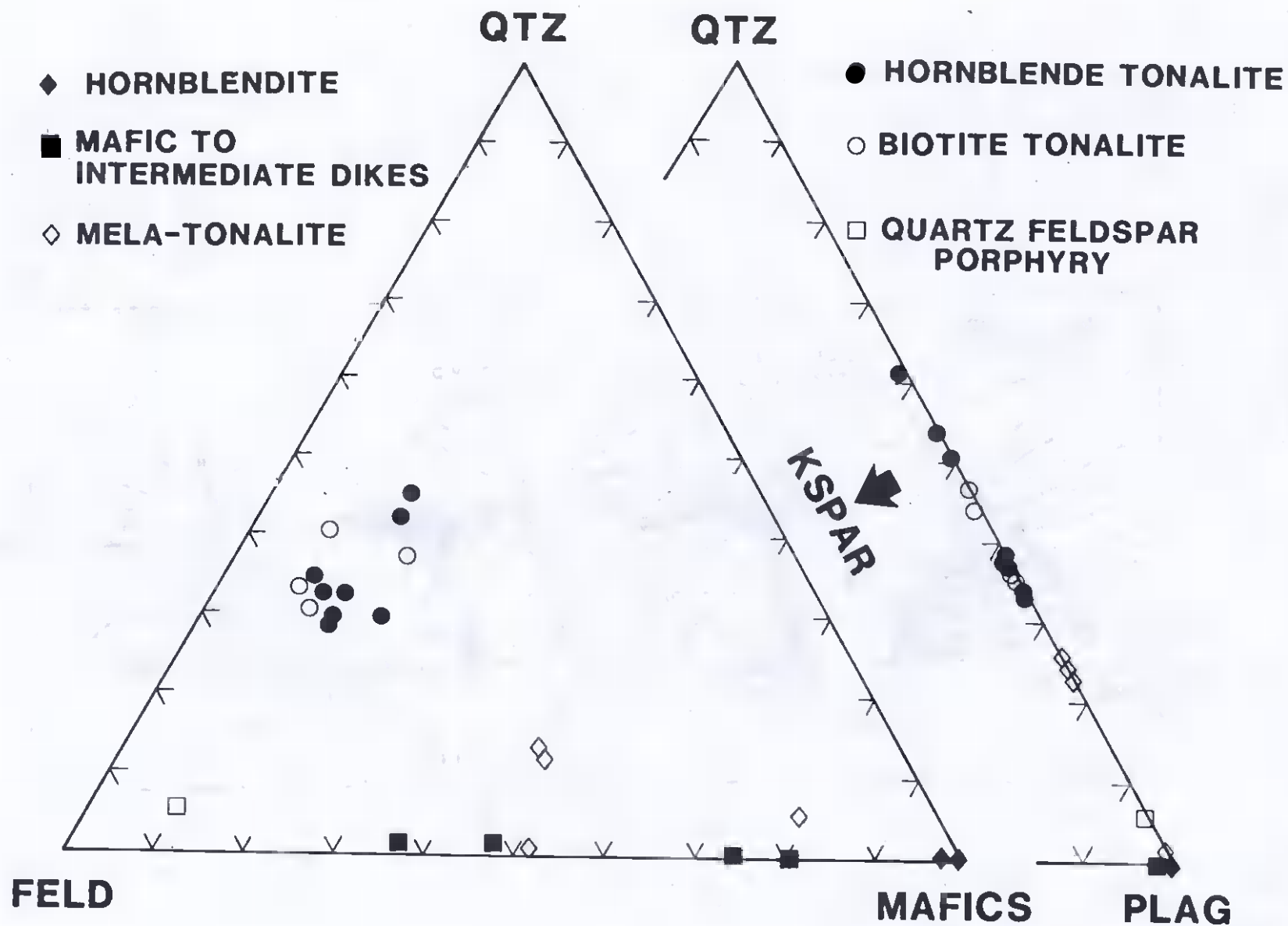


Figure 6.

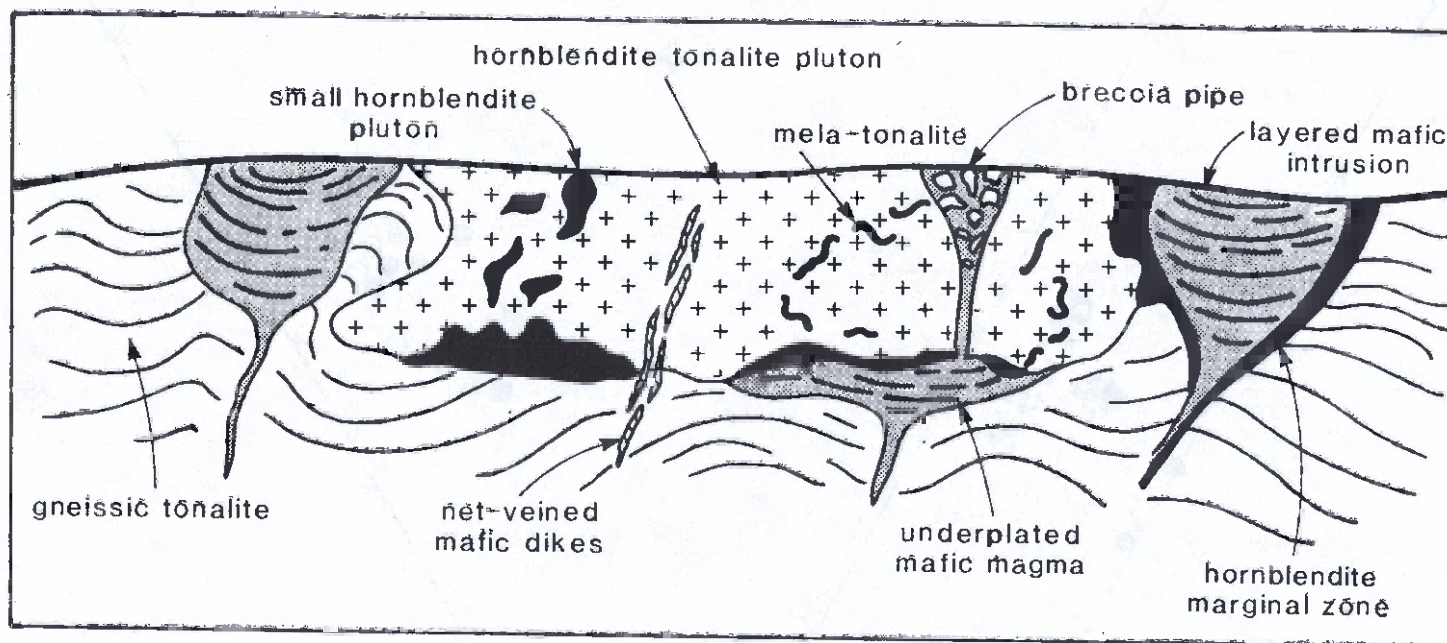


Figure 7.

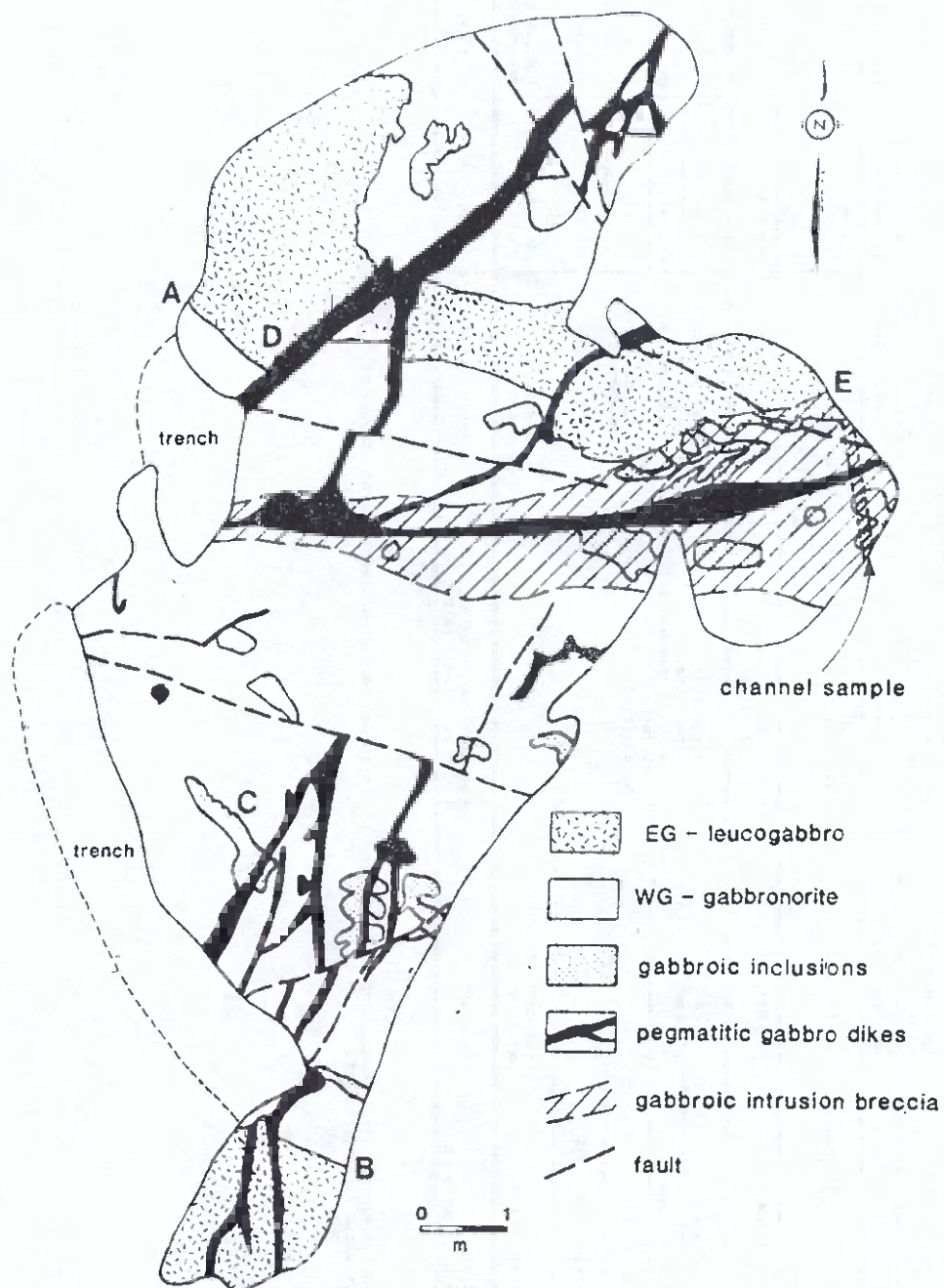


Figure 8.

Table 1. Summary of Petrography of Lithologies from the Lac Des Iles Complex

Rock Type	Cumulus Minerals	Intercumulus Minerals	Alteration	Texture	Grain Size	Crystallization Sequence
Serpentine, weirite	ol, ol + chr	cpx, minor hb	ol to serp + mgt, cpx to talc + anthophyllite, cpx to act, minor carbonate, epidote, chlorite, clinzoisite	mesocumulate	1-4 mm	ol + chr, ol + cpx, cpx, cpx + opx
Ol Clinopyroxenite, Clinopyroxenite	ol + cpx, cpx	cpx, cpx, hb, minor phl		accumulate to mesocumulate	1-4 mm	
Websterite, Gabbro	cpx + opx	cpx, cpx, hb, pl, mgt		mesocumulate	1-3 mm	cpx + opx + pl
Lherzolite, Ol Websterite dike	chr + ol + cpx + opx	pl, hb, phl	as above	mesocumulate to orthocumulate	1-3 mm	ol + chr, opx + cpx
Ol Gabbro dike					1-5 mm	opx + cpx + pl + hb
Hornblende, Pyroxene Hornblende	hb, hb + cpx	pl, qtz, bio	epidote, chlorite, clinzoisite	hypidiomorphic	3mm - 1cm	cpx + hb, cpx + hb + pl + qtz + bio
Norite	ol + pl + opx				2-4 mm	ol + pl, pl + opx,
Gabbro	pl + opx + cpx	pl, cpx, hb, bio	uralite, talc anthophyllite, epidote, clinzoisite iddingsite, sericite, chlorite	mesocumulate to orthocumulate, subophitic to isogranular	2-4 mm	pl + opx + cpx
Leucogabbro, Anorthositic Gabbro	pl, pl + cpx + hb	pl, cpx, hb, act			2-4 mm	pl, pl + cpx + hb,
Magnetite-rich Gabbro	pl + cpx + hb + mgt				2-4 mm	pl + cpx + mgt
Pegmatitic Gabbro and Gabbro	pl + cpx, pl + cpx + opx	pl, cpx, hb, act, qtz, mgt		coarse grained	2mm - 6cm	similar to gabbro and gabbro
Hornblende Gabbro	pl + hb, pl + hb + mgt + ap	bio, qtz	uralite, chlorite epidote, clinzoisite	hypidiomorphic	2-4 mm	pl + hb, pl + hb + mgt + ap

Abbreviations: ol=olivine; chr=chromite; cpx=clinopyroxene; opx=orthopyroxene; phl=phlogopite; bio=biotite; act=actinolite; qtz=quartz; hb=hornblende; pl=plagioclase; mgt=magnetite; ap=apatite;

FIELD TRIP 2

I.L.S.G. FIELD TRIP

Geology of the Shebandowan & Quetico Archean subprovinces

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Proterozoic Rocks

1. Kakabeka Falls. Recessed post-glacial gorge in Proterozoic shale. Soft sediment deformation in Proterozoic shales - car park exposure.
2. Basal conglomerate of Proterozoic sequence. (Care on Highway!) (Kakabeka Formation).

Archean Rocks

3. Pillow lavas and graded tuffs. Structural facing. D₁ fabric. (Fig. 1a, bc).
4. Fourway School. Younger "Timiskaming-like" sequence of Archean. Flap of downward-structural facing sediments. (Fig. 3).
5. Finnmark exposure. "Timiskaming-like" sediments. Intertidal sequence. (General structural style of area, see Fig. 2).
12. Quetico gneisses, Raith. Migmatites, diktyonitic structure, multi-"phase" deformation.
11. Quetico belt gneisses.
6. Shebandowan Mine Road.

"Timiskaming-like" sequence

- (a) Fragmental red igneous rock (!?)
- (b) conglomerate. Fabrics and strain in region (see Fig. 5,6,7).
- (c) slate

Older Keewatin-type Archean rocks

- (d) pillow lava
 - (e) iron formation
7. Swamp Creek Pillow lavas. Strain (see Fig. 1,2)? Younging? Structural facing?
 8. Tuffs and pillow lavas. (Structural facing and folds? see Fig. 8).
 9. Increasing metamorphic grade in metasediments (Kashabowie group) towards belt boundary.
 10. Gneissose rocks of the Quetico Belt.

ARCHEAN LITHOLOGIES



PARAGNEISS



GRANITOIDS



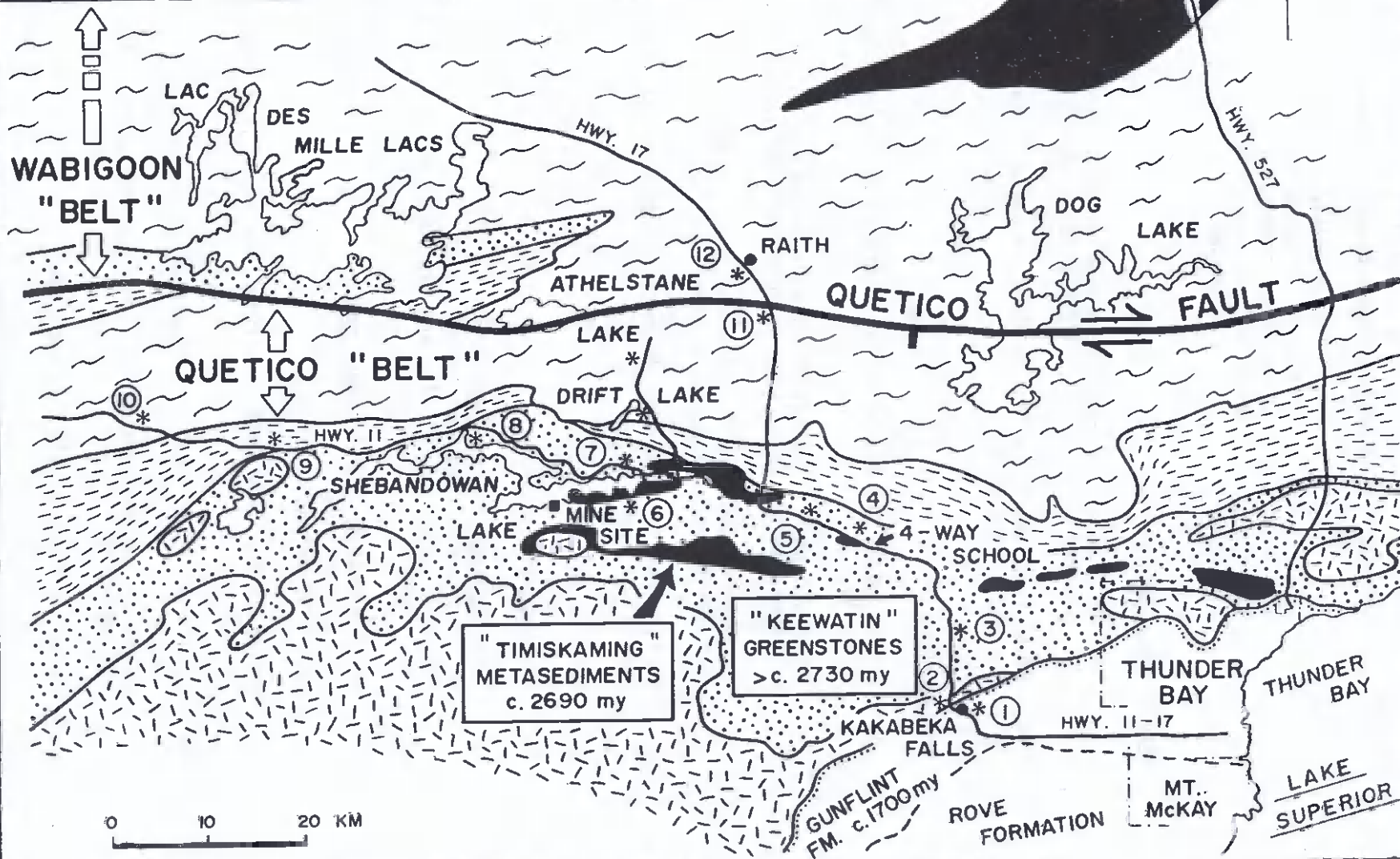
GREENSTONES

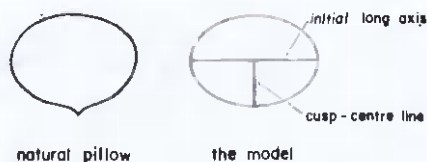


"TIMISKAMING" METASEDIMENTS

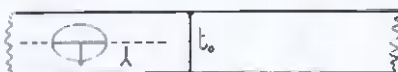


GREYWACKES

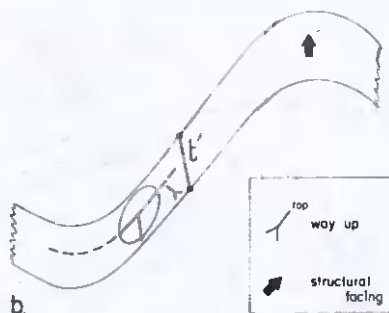




An idealised natural lava pillow before straining and the geometric model of it which is used in this paper.



a.



b.

Fig. 1a

(a) Undeformed bed with idealised pillow. The cusp-to-centre line of the pillow gives the younging direction. (b) Deformed bed with idealised pillow. The cusp-to-centre line of the pillow now tracks the 'stratigraphic thickness' direction, t' . This may approximate to the structural facing direction. The simple younging direction is no longer perpendicular to the long axis of the pillow in the general case, but, by definition, it remains perpendicular to bedding.

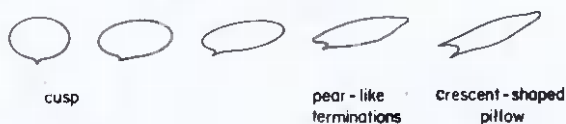


Fig. 1b

undeformed pillow deformation $\left[\sqrt{(\lambda_1/\lambda_2)} > 4.5 \right]$ crescent-shaped pillow

Schematic sequence of progressively deformed pillows inferred from field observations. When the finite strain ellipse ratio reaches about 4.5 the pillows develop pear-like terminations as a result of heterogeneous strain, and subsequently the ends of the pillow become so pointed that it is difficult to identify the cusp

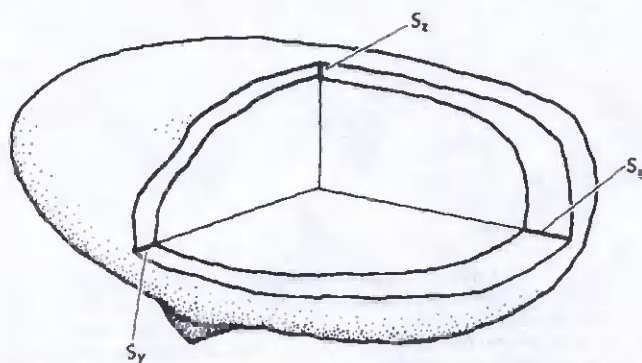
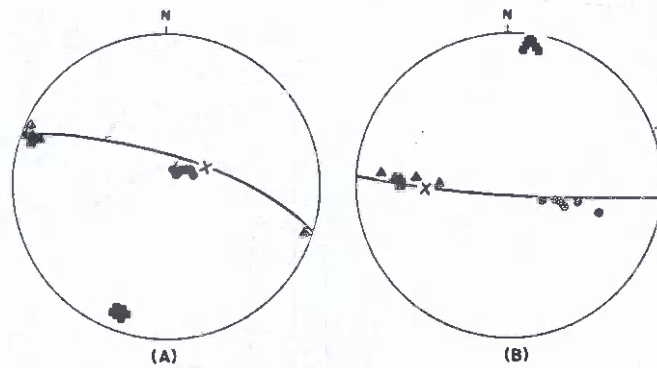


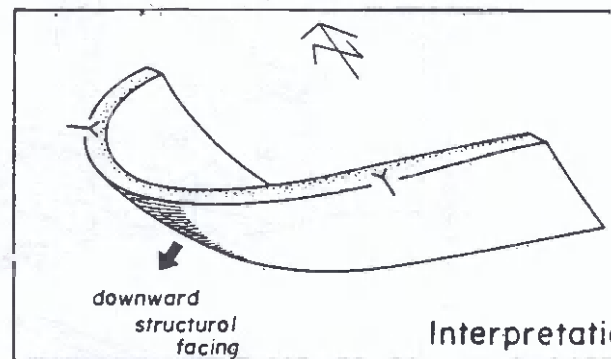
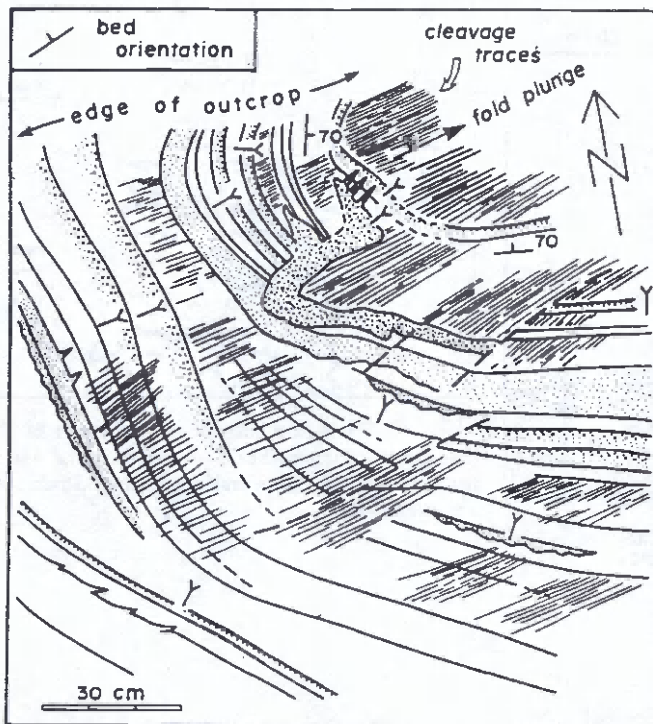
FIG. 1. Sketch of a deformed pillow showing the selvage thicknesses in the principal directions of strain: S_x , S_y , S_z . The ratios of these thicknesses give the ratio of the principal strains if the thickness of the selvage was originally constant. It should be noted that the principal selvage thicknesses are not normally in the same ratio as the deformed pillow's dimensions: one exception is the unlikely event that the original, undeformed pillow was spherical.

Fig. 1c



Susceptibilities : ● maximum ▲ intermediate ■ minimum
X: mean fold plunge / intersection lineation

Fig. 2 Lower-hemisphere equal-area stereonet showing the mean cleavage orientation (great circle) and mean intersection lineation for (A) the Shebandowan mine area and (B) the eastern Finmark area (east of Fourway School) (see Fig. 1). The principal magnetic susceptibility orientations for the slates are also shown.



Plan of an outcrop surface on the south side of Highway 11-17, about 100 m west of Finmark Road.

Where the glacial polish has been weathered away, the vertical cleavage is pronounced; elsewhere the sedimentary details are preserved. Ripple marks, graded bedding, and sand intrusions indicate that the westward-closing fold has older rocks in its core: it is an anticline. However, the fold plunges to the east: it is a synform. This first-generation fold is a downward-facing fold that probably developed when folding affected a locally inverted flap of sediment. The geometrical interpretation is shown in the sketch at bottom right. Younging directions are indicated by "Y."

Fig. 3

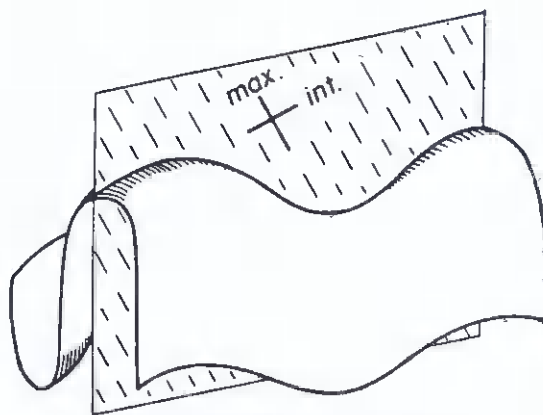


Fig. 4 Stylized regional fold geometry for the Shebandowan group deduced from cleavage-bedding and intersection lineation relationships. The fold plunge variation is inferred to be quite large, although the principal magnetic susceptibilities of the slates ("max." and "int.") show relatively little change in orientation within the axial-planar cleavage. Because the susceptibility is due to the grain fabric of the rock, we suggest that the variation in fold plunge occurred during the development of the folds rather than later.

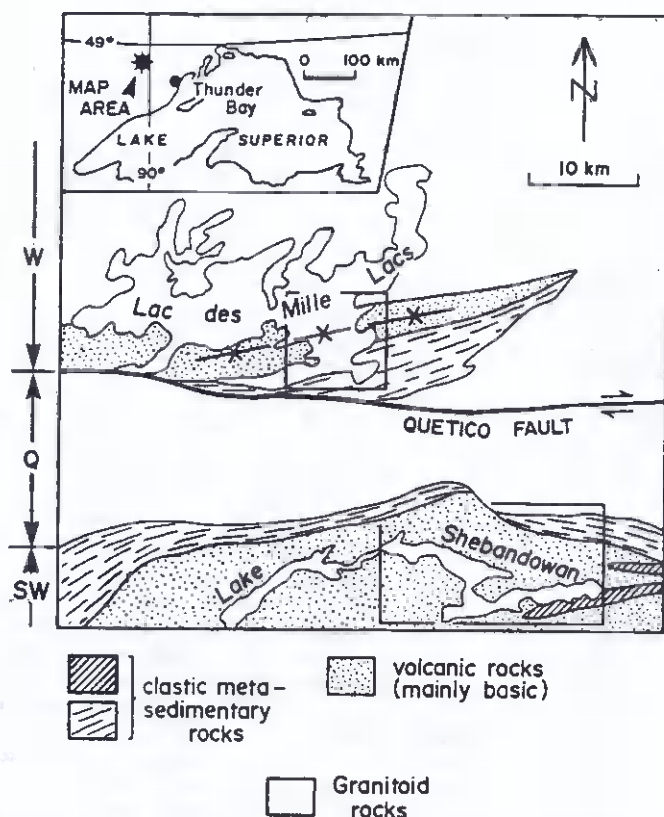


Fig. 5 Simplified regional geology after Pye and Fenwick (1965) with subprovinces (SW = Shebandowan-Wawa, Q = Quetico, W = Wabigoon) after Goodwin (1977). Information on boxed areas is presented in Figs. 2 and 3.

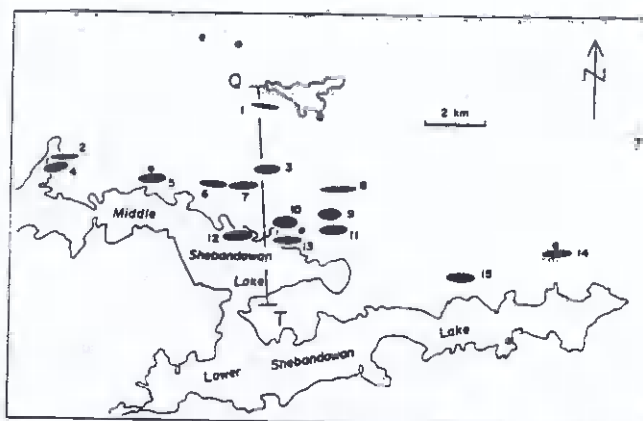


Fig. 6 Geological map of northern part of Shebandowan Lake greenstone structure (SLGS) with horizontal line segment (QT) used for strain estimate of overall shortening. Strain stations (see Table 1) are numbered.

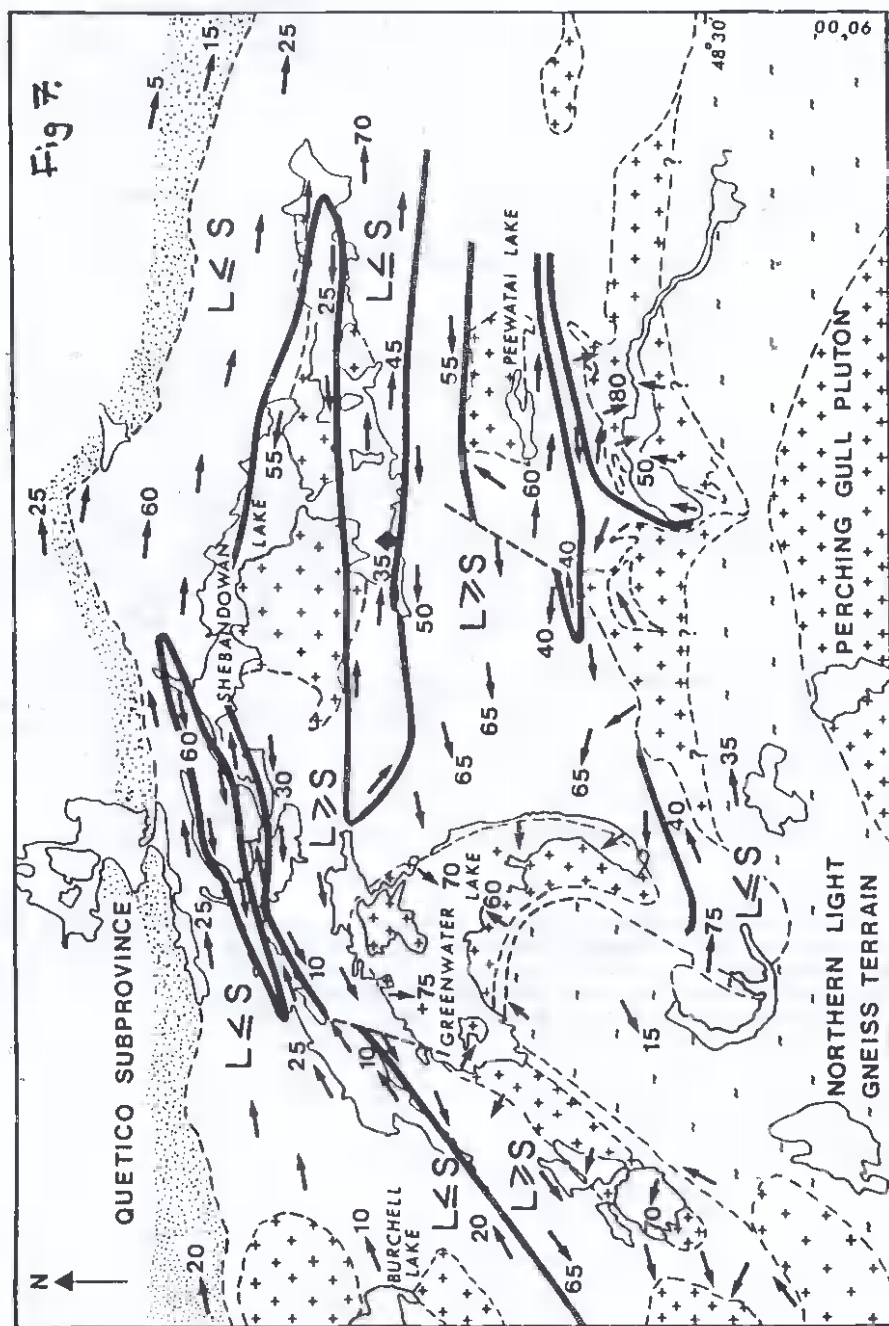


Fig. 7

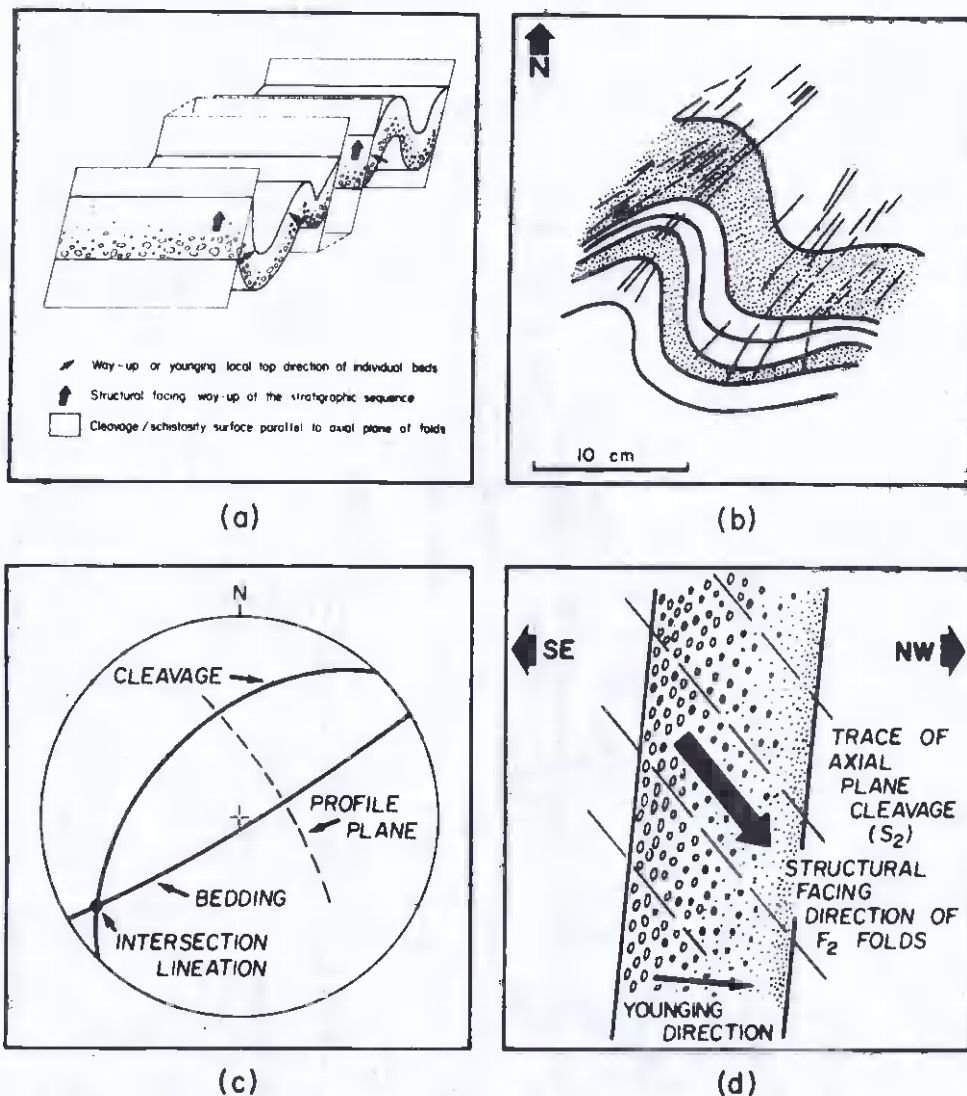


Fig. 8 The structural facing of folds. (a) A general illustration of the concept of structural facing of folds. The structural facing direction remains consistently oriented while younging directions are variable. (b) An outcrop sketch of a small F_2 fold which demonstrates that the S_2 cleavage is axial planar. (c) Lower hemisphere stereographic representation of bedding and cleavage and their mutual intersection. (d) A schematic profile view of bedding and cleavage. F_2 folds face downward since younger beds are encountered downward along the S_2 cleavage.

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FIELD TRIP 3

**Granitoid-related mineral deposits
of the western Lake Superior region.**

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Introduction

The locations of surface exposures of granitoid rocks of the western Lake Superior region are illustrated in Figure 1. The Archean granitoids are found in a western area in the vicinity of Thunder Bay, which is separated by Proterozoic sediments and diabase sills of the Nipigon Plate from various plutons lying south of Lake Nipigon. The western group is associated with the Shebandowan-Wawa greenstone subprovince and its boundary with Quetico Subprovince to the north. The easterly group of granitoids, with the exception of the Black Sturgeon Lake granite (6), lies within the Quetico Subprovince. The plutons in Figure 1 are numbered for purposes of identification in the following discussion. The names of the bodies, with a few exceptions, are those employed by McCrank *et al.* (1981) in their compendium.

The Glacier Lake pluton (3) is a batholith-sized body consisting of two-mica leucogranite in the north grading into biotite granite to the south. The rocks show minor penetrative deformation within a kilometer of their northern contact and are bounded by a weakly defined contact aureole in the intruded metasedimentary rocks, suggesting late tectonic emplacement. The rocks are moderately to strongly peraluminous, and although bearing no aluminous minerals other than micas, have the geochemical characteristics of S-type granitoids ($A/(CNK) \geq 1.1$, normative corundum, high $\delta^{18}O$ and high initial $^{87}SR/^{86}Sr$). Textural and geochemical patterns suggest that the pluton has been tilted to the north since emplacement (Zayachivsky *et al.*, 1989).

The MNW stock (1), with essentially identical characteristics, appears to be a satellite of the Glacier Lake Pluton. The Ilgour Lake Group of Zayachivsky (1985) and Kissin and Zayachivsky (1985), equivalent to the Kilgour Lake-Steen Lake metagabbro of McCrank *et al.* (1981), is a zoned granitoid intrusive with a core of monzogabbro grading outward to tonalites and granodiorites.

Also present in the Georgia Lake area are numerous tabular intrusions of leucotonalite (trondhjemite) to granodiorite of thickness ranging from a few meters to a few tens of meters and strike lengths of up to several kilometres. The petrogenic affinities of these are unclear as is the timing of their emplacement. In fact, contacts among the Glacier Lake pluton, the Kilgour Lake group and the tabular intrusions have not been observed in the field.

The Glacier Lake pluton is well exposed to the south of the Georgia Lake area, but its eastern margin is unmapped at present. It is exposed in the west along Highway 11, but to the west, between Highway 11 and the Nipigon River system, is a region of thick Pleistocene sediments, deposited in a spillway between the Lake Nipigon and Lake Superior basins. Metasedimentary rocks are intruded by granitoids of variable size and form; however, poor availability of outcrop makes interpretation of the overall geologic picture difficult. Farther to the west, a well-defined intrusion, the Church Lake quartz monzonite (4), is seen to intrude Archean metasediments and "migmatites". The petrography of these rocks is similar to that of the Glacier Lake pluton (Coates, 1972), and it is possible that they are parts of the same batholithic-sized body.

The Pine Portage intrusion (5) is exposed in a window in Proterozoic rocks at

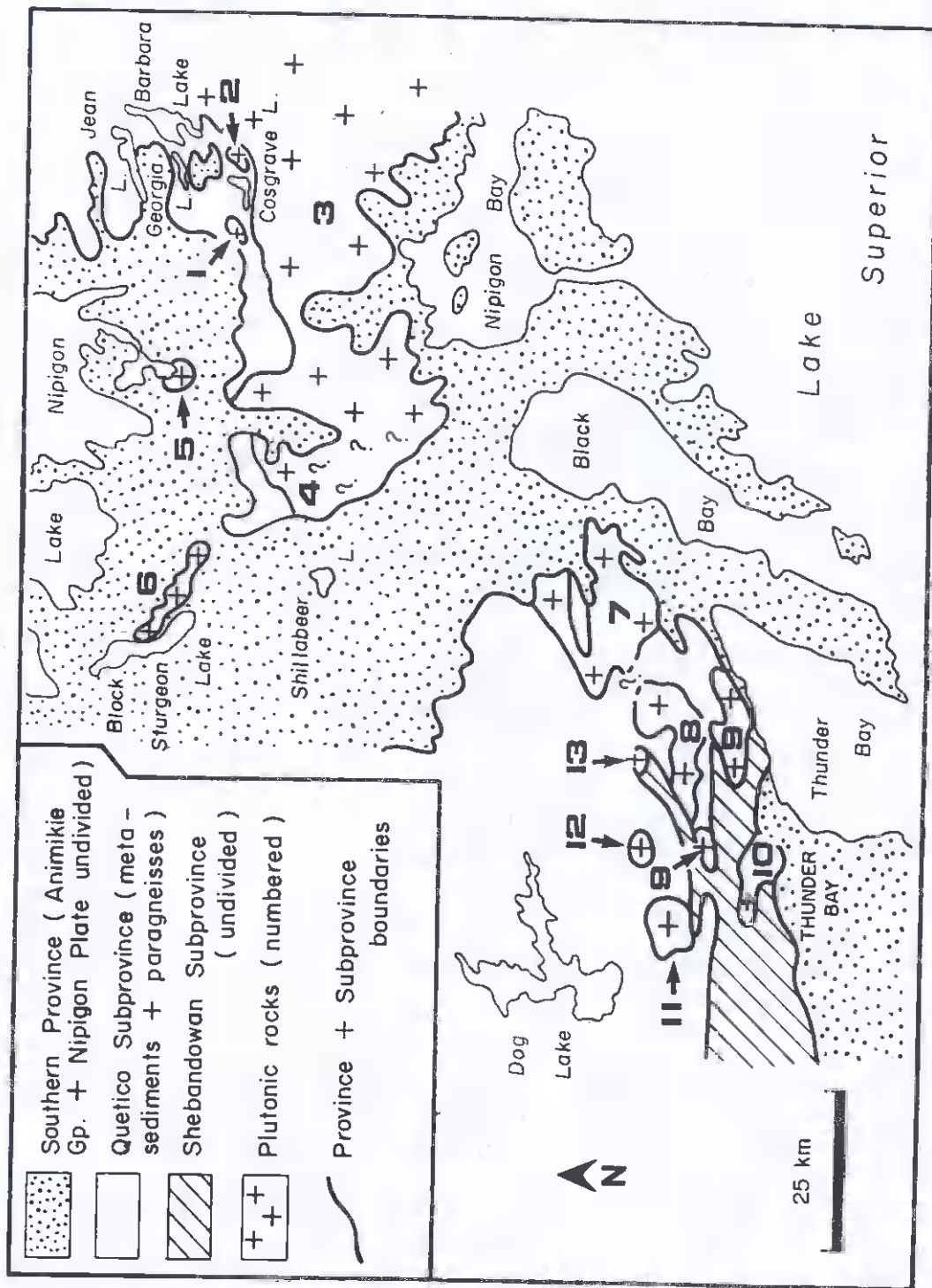


Figure 1. Geology of the western Lake Superior region indicating the locations of major granitoid plutons. Numerical codes are discussed in the text.

the source of the Nipigon River. It is composed of two-mica leucogranites cut by tabular leucotonatites, similar to those in the Georgia Lake area. The granitoids are closely associated with lithium-bearing pegmatite dikes, cutting metasedimentary rocks to the east. Geochemical fractionation trends in rare-element and simple pegmatites suggest that they are derived by fractionation of a leucogranite parent.

The Black Sturgeon Lake granite (6) is exposed in a window of Archean rocks east of Black Sturgeon Lake. It is associated with greenstone-like lithologies and may lie in the Wabigoon Subprovince.

West of the Nipigon Plate lies the Hilma Lake granite (7). The body consists white to pink two-mica leucogranite (Tihor, 1973), although textural compositional variations of the rock are found, including pegmatitic granite and pegmatite. The relationship to the Penassen Lakes stock of Scott (1985), equivalent to the Waller Lake granite of McCrank *et al.* (1981), is unclear owing to a lack of mapping. The two bodies are shown as contiguous by McCrank *et al.*, although the different names were applied to the separate bodies suggested in Figure 1. This differentiation of the two bodies is supported by the rather different petrography of the Penassen Lakes stock, which is mostly coarse-grained porphyritic granite to quartz monzonite with local hornblende syenite and syenite phases.

The MacKenzie granite (9), equivalent to the Waylan Lake granite of McCrank *et al.* (1981), has a complex outcrop pattern intruding Archean metavolcanics and metasediments, lying unconformably below Animike Group (Gunflint and Rove Formations) sediments and intruded by Keweenawan diabase sills and dikes. According to Rogers (1979), it is a medium-to-coarse-grained biotite granite. The MacKenzie granite is possibly equivalent to the Kivikoski granite (10), equivalent to the McIntyre granite of McCrank *et al.* (1981), as granite is known to underlie much of the city of Thunder Bay as basement to the Animike Group. The Gorham granite of McCrank *et al.* (1981) is included on Figure 1 as part of the MacKenzie granite (M.M. Kehlenbeck, pers. comm., 1990).

The Trout Lake granite (11), Barnum Lake quartz monzonite (12) and Whitelily Lake granite (13) are three similar plutons of circular to elliptical outcrop pattern aligned west to east near the Shebandowan-Quetico Subprovince boundary. The rocks are characteristically hornblende and biotite-bearing, porphyritic quartz monzonites. Kehlenbeck (1977) pointed out their late syntectonic emplacement, as indicated by well defined contact aureoles in the intruded Archean metasedimentary and metavolcanic rocks.

Mineral Deposits

Pegmatites

The most spectacular granitoid-related mineral deposits of the area are rare-element pegmatites in the Georgia Lake area and a small but similar pegmatite field in the Pine Portage area (Pye, 1965). These have been studied more recently by Zayachivsky (1985) and Archibald (1987), and these and other results are reported by Kissin & Zayachivsky (1985), Kissin *et al.* (1986) and Kissin and Archibald (1988). The

rare-element pegmatites of the Georgia Lake area are shown in Figure 2. The pegmatites may be subdivided into three groups, a northern group whose members show little difference in relative fractionation of trace element and central and southern group whose members show increasing degrees of fractionation from east to west. Barren pegmatites lie to the east of the central and southern group but are not shown in Figure 2.

Most of the pegmatites are unzoned and contain phenocrystic spodumene; however, the more highly fractionated pegmatites of the central and southern group show weak to well developed zoning and may contain, as well as spodumene, tantalite-columbite group minerals, beryl, cassiterite or staurolite and various Li phosphate minerals. The MNW pegmatite contains spodumene-quartz-intergrowth (SQUI) and is a potential source of ceramic grade spodumene.

An entirely separate group of pegmatites occurs within the Quetico Subprovince to the west of the Nipigon Plate. Franklin (1978) described uraniferous, white albite-muscovite-biotite-quartz pegmatites containing 60-100 ppm U. The pegmatites have been found to contain uraninite, and all such pegmatites containing accessory apatite have been found to be uraniferous. These pegmatites occur in an east-west trending zone within the Quetico Subprovince. Their origin is obscure, as they appear to be unrelated to anatectic mobilizate pods in high-grade metamorphic rocks nor to pink syenitic pegmatites apparently co-magmatic with granitic intrusions in the area. The Hilma Lake granite, however, was mentioned as having an anomalous uranium content with respect to other rocks of the area (Franklin, 1978).

Molybdenum

A sub-economic molybdenite occurrence is located at Anderson Lake, at the western margin of the Hilma Lake granite. Molybdenite occurs as coarse aggregates in quartz veins and within pink pegmatitic granite. It is unclear as to whether the mineralization process was pegmatitic or hydrothermal. The quartz veins lack characteristics of pegmatites; however, hydrothermal alteration characteristic of vein- or porphyry-style molybdenite deposits is also lacking. The average grade of the occurrence is reportedly 2.85% Mo; however, the apparent small size of the reserves is responsible for a lack of exploitation.

Non-magmatic deposits, hosted in granite

Uranium

Uranium occurrences in the western Lake Superior region were summarized by Franklin (1978). Those related to granitoid rocks include uraniferous pegmatites in the Quetico Subprovince and vein-type occurrences in granitic basement or cutting Sibley Group sediments. The vein-type occurrences are not of direct hydrothermal derivation from granitic magmas, but may be related to uraniferous pegmatites and uranium-rich granites as protore.

The uraniferous pegmatites are prominent in the Greenwich Lake area (U(1) in Figure 4), and exploratory work has indicated that they occur about 40 km to the west

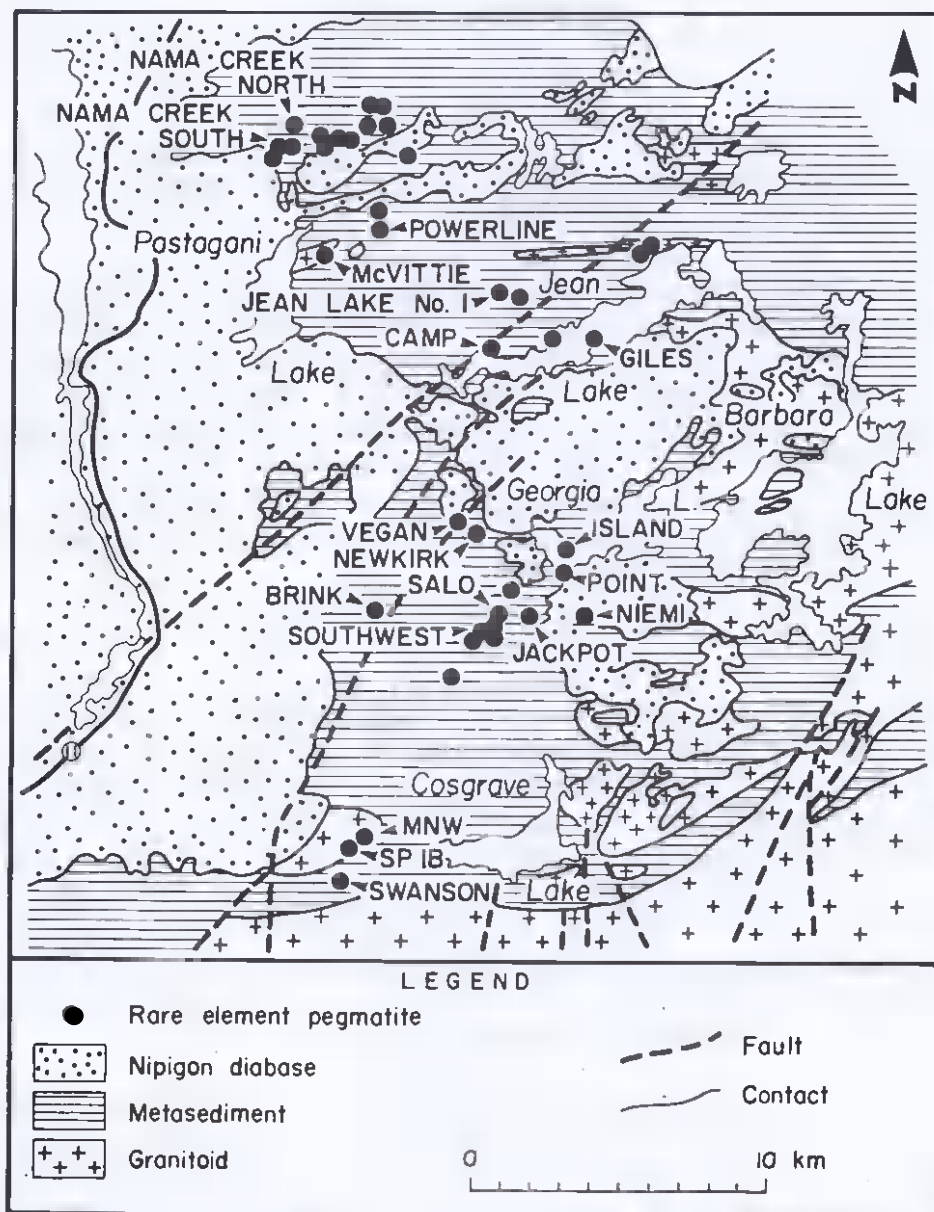


Figure 2. The Georgia Lake Pegmatite Field illustrating the granitoid rocks and rare-element pegmatites. Named pegmatite bodies were studied by Zayachivsky (1985).

along Highway 527. As described previously, these pegmatites are contained within a relatively narrow east-west trending zone. The small size of the pegmatite pods and lenses and their low uranium content preclude the possibility of economic deposits.

The vein-type occurrences (Figure 4) at Greenwich Lake (U(1)) and Innes Lake (U(2)) have been described by Franklin (1978) and Yule (1979). Both occurrences are sub-economic but were the source of considerable excitement in the late 1970's. Both occurrences are localized in brecciated shear zones cutting all Archean lithologies. The Greenwich Lake occurrence occurs in well defined veins with quartz and pyrite gangue and minor pitchblende yielding grab sample grades of 0.5 to 2.0% U. Hydrothermal alteration of wall rock is distinct, and the vein contains altered Sibley Group fragments indicating its Proterozoic age. The Innes Lake occurrence has no recognisable uranium minerals, although the shear zone is much enriched in apatite, which is thought to be the carrier of uranium. Hydrothermal alteration including chloritization, sericitization and hematization is distinctive about the shear zones. The shear zone yielded values as high as 633 ppm U_3O_8 .

Franklin (1978) also described minor uranium occurrences (up to 540 ppm U) in the Enterprise Mine, one of a group of vein-type deposits known as the lead-zinc-barite veins. These are discussed more fully in the next section; however, both Franklin (1978) and Yule (1979) believed that the Greenwich Lake and Innes Lake occurrences were formed by the same process that formed the sulfide-rich lead-zinc-barite veins. Briefly, the hypothesis favored is circulation of basinal brines through uranium-enriched source rocks, the uraniferous pegmatites and anomalous granitic rocks. The brines circulated through faults marginal to the Sibley Group depositional basin and rose and cooled in response to hydraulic gradients.

Lead-zinc-barite

The lead-zinc-barite veins (Figure 3) are near the margins of the present outcrop of the Sibley Group or its inferred former outcrop. The veins were studied in detail by Franklin & Mitchell (1977), who found them to contain galena, sphalerite and barite, with minor chalcopryite and marcasite in quartz and calcite gangue. The veins occur in three settings: (A) fractures within the Sibley Group, (B) fractures at the Sibley-basement unconformity and (C) fractures within granitic basement. Franklin & Mitchell believed that the deposits formed by mixing of bacterially reduced H_2S with metal-rich brines derived by leaching of the Sibley basinal fill. Deposition occurred in structural traps at or near the basin margin. The basinal character of the brines was confirmed in recent fluid inclusion studies by Haynes (1988).

The differentiation of structural settings by Franklin & Mitchell suggests that type (C) may be a portion of a continuum of deposits extending to the uraniferous occurrences at Greenwich Lake and Innes Lake. The lead-zinc-barite veins represent a sulfide-rich but uranium-poor end-member, whereas the uranium occurrences are sulfide-poor but uranium-rich. The differential factor seems to have been the relative amount of interaction with basinal fill (base metal source) or granitic basement (uranium source).

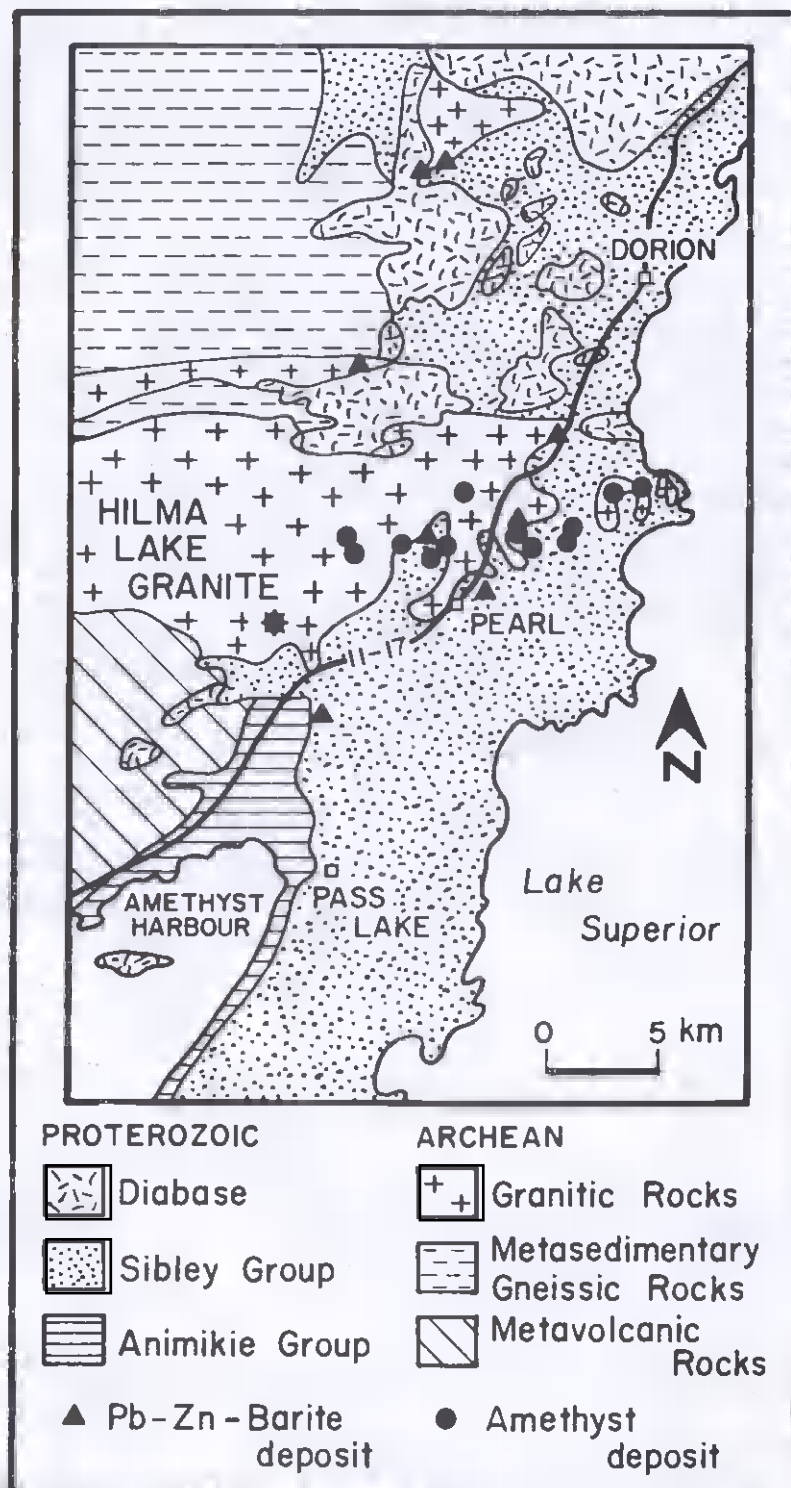


Figure 3. The Dorion area illustrating the locations of major amethyst deposits and lead-zinc-barite veins. The star is the location of the Thunder Bay Amethyst Mine.

Amethyst

The largest deposits of amethyst in North America are located in the Thunder Bay area (Figure 3). Amethyst is the provincial gemstone of Ontario and is a significant generator of economic activity in the region. Nearly all the important deposits occur in a 10 km-long, east-west trending near the margin of the Sibley Group outcrop and with Hilma Lake granite as the basement. A few other scattered occurrences are spatially related to the Sibley Group outcrop and are invariably hosted in granite.

The amethyst deposits consist mostly of quartz including much of the variety amethyst with minor calcite, barite and sulfides such as pyrite, chalcopyrite and bornite. The similar geologic setting of these deposits to that of the lead-zinc-barite veins suggests that they are cogenetic. Fluid inclusion studies by McArthur (1988), reported in McArthur & Kissin (1988), indicate that most amethyst formed near the surface at temperatures ranging from 60 to 90 °C. The deposits appear to be low-temperature derivatives of lead-zinc-barite veins, poor in sulfides and richer in quartz. A crucial factor appears to have been passage of the quartz-depositing solutions through a uranium enriched host rock, e.g. the Hilma Lake granite. A source of silica (from sericitization of feldspar) and radioactivity are probable requirements for the formation of amethyst. The formation of the electronic colour center in quartz producing the blue of amethyst has been attributed to the coupled electronic transition:



The associated Fe^{4+} (substitutional) + Fe^{2+} (interstitial) defect (Cohen & Hassan, 1974) is evidently unstable at high temperatures, as older generations of amethyst were observed to be bleached by influx of hotter fluids.

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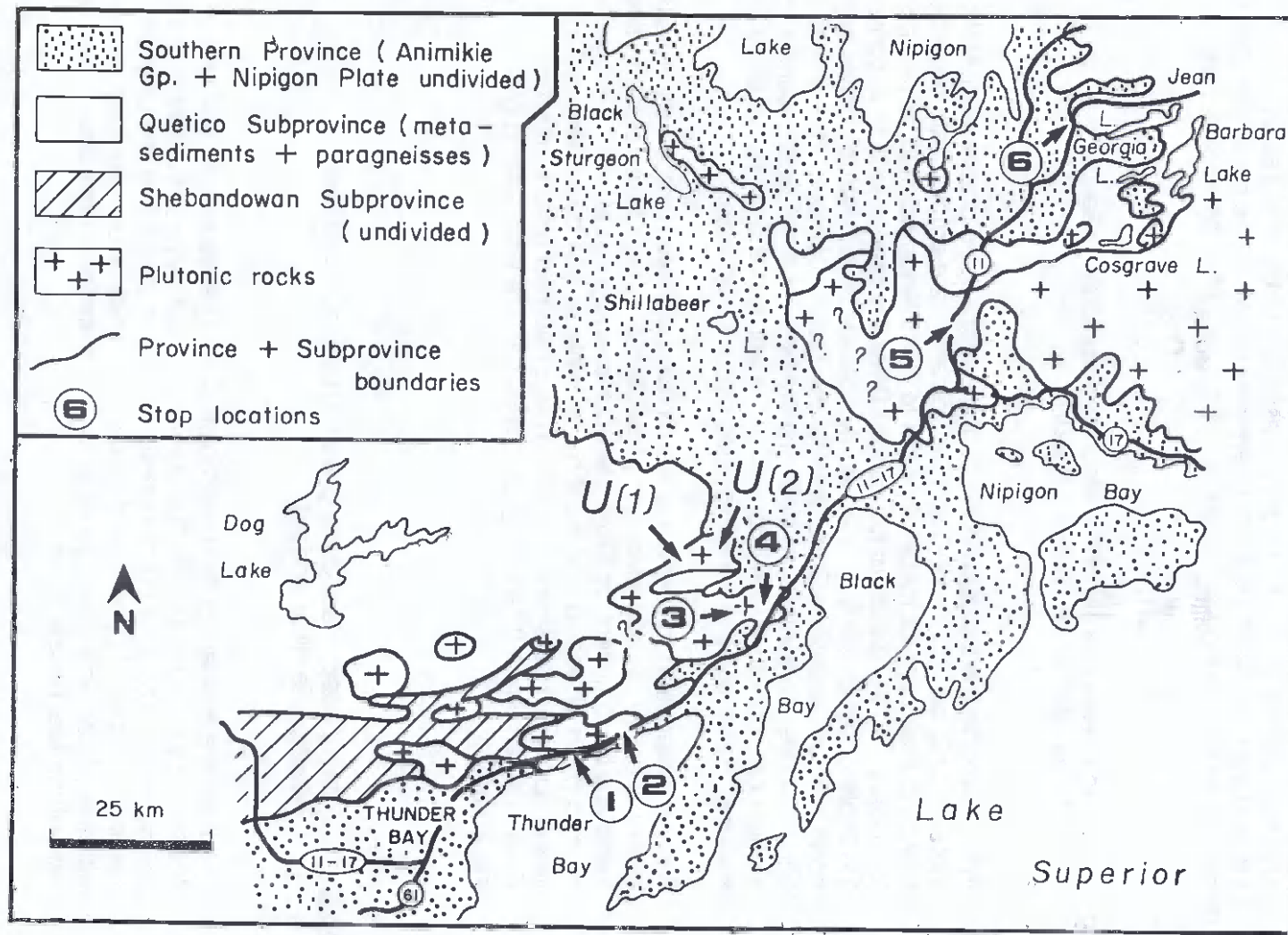


Figure 4. Geology of the western Lake Superior region showing numbered field trip stops. U(1) and U(2) indicate the locations of the Greenwich Lake and Innes Lake uranium occurrences.

FIELD TRIP LOG

This field trip will attempt to illustrate as many of the features as possible in one day. Unfortunately, some of the most interesting ones are the most inaccessible. The trip route is illustrated in Figure 4, together with a simplified geological map.

Stop 1: Southeast margin of the MacKenzie granite (Lakeshore Drive near Silver Harbour Road).

In this location the intrusive contact of the MacKenzie granite with Archean metavolcanic can be seen. The metavolcanics are amphibolitic near the contact reflecting their contact metamorphism to hornblende hornfels facies, according to Rogers (1979). South of Lakeshore Drive the granite is in fault contact with the Proterozoic Gunflint Formation. A small Gunflint outcrop occurs on the north side of Lakeshore Drive, west of the granite contact. The granite is jointed and in places the joints have been mineralized. A small quartz-calcite vein is visible here, but other veins contain pyrite and small amounts of base metal sulfides. Small veins of amethyst also occur in the area.

A short distance to the west and south toward Silver Harbour are a series of silver mines, which operated in the late 19th Century. From north to south these are the 3A, Beck or Silver Harbour and Algoma Mines. The shafts have been capped or sealed and most dump material has been removed in recent years. Little can be seen of these. These deposits were formed in Keweenawan times (Franklin *et al.*, 1986) and are unrelated to granitic rocks of the area.

Stop 2: Main phase, MacKenzie granite (Highway 11-17 near MacKenzie River bridge).

The massive and leucocratic character of the granite can be seen here. According to Rogers (1979), the color index is always ≤ 5 , making the granite a leucogranite. The only primary mafic mineral is biotite and the only accessories noted are traces of sphene and apatite. No mineral occurrences are associated with this pluton apart from some small but commercial amethyst-bearing veins east of the MacKenzie River.

Stop 3: Anderson Lake Molybdenum occurrence (East Loon Road from Highway 11-17. Continue 1.8 km to north side of Loon Lake. Turn right immediately past bridge over small stream. Continue about 200 m to Hydro transmission lines and clearing. Park and continue north on road by foot for about 3 km to Anderson Lake.)

The molybdenite occurrence described in the introductory text was diamond drilled and blasted at surface in the 1960's, giving rise to the chaotic scene here. Good specimens of coarse-grained molybdenite and pegmatitic granite host rock may be collected.

Stop 4: Thunder Bay Amethyst Mine (back-track on East Loon Road to junction with road to Thunder Bay Amethyst Mine. Turn right and continue for approximately 10 km to the Mine.)

Note: This is a commercial operation entered by permission of the owner. There is to be no hammering nor collecting in the pit. Collecting may be done in the designated dump area, but material collected here is not to be broken by hammering.

The amethyst-bearing vein system occupies an east-west striking, near vertical fault zone. The fault is parallel to the trend of occurrence of other amethyst deposits in the area (Figure 3) and may form a major margin to the Sibley Group depositional basin. The fault is offset by steps of a few meters by a series of northerly striking en-echelon faults. The easterly and westerly limits of the fault have not been determined.

The amethyst reaches its most spectacular development in large vugs where crustiform aggregates contain crystals of quartz 30 to 40 cm long and 10 to 15 cm in diameter. The crystals display growth zoning, which is consistent throughout the deposit. The veins contain little apart from quartz, although minute sulfide crystals occur in some growth zones. The altered granite countryrock contains occasional accumulations of pyrite, chalcopyrite, bornite and other minor sulfides. The vein system is surrounded by a well-defined, sericitized, hydrothermal alteration zone.

The vuggy character of the vein indicates that it must have formed near the surface. However, even more convincing is the presence of breccia fragments in the vein consisting of Sibley Group lithologies. The Sibley Group has been entirely removed from the area by erosion, but its former presence is obvious. On a granite knob north of the vein, a chloritized and hematitic surface possibly may represent a Precambrian paleoregolith.

Stop 5: High-grade central portion of the Quetico Subprovince (Highway 11, east side of Lake Helen).

In this area, well developed migmatites are indicative of the highest metamorphic grades attained in the Quetico Subprovince. Metamorphic grades decline rapidly to the north, and much of the metasedimentary terrane in the Georgia Lake area is lower amphibolite to upper greenschist facies.

Stop 6: McVittie Pegmatite cutting the Postagoni Lake sill (turn off Highway 11 at George Creek and continue from approximately 8 km to well-defined clear cut on west side of road at Postagoni Lake. Park and proceed west on foot for 1/2 km to north side of Dive Lake to McVittie pegmatite.)

The McVittie Pegmatite is a north striking vertical dike approximately 10 m wide, which cuts the east-west striking tonalites of the Postagoni Lake sill. The pegmatite is typical of the north group pegmatites of the Georgia Lake area in that it is unzoned and contains no rare-element-bearing minerals other than spodumene. The spodumene is typical phenocrystic spodumene containing several percent iron and forming euhedral, prismatic crystals up to 10 cm long. The spodumene appears greenish owing to its iron content and partial decomposition to chlorite and sericite.

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FIELD TRIP 4

BASE METAL MINERALIZATION

IN THE

SHEBANDOWAN GREENSTONE BELT

Introductory Discussion and Field Guide
36th Annual Institute on Lake Superior Geology
Thunder Bay, Ontario

by

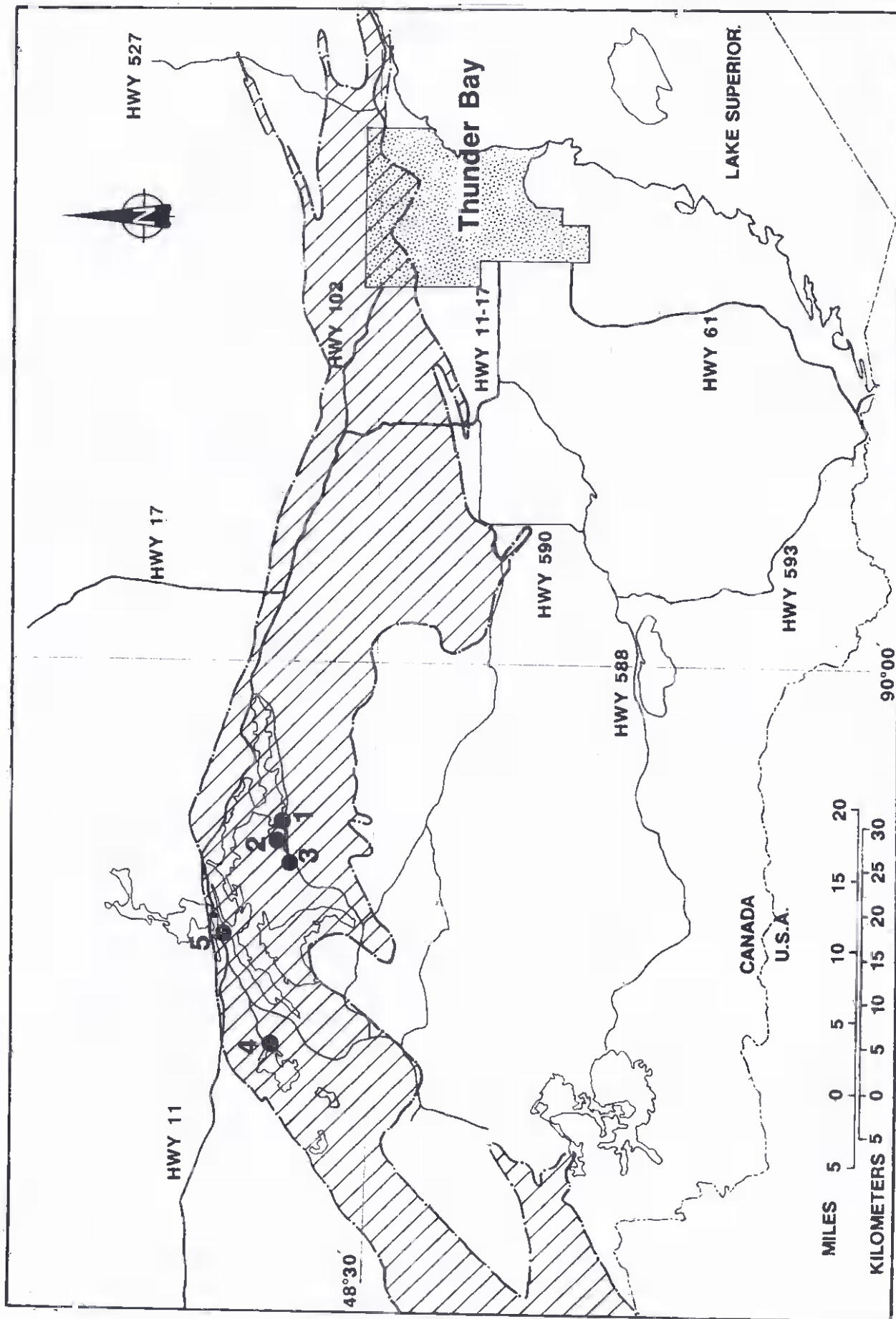
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INTRODUCTION

Mineral deposits of the Shebandowan greenstone belt bring about an extra degree of controversy with respect to their origins as the result of the destruction of prerequisite diagnostic textures and relationships by intense deformation. This is especially true for deposits which predate deformation, such as synvolcanic mineralization. Although unintended, this field trip highlights the trial and tribulations of trying to "pigeon hole" highly deformed mineralization. Three styles of base metal mineralization will be examined during the course of this trip, and all are proposed candidates for synvolcanic origins. Allan Aubut will be presenting arguments which favour that the Shebandowan nickel-copper mine is komatiite flow hosted, ie., Rambalda type. The other authors will bring forth arguments that the North Coldstream Mine is the stringer zone of a Kuroko type, volcanogenic, polymetallic sulphide deposit and that the Vanguard prospect is the exhalative component of such a deposit.



STOP 1; SHEBANDOWAN MINE

INTRODUCTION

The Shebandowan Mine is a nickel-copper deposit owned by Inco Limited and is presently being operated by MacIsaac Explorations. It is located in northwestern Ontario, 75 kilometres west-northwest of Thunder Bay. The mine is situated on the south-western shore of Lower Shebandowan Lake (Fig. 1).

It has been in semi-continuous production since 1972 at an average production rate of about 2000 tons per day. Presently reserves of broken and developed material stand at approximately 2 million tons grading 2.10 per cent nickel and 0.95 per cent copper. Accessory platinum group metals and gold are also recovered. Concentrate produced is shipped by truck to Sudbury, Ontario, for smelting and refining.

The Shebandowan Mine has had a long history with nickel-copper sulphide mineralization being first discovered in 1913 at what is now Discovery Point, on Lake Shebandowan (just east of stop 2, Fig 1). The International Nickel Company of Canada, now Inco Limited, optioned the property in 1936. The property was purchased for \$250,000 in 1937 (Daily Times Journal, 1937). Inco explored the property off and on for the next 28 years. Between 1965 and 1968 exploration was intensified with the collaring of an exploration shaft in 1966. A production decision was announced in 1968 and the first shipment of concentrate was made in 1972.

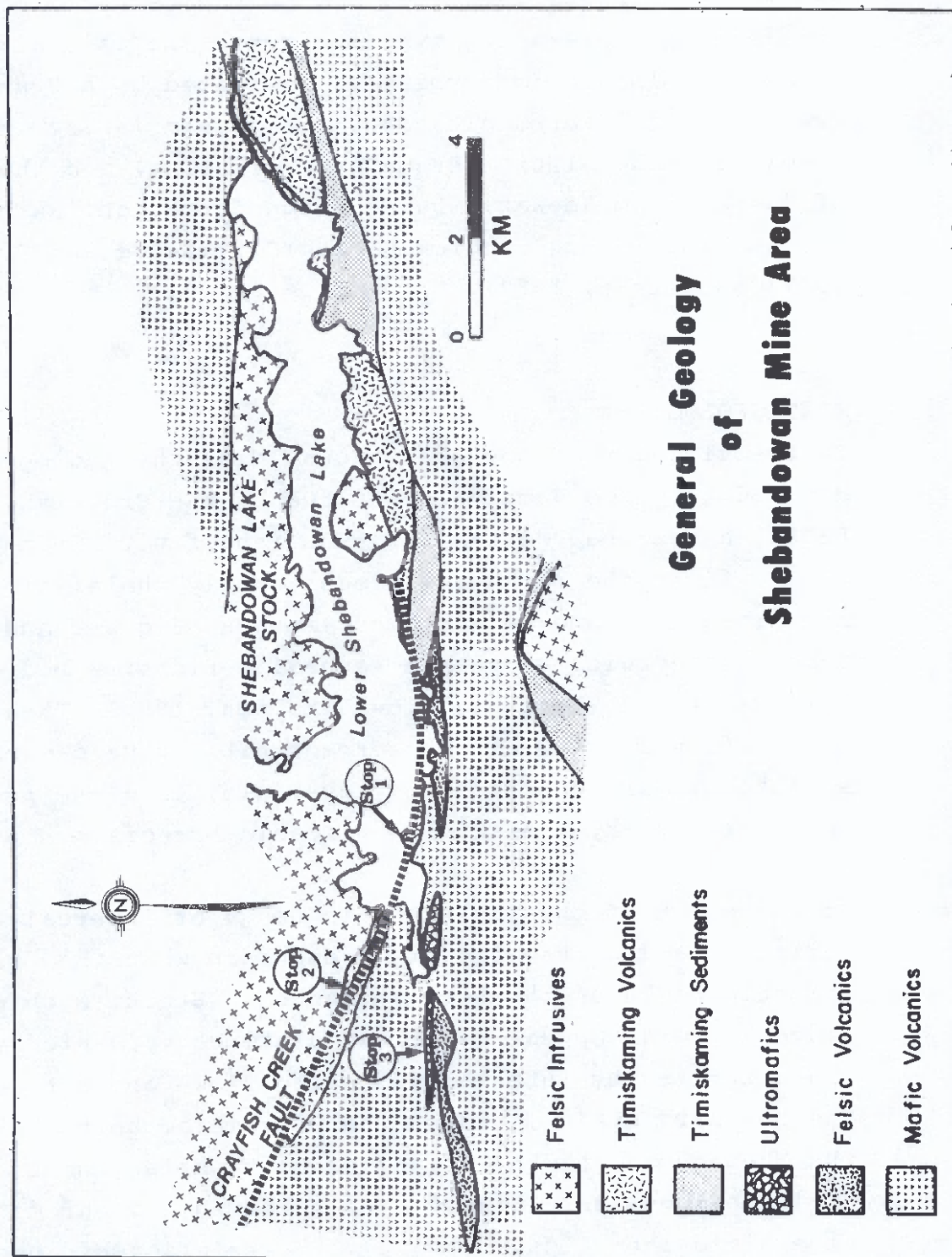


Figure 2. Shebandowan Mine - General Geology

REGIONAL GEOLOGY

The area in the vicinity of the Shebandowan Mine (Fig. 2) is underlain by Keewatin tholeiitic volcanics unconformably overlain by Timiskaming type sediments and calc-alkaline volcanics. This unconformity is indicated by a truncation of Keewatin axial surface trends, the dissimilarity in chemistry (tholeiitic versus calc-alkaline) and the presence of jasper from Keewatin banded iron formation, as clasts within Timiskaming conglomerates (Borradaile and Brown, 1987; Shegelski, 1980).

MINE GEOLOGY

In the vicinity of the Shebandowan Mine the geology can be divided into two domains separated by the Crayfish Creek Fault, a regional dextral transcurrent fault. To the south of the fault the rocks are predominantly tholeiitic basalts, andesites and some felsic pyroclastics (Figs. 2 and 3). Intercalated with these are several peridotite bodies believed to be komatiite flows (Morton, 1982). These ultramafics are now either serpentinite or talc-carbonate schist. All are unconformably overlain, in close proximity to the fault, by Timiskaming volcanic breccia.

To the north of the fault a thin wedge of intercalated mafic volcanics and ultramafics is present between the fault and granitic rocks of the Shebandowan Lake Stock. Within this wedge are what appear to be two distinct volcanic cycles characterized by intercalated mafic flows and ultramafics. The northern cycle includes the ultramafic unit that hosts the Shebandowan nickel copper deposit while the southern cycle includes an ultramafic unit that hosts chromite mineralization (Figs. 3 and 4). The relationship between the mafic volcanics and the ultramafics is unclear due to widespread deformation associated with the Crayfish Creek Fault deformation zone.

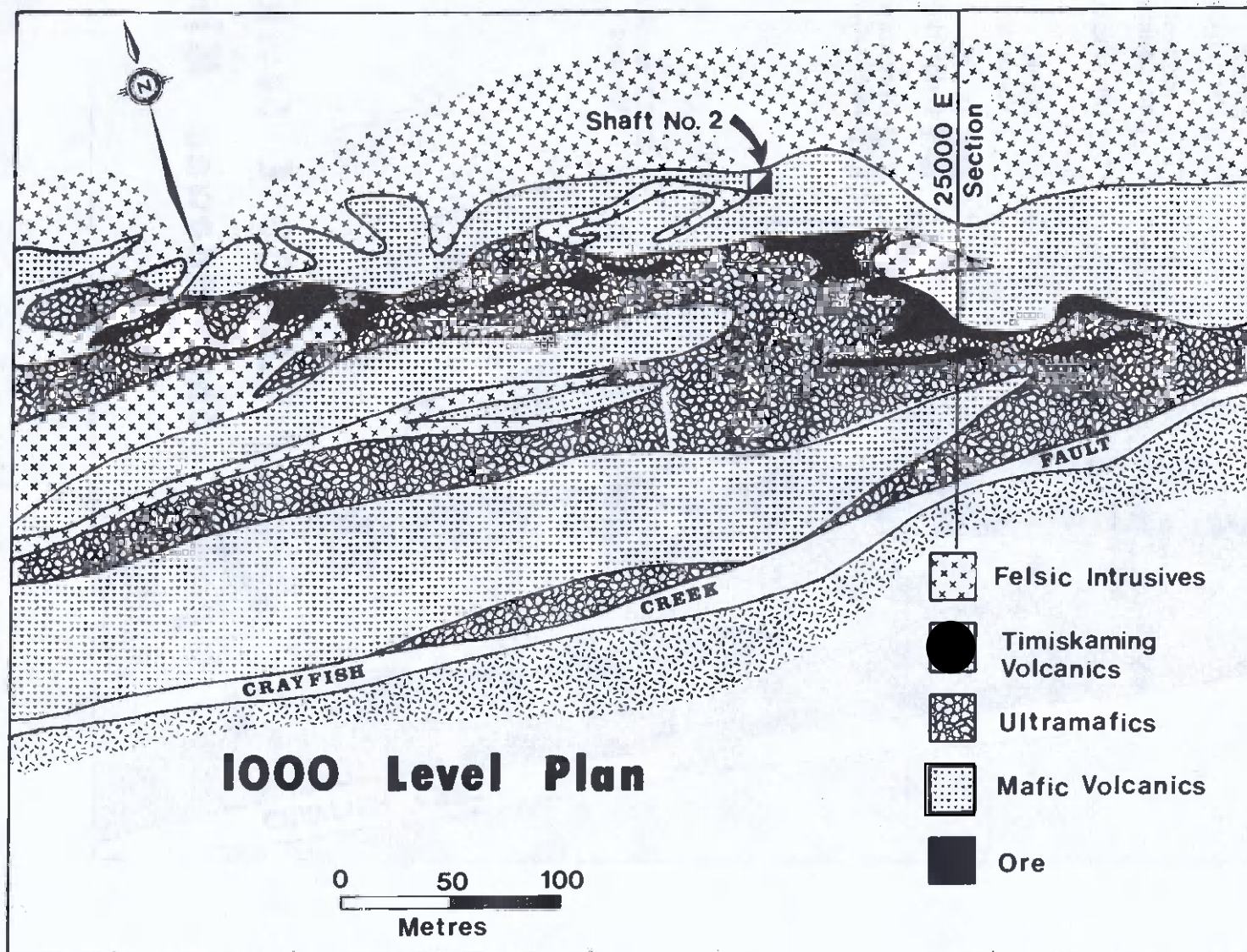


Figure 3: 1000' Level Plan - Shebandowan Mine

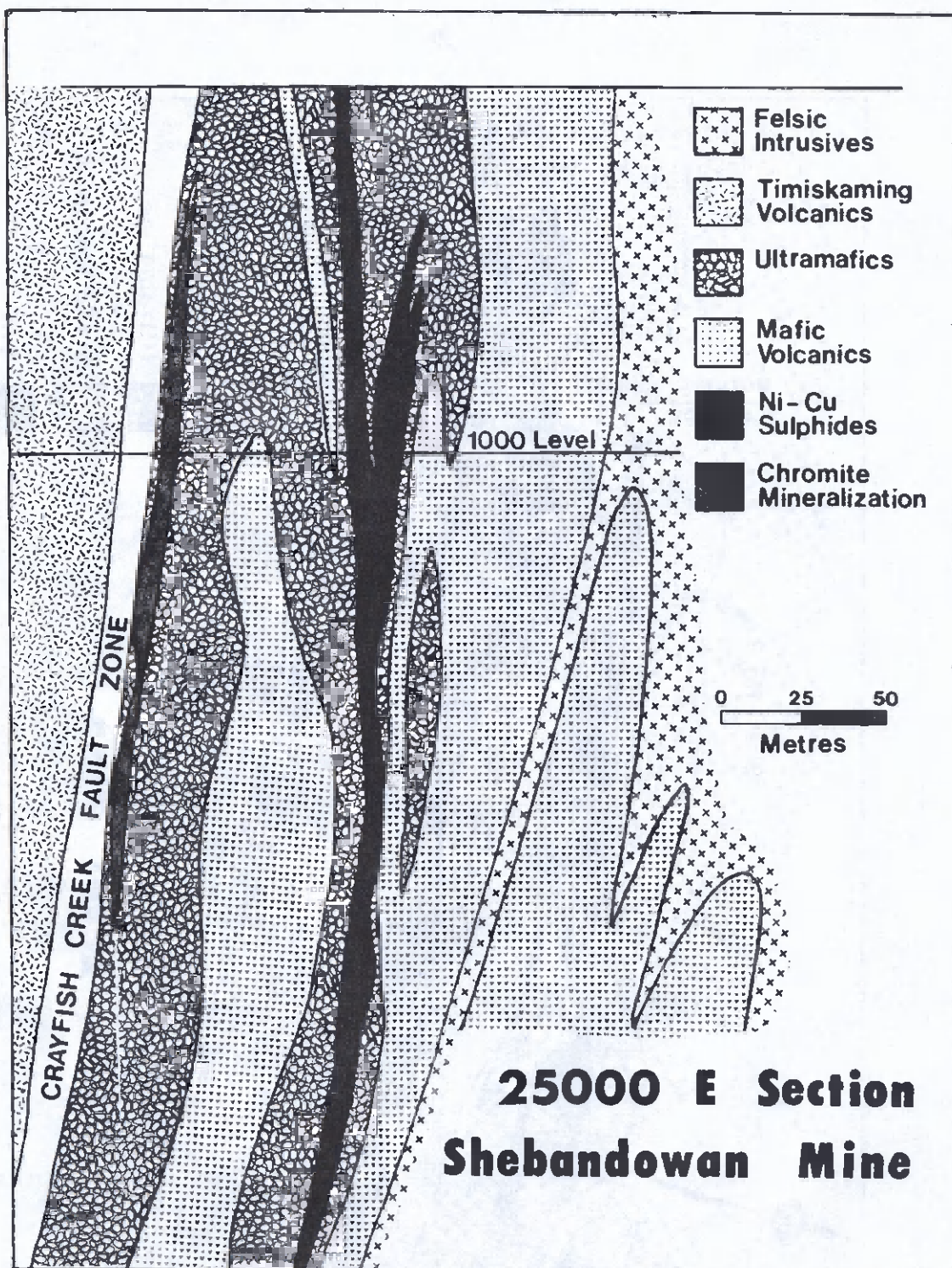


Figure 4: 25000E Section
Shebandowan Mine

STRUCTURAL GEOLOGY

The main structural feature of the Shebandowan Mine is the Crayfish Creek Fault. This major dextral transcurrent fault dips steeply to the south. Though a distinct fault breccia zone up to several metres thick is present a wide zone of deformation characterized by extensive shearing is present. To the south of the fault much of the deformation is hosted by the Timiskaming volcanic breccia resulting in a highly foliated quartz sericite schist possessing remnant breccia fragments and relict hornblende-phyric texture.

North of the fault shearing is not as pervasive, likely due to the more incompetent nature of the ultramafic-mafic contacts and the ultramafics themselves. This resulted in deformation being focused into numerous narrow shear zones, commonly concentrated along contacts with dextral transposition of slices. This tectonic reworking has had a major effect on the nickel-copper mineralization producing significant sulphide movement. The sulphide zones show numerous signs of this remobilization such as: disseminated zones occurring above, below and in blocks within massive sulphide zones; as sharp contacts between massive or semi-massive ore and disseminated sulphide bearing ultramafic; and the presence of pentlandite banding within the massive sulphides.

Another factor that has greatly added to the complexity of the deposit is the presence of numerous apophyses of the Shebandowan Lake Stock. These frequently follow along contacts, shears and sulphide zones as well as cross-cutting these features. This has made interpretation of stratigraphy in the mine environment very difficult as well as producing dilution problems in the ore.

MINERALIZATION

A) Nickel-Copper

The Shebandowan Mine hosts three main types of ore: stringer sulphides, breccia sulphides and massive sulphides. Though disseminated sulphides are present they are not of economic importance.

The stringer sulphides are confined to mineralized shear zones. Sulphides present include chalcopyrite, pyrite and minor pyrrhotite and pentlandite.

Breccia sulphides consist of fragments of ultramafic, volcanic and granite set in a matrix of pyrrhotite, chalcopyrite and pentlandite. Frequently chalcopyrite coats many of the inclusions within the breccia sulphide. Chalcopyrite is also commonly concentrated along the margins of the breccia zones.

Massive sulphide is composed of pyrrhotite, chalcopyrite and pentlandite. The latter is usually present as discontinuous layers that are generally parallel to the ore zone contacts. These layers also wrap around large inclusions.

The average copper to copper-nickel ratio is 0.36 with it being higher, or more copper rich, in the western portions of the mine and lower, or more nickel rich, in the eastern portions of the mine. This variation in copper-nickel ratio corresponds to the higher proportion of the more chalcopyrite rich breccia sulphide in the western part of the ore body.

B) Chromite

Uneconomic chromite mineralization is also present in the Shebandowan Mine environment. This chromite is hosted by ultramafics within the second volcanic cycle, close to the Crayfish Creek Fault. Chromite is present as finely disseminated grains in all the ultramafics in the area (Morton, 1982).

The second mafic volcanic-ultramafic cycle hosts chromitite (chromite rock) as massive to strongly contorted bands varying from a few millimetres to several metres in thickness. On either side of this main zone chromite is found as brecciated chromitite and chromitiferous peridotite. Chromitite mineralization has been traced discontinuously along strike for a distance of about 5 kilometres.

Chromium trioxide assays have been reported as high as 44 per cent but are typically much lower. Minor platinum group metals are associated with the chromite. The chrome-iron ratio varies from 1.2 To 1.8 although it rarely exceeds 1.6.

GENESIS OF THE NICKEL-COPPER ORE

The ore body has been subjected to intense structural deformation which has resulted in the remobilization of nickel-copper sulphides. Primary textures have been obscured if not erased. Regardless, away from the Crayfish Creek Fault Deformation Zone there is evidence that many if not all of these ultramafics are komatiites.

A komatiite is defined simply as being an ultramafic volcanic rock with 18% or greater MgO (Arndt and Brooks, 1980). These high magnesian flows have a number of physical

characteristics which can be used to identify them. A conclusive feature of komatiite flows is the presence of a relatively thin layer with spinifex texture. According to Arndt et al. (1979) most ultramafic flows are not spinifex textured but massive with polyhedral jointing. These polyhedra are coarse in the centre of the flow and become finer as the contacts are approached. It is this polyhedral jointing that is common in many of the ultramafic bodies at Shebandowan.

Another common feature at Shebandowan is the presence of chemical sediments, as both banded cherty iron formation and chert, immediately overlying the ultramafics (Morton, 1982). This feature implies that when the sediments were deposited the ultramafics must have been present as flows on the ocean floor.

The Shebandowan Mine ore body shows a lot of physical similarities with the komatiite hosted Redross deposit, in the Kambalda region of Australia. The Redross deposit is located 58 kilometres south of Kalgoorlie. It has been affected by regional dynamic metamorphism with faulting and shearing, primarily along the footwall contact between the sulphide bearing komatiite flow and mafic volcanics (Barrett et al, 1977).

The sulphides have undergone major remobilization along the footwall contact and shear zones. A major ore type consists of brecciated wall rocks cemented by sulphides. Contacts between massive or semi-massive ore and disseminated ore are commonly very sharp with none of the intervening matrix ore associated with other Kambalda deposits (ibid.).

The ore zones at Redross possess structures and textures produced by remobilization of both disseminated and more massive mineralization (ibid.).

At the Shebandowan Mine the same characteristics exist including an affinity of sulphide mineralization for one side of the ultramafic body. It is therefore suggested that the Shebandowan nickel-copper deposit is the product of dynamic metamorphism of a sulphide bearing komatiite flow with remobilization and subsequent redeposition of the sulphides in low strain areas.

FIELD TRIP STOPS

STOP 1; Shebandowan Mine Headframe, Fig. 5

At this stop we will examine strongly deformed Timiskaming calc-alkaline volcanic breccia as well as the ultramafic unit hosting the Shebandowan nickel-copper deposit.

The Main structural feature at the Shebandowan Mine is the Crayfish Creek Fault. South of the fault Timiskaming volcanics have been strongly deformed and are now quartz-sericite schist with possessing relic textures. North of the fault deformation has not been as focused into narrow shear zones separating lenses of serpentinized peridotite. This deformation has resulted in the remobilization of the Ni-Cu sulphides and subsequent deposition in lower strain areas within the host ultramafic.

STOP 1A.

This outcrop consists of strongly deformed hornblende phyric Timiskaming type volcanic breccia. Note the highly foliated nature reflecting its proximity to the Crayfish Creek Fault, approximately 100 metres to the north. Though strongly foliated note the many similarities with the undeformed Timiskaming volcanic breccias seen to the east along the road leading to the Shebandowan Mine.

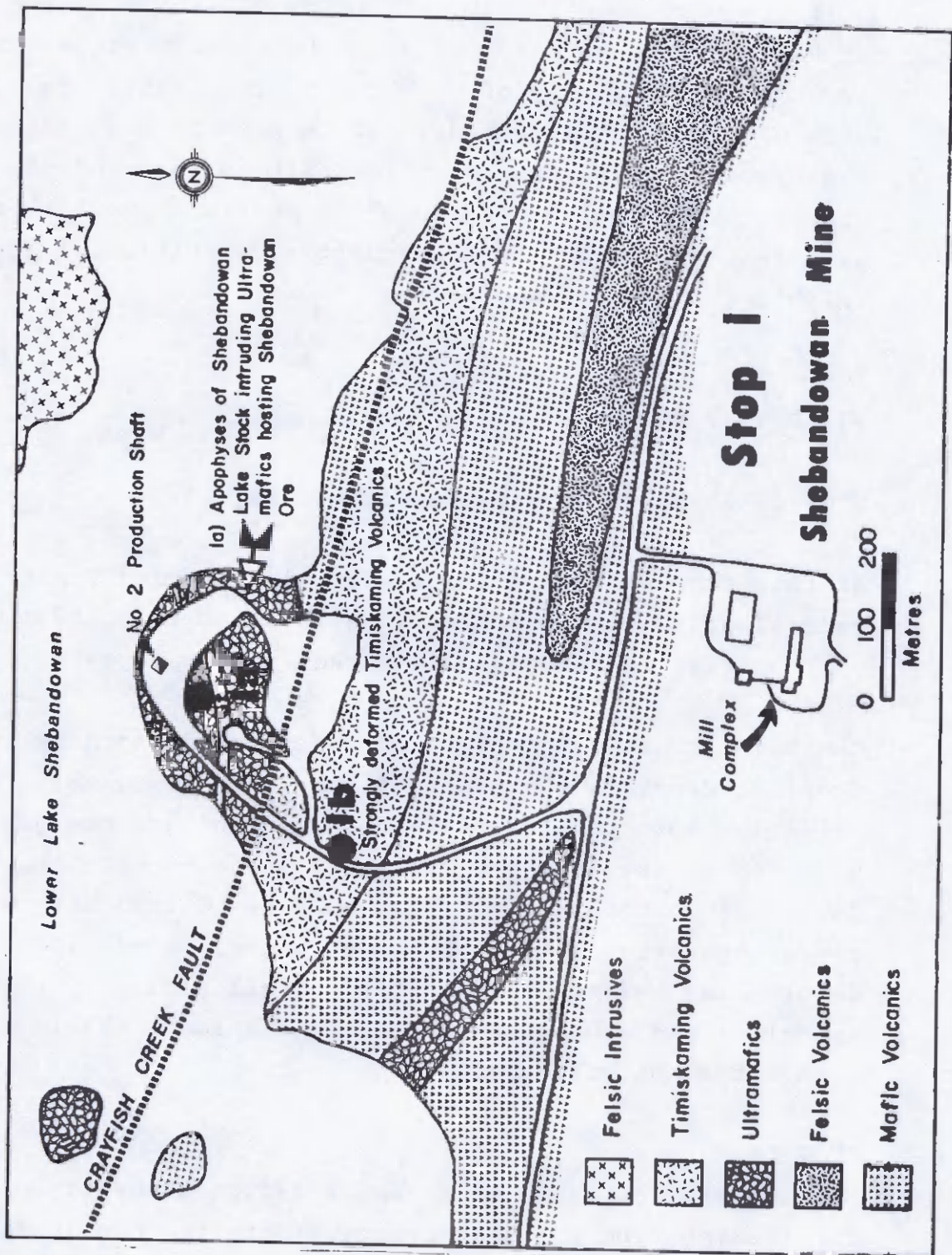


Figure 5: Stop 1 - Shebandowan Mine

STOP 1B.

This outcrop, beside the #2 Shaft headframe, consists of serpentized peridotite with numerous narrow zones of talc-carbonate schist that form an anastomosing network. It is this unit that is host to the nickel-copper sulphides. Note on the north side of the outcrop a feldspar porphyry dyke, an apophysis of the Shebandowan Lake Stock, cutting across the peridotite.

STOP 2, Fig. 6

Approximately 3 km w-nw of Stop 1 is a narrow unit of serpentized peridotite hosting Ni-Cu sulphides hosted by banded mafic volcanics. These rock units are exposed in a small trench on the side of a hill, approximately 75 metres north of the Crayfish Creek Fault. Points of interest at this outcrop are: the banding within the mafic volcanics, the thin wedge of ultramafic and the breccia sulphides. The latter consist of numerous mafic to ultramafic volcanic fragments cemented by pyrrhotite, chalcopyrite and pentlandite. This sulphide zone averages 1.88% Ni and 1.61% Cu. Note the porphyry dyke, an apophysis of the Shebandowan Lake Stock, intruding the sulphides.

STOP 3, Fig. 7

4.5 Km w-sw of the Shebandowan mine headframe is a typical exposure of ultramafic rock, common in the area, interdigitated with felsic volcanic fragmentals, mafic volcanics and and intruded by a felsic dyke. With the exception of portions of the ultramafic rock and the dyke, the entire outcrop is strongly foliated to schistose. Well developed mineral aggregate lineations and clast elongations are vertical. Stratigraphic tops are to the south, as

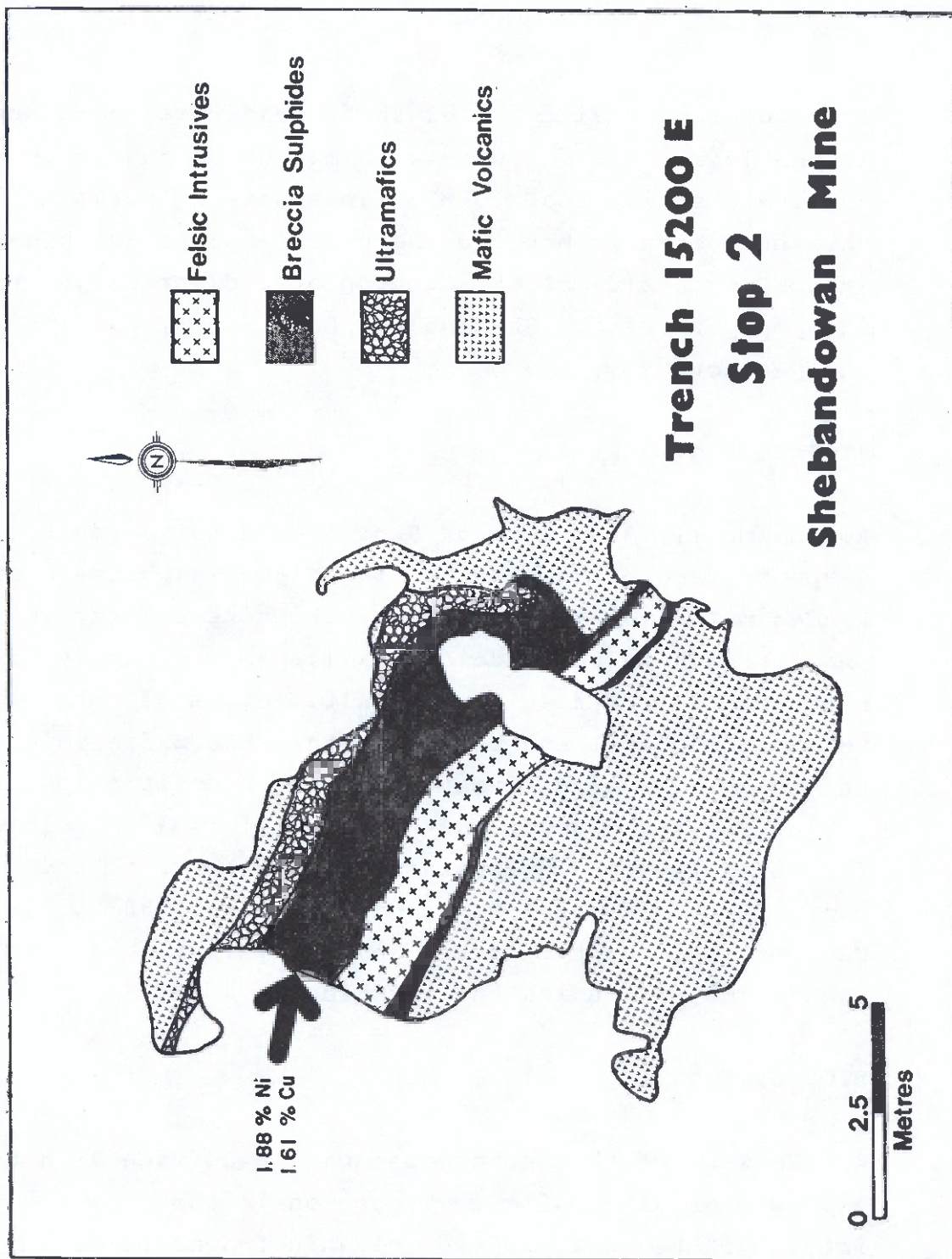


Figure 6: Stop 2 - Trench 15200E - Shebandowan Mine

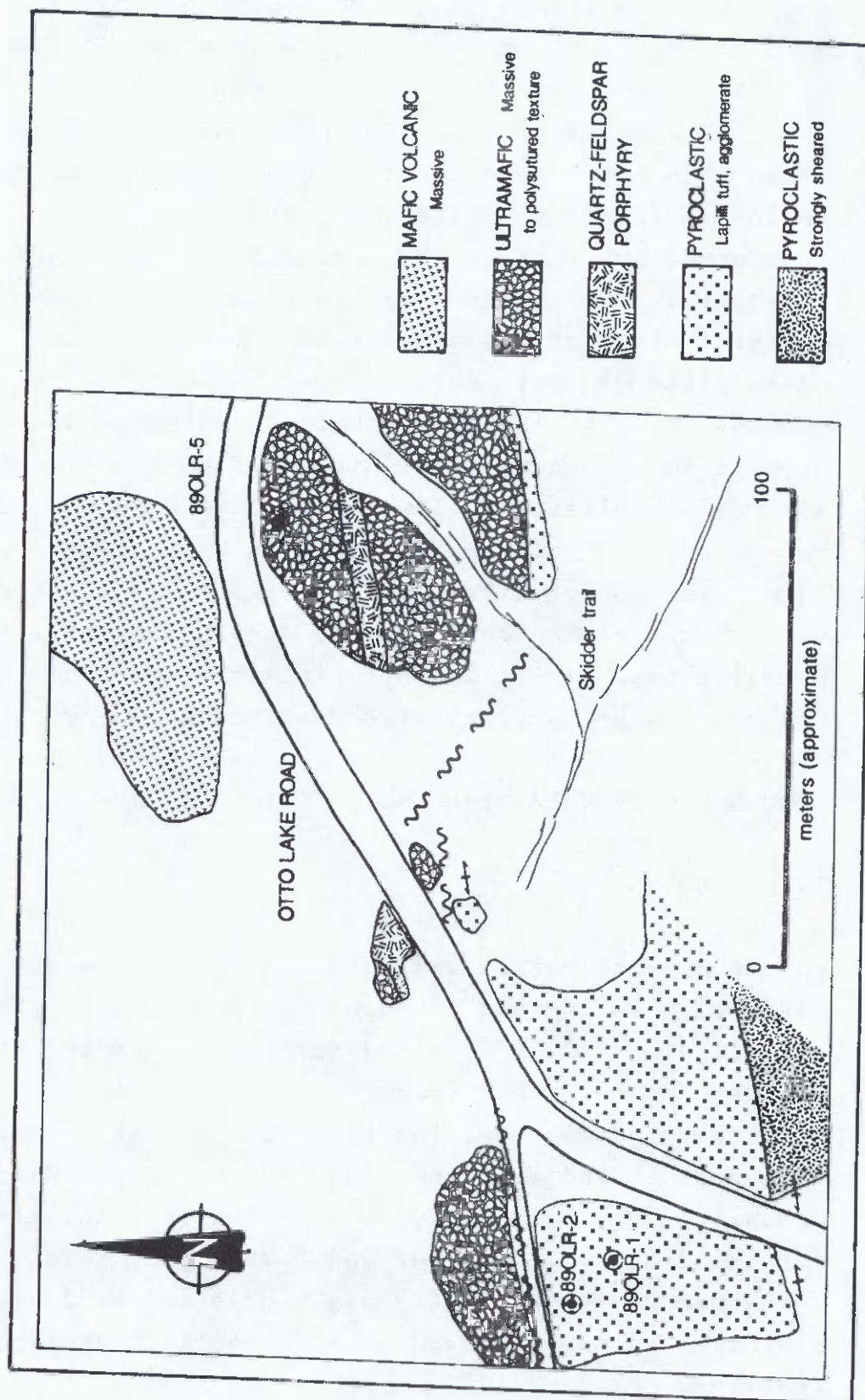


Figure 7: Stop 3 - Otto Lake Rd.

interpreted from the occurrence of black chert fragments in the lower few metres of the felsic fragmental at site A on Fig. 7. These clast are interpreted to have been derived from an underlying chert bed, exposed at site B on Fig. 7.

The margins of the ultramafics are schistose, and more often than not, covered. The relationship with the enclosing felsic volcanic rocks is therefore unclear. The undeformed interior of the ultramafic is fine grain, and has polyhedral jointing. It can be argued that these two textural observations support an extrusive origin for this unit, although they could be considered equivocal. The common juxtaposition with chemical sediments also lends some support to an extrusive origin. Definitive flow features such as spinnifex and flow top breccia have not been observed in the ultramafic rocks in this area. The positive identification of ultramafic flow rocks in the area would support the contention that the Shebandowan Mine is a Kambalda type nickel deposit since the host rock at the mine has had its primary textures destroyed by shearing.

STOP 4; North Coldstream Mine, Burchell Lake

INTRODUCTION

Copper mineralization was discovered on the north side of Burchell Lake in 1872. Copper was produced in 1903, 1906, 1916-17, 1957-58 and 1960-67 for a total of 102 000 000 pounds. Most of this production occurred from 1960 to 1967 when the mine was operated by Noranda at a capacity of 1000 tons per day and was producing 14 000 000 pounds of copper annually. The mill produced concentrate containing 26% copper, 2.63 o.p.t. silver and 0.09 o.p.t. gold. The ore contained 2% copper, 0.012 o.p.t gold and 0.22 o.p.t. silver. The mine closed as the result of depleted reserves.

REGIONAL GEOLOGY

The North Coldstream Mine is located in the western portion of the Shebandowan greenstone belt. Stratigraphic and structural trends are northeasterly. Outcrops have well developed foliation and are often schistose. According to Stott and Schwerdtner, 1981, deformation in this area was dominated by sinistral shear. The mine is on the southwestern limb of an anticline with an axial plane striking northeast (Fig. 8). The anticline is cored with felsic volcanics which are overlain by mafic volcanics. At a regional scale, the mine is at this contact, with felsic volcanics as stratigraphic footwall and mafic volcanics as hangingwall. Along its entire length, this mafic-felsic volcanic contact is a 1 km wide schist zone. At the minesite, this contact was intruded by a gabbroic stock which is the immediate footwall to the mine. Foliation orientation at the minesite shows a departure from the regional orientation, in that it consist of an enclave of east trending foliation, with dextral kinematic indicators. Wether or not this foliation orientation is independent of felsic plutons, both to the north and south of the mine is as of yet undetermined.

MINE GEOLOGY

Copper mineralization at the North Coldstream Mine consist of lenses with a high density network of chalcopyrite and pyrite veinlets (stringers). Some massive and disseminated mineralization also exist. As seen on the 500 foot level plan (Fig. 9), cross-section geometry of these lenses is variable, with maximum horizontal dimensions of 200 by 200 feet. The greatest dimension of these ore lenses (plunge) is oriented at 50 degrees to the east. Giblin, (1964) reported three types of coarse grain veins which crosscut the orebodies. The most common type of vein consist of

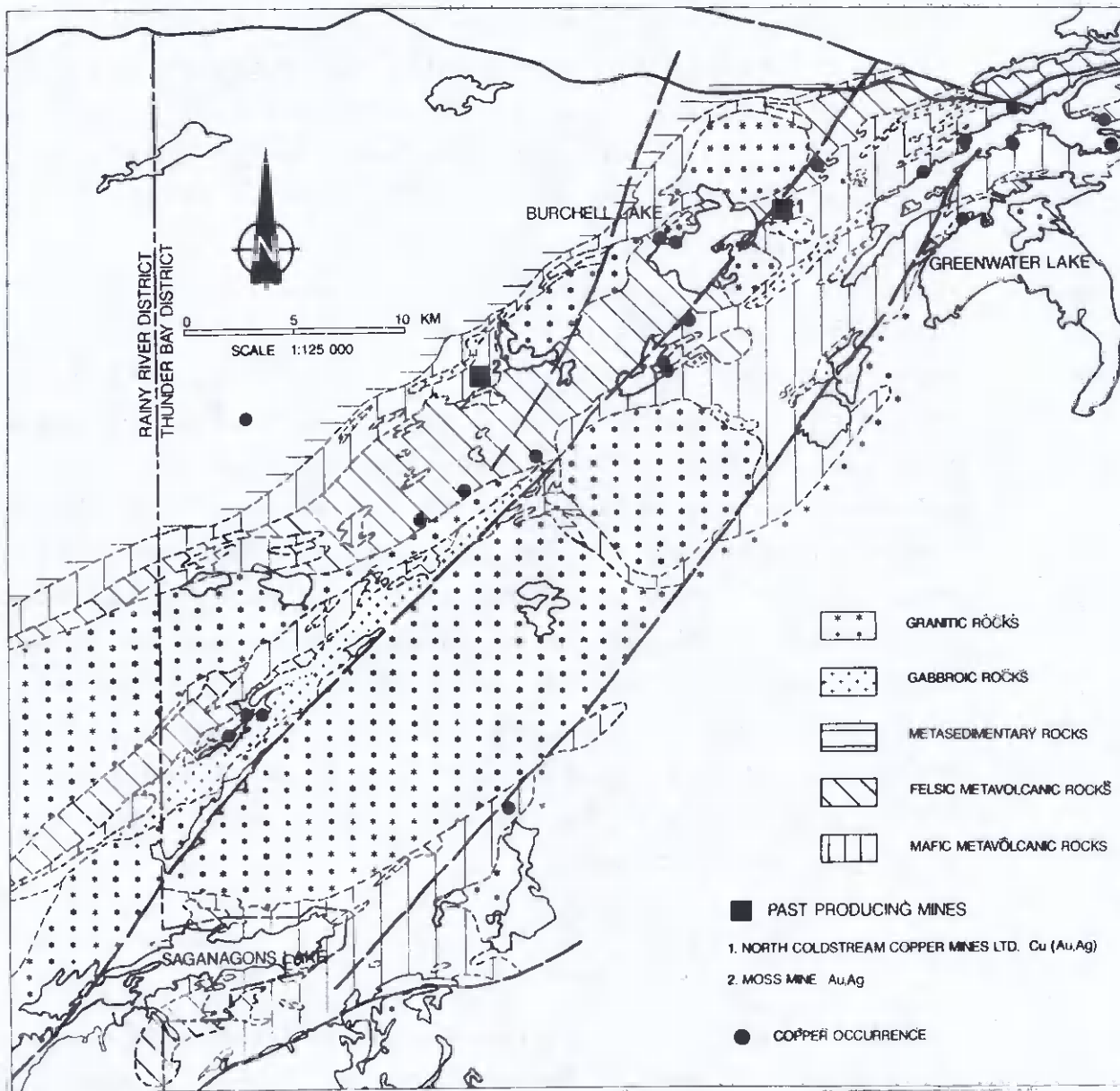


Figure 8: Geology Western Shebandowan Volcanic Belt

quartz, carbonate, pyrite and chalcopyrite. Carbonate veinlets constitute the second commonest type. Relatively rare are veins consisting of quartz, carbonate, barite, purple fluorite, dark brown sphalerite and chalcopyrite.

The ore is hosted by a massive siliceous lenses, 1000 by 400 feet, consisting almost entirely of aphanitic quartz and historically has been referred to as chert, with or without genetic connctations. Locally, fabric is discernable megascopically, and generally in these cases the siliceous rock is white. The most common colour, and always in close association with the ore, is brownish mauve. A subtle variant to this last colour is buff, and appears to be restricted to the south central portion of the mineralized complex, in close proximity to a non-siliceous buff alteration of mafic rock.

The dominant lithology surrounding the siliceous rock is a chlorite schist, then sericitic schist. The northern band of chloritic schist separates the ore complex from a gabbroic stock. At the eastern end of the complex a lens of sericitic schist separates this chlorite schist from the ore complex. At the western end, the chlorite schist contains abundant bluish quartz segregations and magnetite. All contacts between the ore complex and the sericitic schist are sharp, and the smaller siliceous pods are boudin like in that the sericitic foliation wraps around them. The southern contact has been described by Scott, (1963) as sharp, although as described above, there appears to be a transition locally from buff siliceous to buff alteration, in less strained rock. The sharp contacts are most likely the products of shear juxtaposition and the destruction of gradational contacts. The gradational contact between the gabbro stock and the chlorite schist indicate a gabbroic protolith at this location for the schist. Giblin (1964), noticed an increasing amount of bluish quartz in this

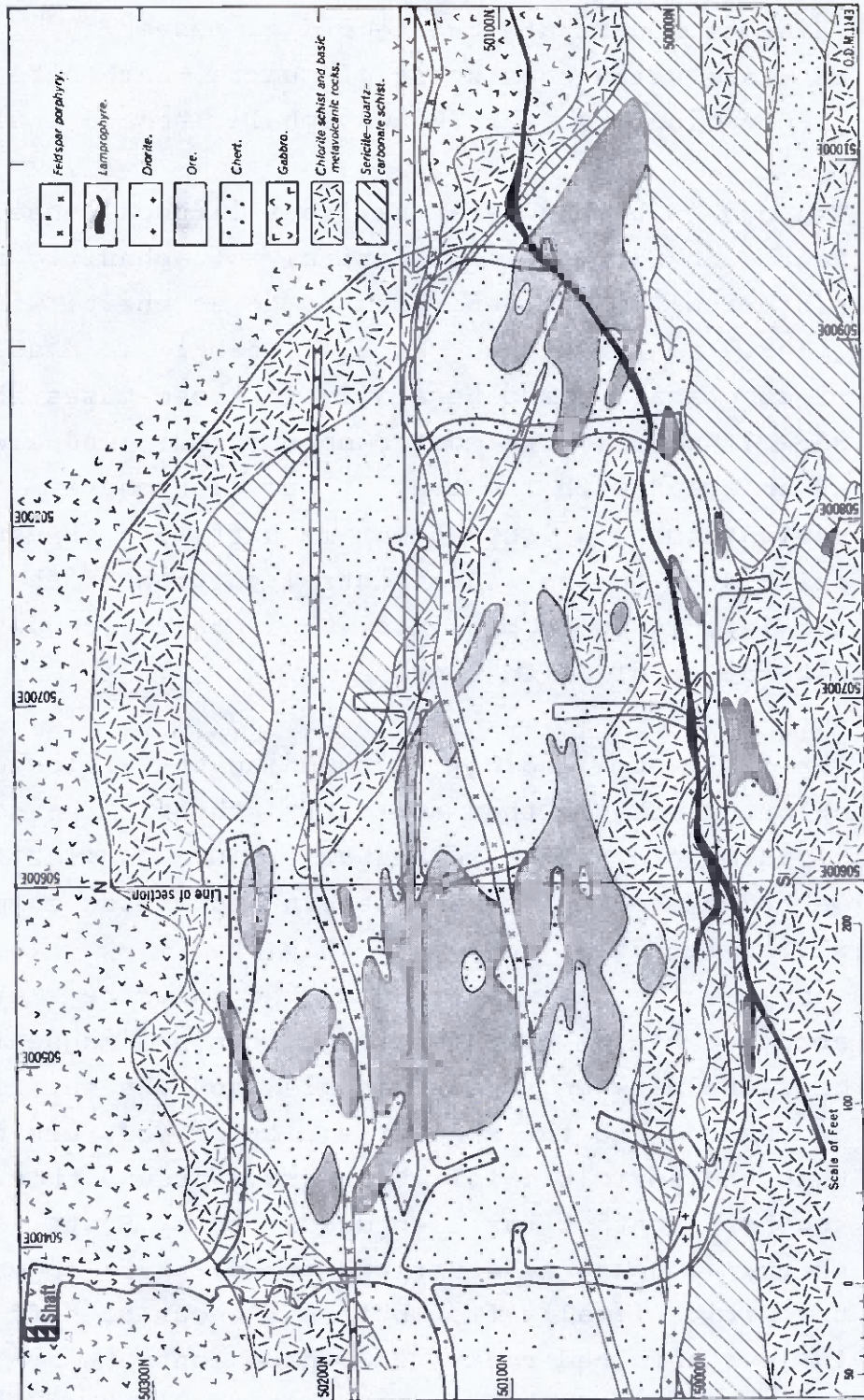


Figure 9: Geology 500' Level
North Coldstream Mine

chlorite schist towards the siliceous rock. This may be interpreted as increasing silicification of the gabbro. The protolith for the chlorite schist at the southern margin of the ore complex may be mafic volcanic, although a gabbroic protolith is also possible.

The protolith for the sericitic schist may be felsic volcanic as indicated by poorly preserved, questionable, fragmental texture. As intense alteration is probable, the protolith to the sericite schist could also be mafic volcanic or intrusive.

GENESIS OF THE NORTH COLDSTREAM MINE

Several fundamental questions must be answered in order to contemplate the origin of the North Coldstream Mine. What is the siliceous host? -chert or silicification? What is the host rock? -mafic or felsic volcanic -or gabbro? The massive nature of the siliceous zone, the lack of sedimentary features, and other sedimentary lithologies, its absence on strike, and the gradational contacts all support silicification. Silicification of this intensity is only known to occur in hydrothermal gold deposits (of all ages), and in the Keiko zone of some Kuroko type volcanogenic massive sulphide deposits. The former can be negated as a gold content at the North Coldstream of 450 ppb is much too low, and is typical of the latter. Silicification in gold systems tends to be "late", ie post tectonic, is of the highest ore grades and is superimposed on fabrics. Silicification at this site predates deformation as evidenced by siliceous boudins and inherited fabric. A volcanogenic origin is also supported by the "stringer" style of the ore, the piped shaped host silicification and to some extent, the sequence of alteration. It therefore appears that the North Coldstream Mine is the stringer zone of a volcanogenic massive sulphide deposit.

At one location there is evidence that the gabbro may be a host rock. If one considers the apparent increasing silicification of the gabbro to indicate its presence during the mineralizing event, then the mineralizing event must be post volcanic. However, the lack of fabric within the gabbro at the east end of the mineralized complex indicates that the gabbro is post tectonic, and therefore post mineralization. Taking this under consideration, the chloritic schist along the margin of the gabbro may be related to emplacement and the blue quartz generated during feldspar destruction and chloritization.

This mineralized complex appears to be located at a felsic-mafic volcanic contact. This is certainly true at a regional scale, but difficult to demonstrate at mine scale.

FIELD TRIP STOPS

STOP 4a;

This is an outcrop of altered hornblende gabbro on the northern margin of the mineralized complex. Despite epidote-chlorite alteration and multiple shear sets, igneous textures and mineralogy are preserved. This gabbroic stock has variable composition and texture and at this location we find a medium grain, equigranular gabbro and a pegmatitic phase. In thin section the hornblende is fibrous and secondary.

STOP 4b;

Group of outcrops on south side of road, only 25 feet from stop 4a. These outcrops have rare, ghost like, relic igneous textures which indicate that the protolith is gabbro. The gabbro has been sheared and increasingly chloritized at the

expense of epidote. The chlorite is pervasive and also occurs as segregation. Blue quartz segregations and magnetite are irregularly distributed. Leucoxene and magnetite are dispersed as sub-millimetre grains, magnetite also occurs as acicular crystals up to 5 cm in length. Some layers in the chlorite schist are in fact, fibrous hornblende, which are later than the magnetite-leucoxene alteration, and have a more random orientation than chlorite, indicating post tectonic growth. The distribution of quartz as irregular blebs, and the volume of this quartz could be produced in situ during shearing and chlorite alteration, as opposed to silicification from an external source. Those quartz rich portion of the chlorite schist zone are termed "quartz complex" in mine terminology.

STOP 4c;

This white siliceous zone is the transition from the quartz complex to the more massive brownish mauve host rock. It consist of strained quartz and 10 % sericite and leucoxene.

STOP 4d;

This brownish mauve siliceous rock is the ore host. It's only dissimilarities with the white siliceous unit is that it unstrained, and as such is an even grain mosaic of fine, equant quartz crystals,

STOP 4e;

A third variety of the silicification is buff coloured and only a subtle difference exist with the brownish mauve variety. This variety is spatially related to a buff alteration zone along the south central portion of the mineralized complex. This buff alteration is the

sericitization of a felsic rock bearing quartz eyes, possibly felsic volcanic. The colour of the silicification may reflect the composition of the protolith.

STOP 4f;

This is an outcrop of the buff alteration described in 4e. The rock is massive, fine grain, and textureless.

STOP 4g;

An outcrop of ore in sharp contact with a sericite schist. At the top of the outcrop, fragmental textures of equivocal origins and a siliceous boudin wrapped in sericitic schist.

STOP 4h;

This buff outcrop is included in the sericite schist unit because of its felsic appearance, although it may contain more chlorite than sericite. At this location a fragmental texture may be interpreted as volcanic, but is considered an equivocal texture.

STOP 4i;

This large ridge at the east end of the mineralized complex is gabbro. The lack of fabric indicates that it was not affected by the deformation that produced the adjacent schist and must therefore be younger and crosscut the ore complex. It cannot therefore be the host rock.

STOP 5; Vanguard Prospect

INTRODUCTION

The Vanguard Prospect is situated south of the junction between Highway 802N and Highway 11 at Kashabowie, Ontario (Fig. 8). A private bush road trends south of this intersection and accesses the showing area. The prospect is situated on Mining Locations K56 and Z71.

The property is private land and permission from the owners is required to visit the prospect. The property is currently under option to Minnova Inc.

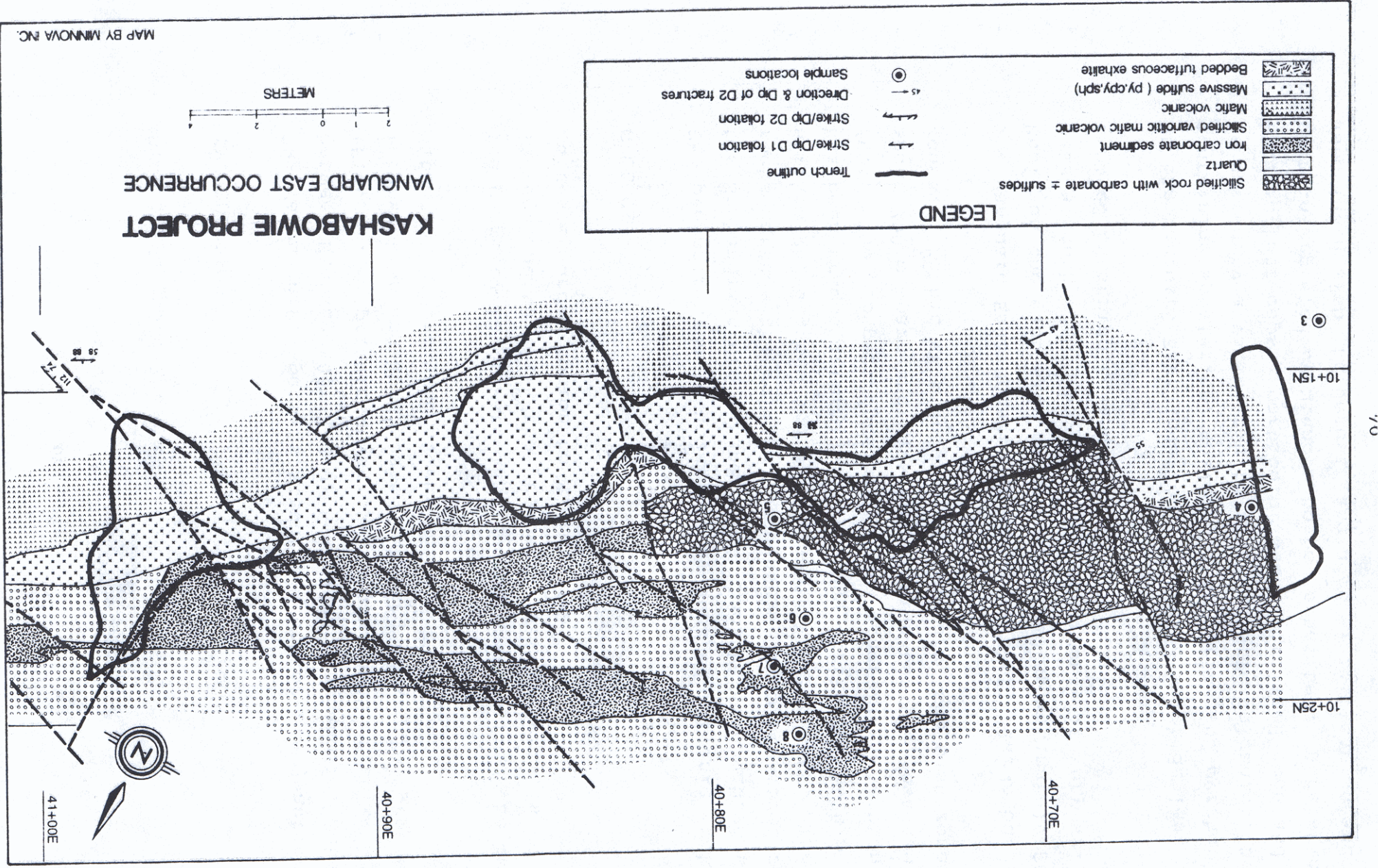
GEOLOGY

The prospect area is underlain by sheared volcanic rocks of the Shebandowan Greenstone Belt. Anorthositic gabbro suite rocks intrude the volcanic sequence. The mineralized zone strikes 060 degrees and dips steeply to the south. The zone has been traced intermittently along strike for 1310 m (4300 ft.). The prospect consists of two mineralized sections about 700 m apart and have been designated the West and East zones. At this stop we will be examining the east zone.

The Vanguard East Occurrence forms part of a large outcrop knoll on the east side of the bush road approximately 800 metres south of Highway 11. The occurrence is very well exposed by stripping and trenching.

At the Vanguard East location, a massive sulphide lens containing sphalerite, chalcopyrite, pyrite and pyrrhotite is hosted by a variolitic to massive mafic volcanics; bedded silicic tuff caps a portion of the mineralized zone and has been interpreted by Minnova to be "exhalite". Alteration associated with the deposit includes silicification and iron

Figure 10: Geology of the Vanguard East Occurrence



carbonate development, coupled with Na and Ca depletion. The rocks are chloritized.

With reference to Fig. 10, the sulphide mineralization is best exposed in the main trench. A channel sample across this zone yielded 2.47% Cu, 6.92% Zn, 40.6 grams per ton Ag, and 1.39 grams per ton Au across an apparent width of 4 meters. The sulphide ore is finely banded and in the sample examined, consisted of approximately 25% sphalerite, 70% pyrite, and the balance silica, rock fragments, and other gangue minerals. Chalcopyrite also occurs in the sulphide body.

The two other major rock units exposed near the main trench are a silicified variolitic mafic volcanic and a chocolate brown (weathered surface) carbonate-rich rock. An alteration zone consisting of massive to bedded silica/carbonate with disseminated iron sulphides is situated west of the south end of the main trench. The contact area between this zone and the variolitic mafic volcanic to the west is marked by quartz (veins?).

Sheared but otherwise well preserved volcanic features such as varioles can be examined throughout the large flat stripped area north of the sulphide lines.

Most of the faulting trends approximately 270 degrees with apparent dextral offsets. The intense deformation of the rocks of the showing area is perhaps explained by the proximity of the Crayfish Creek Fault. This fault presumably underlies the low ground to the east of the stripped area.

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