FIELD TRIP GUIDEBOOK

WINNIPEG '96

GEOLOGICAL ASSOCIATION OF CANADA - MINERALOGICAL ASSOCIATION OF CANADA
ASSOCIATION GÉOLOGIQUE DU CANADA - ASSOCIATION MINÉRALOGIQUE DU CANADA
JOINT ANNUAL MEETING — RÉUNION ANNUELLE CONJOINTE
27-29 MAY/MAI 1996 — THE UNIVERSITY OF MANITOBA

WESTERN SUPERIOR PROVINCE

(FIELD TRIPS A5 and B6)

by

G.P Beakhouse[1], G.M. Stott[1], C.E. Blackburn[2], F.W. Breaks[1], J. Ayer[2], D.Stone[1], C. Farrow[1] and F. Corfu[3]

1 Ontario Geological Survey, Willet Green Centre, 933 Ramsey Road, Sudbury, Ontario, P3E 6B5

2 Ontario Geological Survey, Box 5200, 808 Robertson Street, Kenora, Ontario, P9N 3X9

3 Royal Ontario Museum, 100 Queen's Park, Toronto, Ontario, M5S 2C6

Recommended citation. Beakhouse, G.P., Stott, G.M., Blackburn, C.E., Breaks, F.W., Ayer, J., Stone, D., Farrow, C. and Corfu, F., 1996. Western Superior Province - Field Trip Guidebook A5/B6, Geological Association of Canada/Mineralogical Association of Canada Annual Meeting, Winnipeg, Manitoba, May 27-29, 1996.

WINNIPEG '96

GAC/AGC - MAC/AMC 27-29 MAY/MAI 1996

Electronic Capture, 2008

The PDF file from which this document was printed was generated by scanning an original copy of the publication. Because the capture method used was 'Searchable Image (Exact)', it was not possible to proofread the resulting file to remove errors resulting from the capture process. Users should therefore verify critical information in an original copy of the publication.

Winnipeg '96

Geological Association of Canada - Mineralogical Association of Canada
Joint Annual Meeting

Western Superior Province Fieldtrips Guidebook

G.P. Beakhouse, G.M. Stott, C.E. Blackburn, F.W. Breaks, J. Ayer, D. Stone, C. Farrow and F. Corfu

This volume consists of three parts, comprising an introduction and two guidebooks describing two separate but complementary field trips in Ontario as follows:

Part A
An Introduction to the Regional Geological Setting of the Western Superior Province

G.P. Beakhouse, C.E. Blackburn, F.W. Breaks, J. Ayer, D. Stone and G.M. Stott

Part B
Western Superior Transect: English River - Winnipeg River - Wabigoon Portion

G.P. Beakhouse, C.E. Blackburn, F.W. Breaks and J. Ayer
(May 30-June 2, 1996)

Part C
Western Superior Transect: Wabigoon - Quetico - Shebandowan Portion

G.M. Stott, D. Stone, C. Farrow and F. Corfu
(May 24-26, 1996)

WESTERN SUPERIOR PROVINCE FIELDTRIPS GUIDEBOOK

TABLE OF CONTENTS

Part C

Part A
An Introduction to the Regional Geological Setting of the Western Superior Province

G.P. Beakhouse, C.E. Blackburn, F.W. Breaks, J. Ayer, D. Stone and G.M. Stott

Introduction

The Superior Province is the largest Archean craton in the world. The Superior Province can be further subdivided into subprovinces that are themselves very large and comparable in size to major elements of modern orogenic belts (Figure 1). These scale considerations suggest that the Superior Province is one of the few places where we can examine large scale tectonic elements and their interrelationships and compare these to those of modern orogenic belts. Other attributes and developments that make the central portion of western Superior Province an ideal natural laboratory for the study of the tectonic development of Archean crust include:

- The subprovince structure is particularly well developed and distinct in the western Superior Province.
- Much of this area is relatively accessible.
- Bedrock exposures are relatively abundant and freshly exposed by Pleistocene glaciation.
- The general, and in many cases detailed, distribution of lithological units is well mapped.
- There are currently hundreds of high-precision U-Pb age determinations that provide an absolute chronological control comparable to that provided by zone fossils in Phanerozoic orogens.
- The crust of the western Superior province has been imaged by a number of geophysical methods and this aspect will be augmented in the coming years by geophysical investigations carried out under the Lithoprobe Western Superior Transect Program.
- Numerous testable hypotheses have been advanced.

The goal of this field trip is to provide the participant with a flavour of the characteristics of the diverse subprovinces/domains that comprise the belt structure of the central portion of the western Superior province. The emphasis is on relationships that are relevant to the independent development of terranes and their subsequent assembly and stabilization in the context of a uniformitarian, accretionary tectonic model. The intent is to stimulate discussion and further work that will both develop and challenge this model.

Subdivisions of the Western Superior Province

It has long been recognized that the Superior province is divisible into a number of generally east trending entities that have been variously described as belts, blocks and subprovinces. Card and Ciesielski (1986) summarized the historical development of subdivision terminology and proposed a subdivision that is widely used today (Figures 2 and 3). These authors recognized 4 types of lithotectonic domains within the Superior Province that are briefly reviewed below.

Volcanoplutonic Domains

Volcanoplutonic domains include the Sachigo, Uchi, Bird River, Wabigoon and Abitibi-Wawa subprovinces. Of these, the Bird River, Wabigoon and Wawa subprovinces will be examined during this field trip. Volcanoplutonic domains are comprised of dominantly volcanic supracrustal sequences (the greenstone belts) intruded by syn-volcanic to post-tectonic granitoid plutons. The magmatic components of the greenstone belts include ultramafic to felsic varieties with tholeiitic, calc-alkalic and alkalic affinities. Ultramafic and mafic varieties are predominantly effusive whereas pyroclastic deposits are well represented among the more felsic varieties. The sedimentary component of the greenstone belts includes both clastic and chemical deposits. The proportions of different supracrustal rock types varies from belt to belt (Ontario Geological Survey, 1991). Plutonic rocks in the volcanoplutonic domains include synvolcanic tonalitic, quartz dioritic and granodioritic plutons, the emplacement of which has deformed the greenstone belts into arcuate forms. Later plutons tend to be smaller and more compositionally diverse, ranging from dioritic to granitic or syenitic.

Metamorphic grade in the greenstone belts is generally greenschist or sub-greenschist grade except for narrow belts or the margins of larger belts which commonly display mineral assemblages typical of low-pressure amphibolite grade.

Metasedimentary Domains

Metasedimentary domains include the English River, Quetico and Pontiac subprovinces. Clastic metasedimentary rocks are the predominant supracrustal component. These rocks have been interpreted to represent wacke, siltstone and shale derived from adjacent volcanoplutonic domains and deposited in extensive basins from turbidity currents (Breaks, 1991, Williams, 1991). Mineral assemblages indicative of low metamorphic grades occur locally near the margins of these domains but they are, for the most part, characterized by upper amphibolite to granulite grade metamorphism and widespread partial melting of the metasedimentary gneiss.

Granitoid rocks include strongly peraluminous granite to granodiorite derived from the partial melting of the metasedimentary rocks as well as sheet-like intrusions and elliptical plutons and batholiths of meta-aluminous to mildly peraluminous granite to tonalite.

Plutonic Domains

Plutonic domains include the Berens River and Winnipeg River subprovinces and possibly the central portion of the Wabigoon subprovince. These domains are characterized by the dominance of intermediate to felsic plutonic rocks. Supracrustal rocks comprise less than 10 % of these

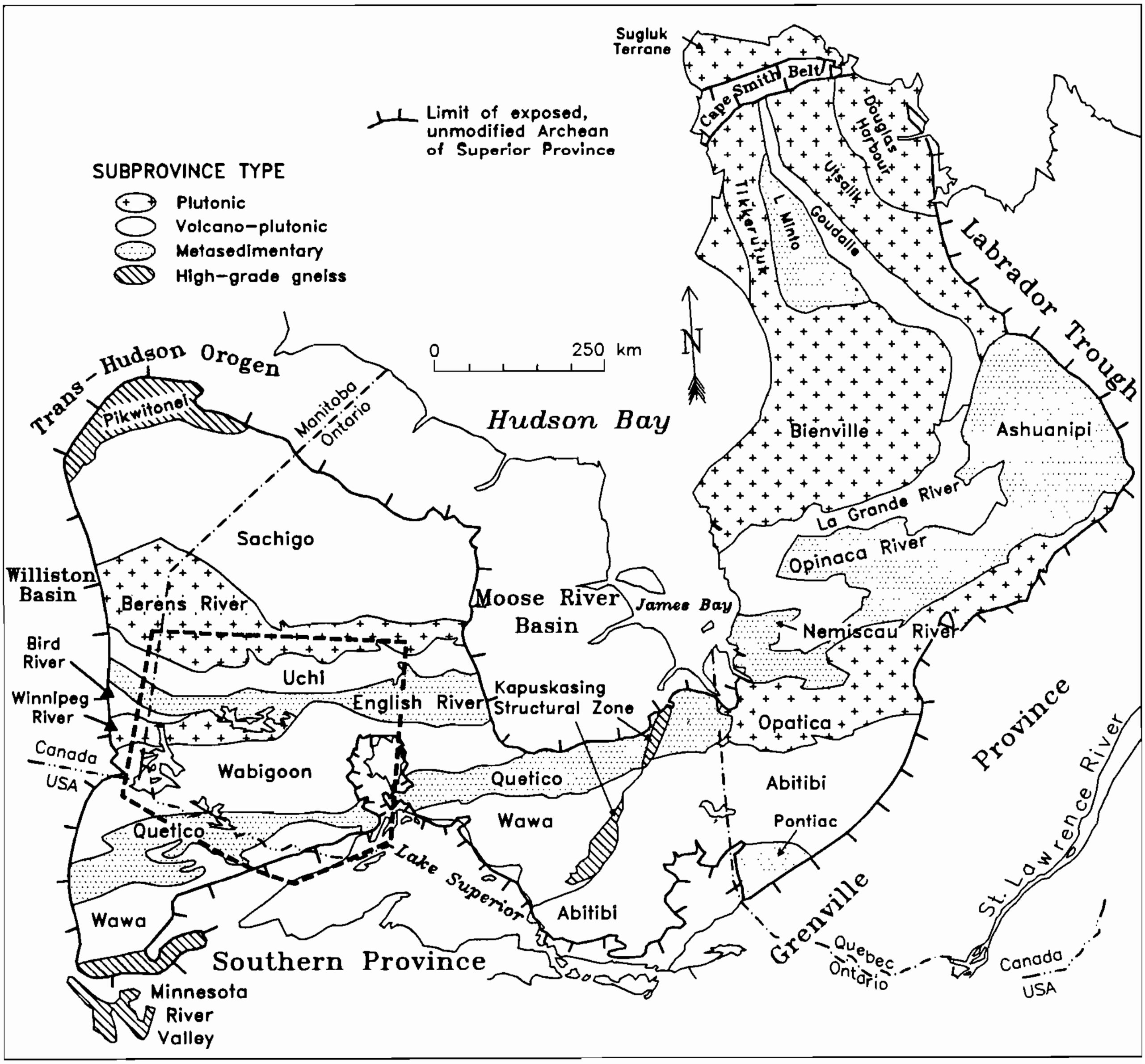


Figure 1. Subdivisions of the Superior Province. The dashed line indicates that portion of the Superior Province covered in figures 2 and 3.

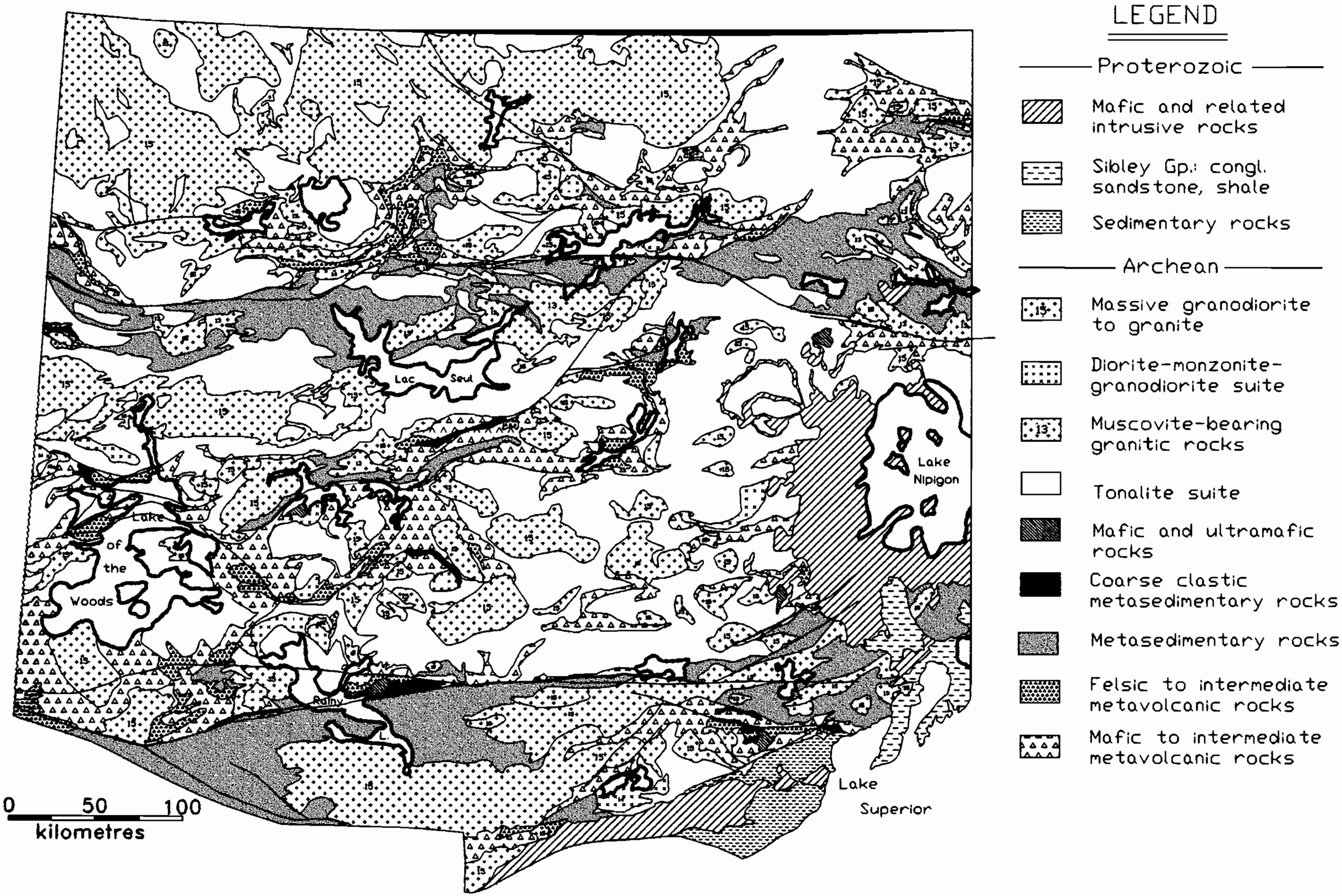
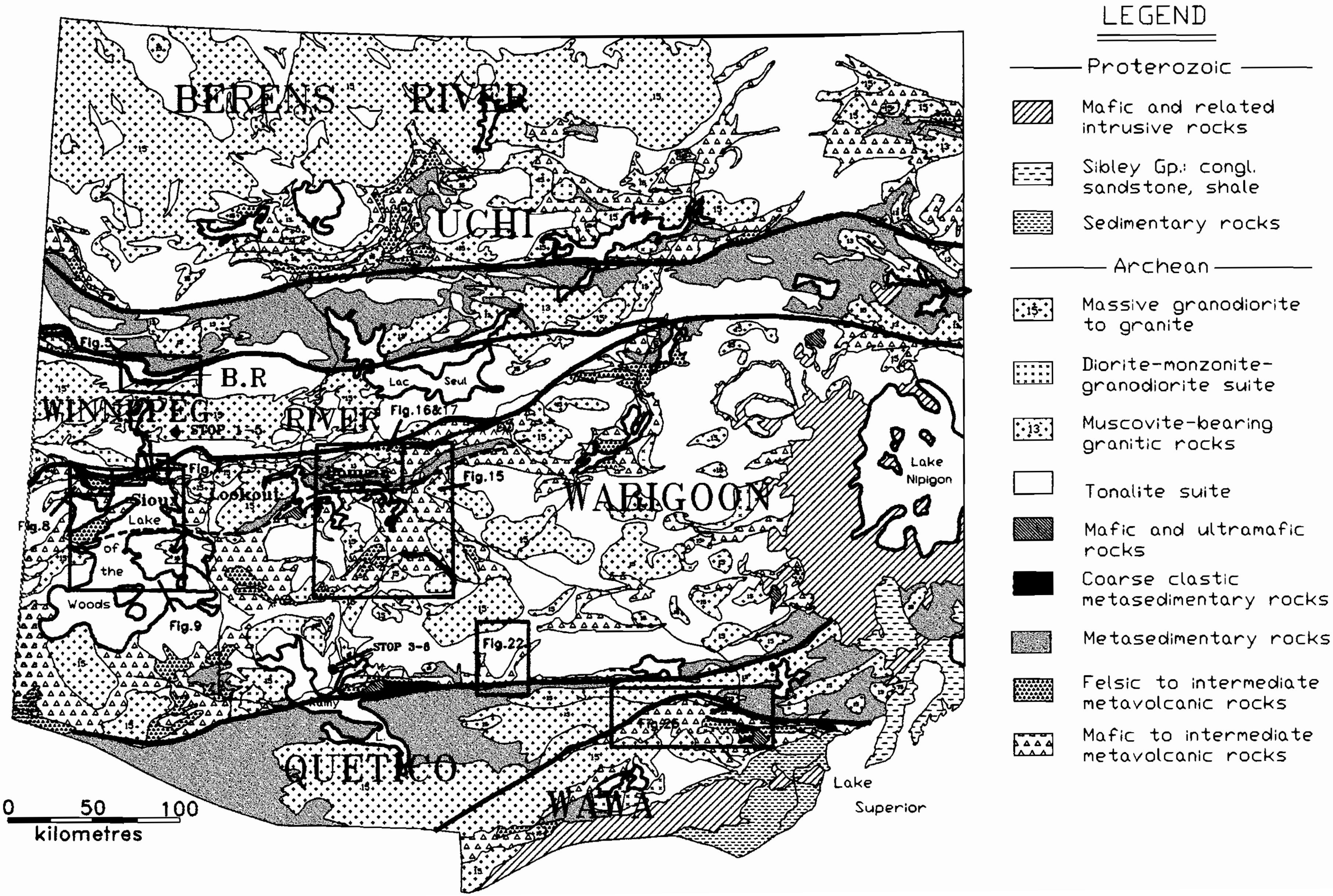


Figure 2. Geological map of the central portion of the western Superior Province (modified from OGS map 2542).

Figure 3. Geological map of the central portion of the western Superior Province illustrating the location of subprovinces and domains, more detailed maps and several isolated stops.



domains and are dominantly of mafic metavolcanic parentage. Granitoid rocks include earlier, foliated to gneissic, tonalite-granodiorite-quartz diorite as well as later, more massive granodiorite to granite. These domains are characterized by upper amphibolite to granulite grade metamorphism.

High-grade Gneiss Domains

High-grade gneiss domains are not available for examination in the area covered by this field trip. Many of the high-grade domains occur near the margins of the Superior Province (e.g., Pikwitonei) or along discordant structures (Kapuskasing) where latest Archean or Proterozoic tectonic processes have exposed deeper levels of Archean crust, often with a preserved transition into higher crustal levels (Percival and Card, 1983; Card, 1990). This has led to the interpretation that mid-crustal levels exposed in the Kapuskasing may be representative of the buried mid-crust elsewhere in the Superior Province. It is not clear if higher metamorphic grade portions of the other types of domains (e.g., plutonic Winnipeg River subprovince) are transitional into Kapuskasing-like mid-crust. An alternative point of view is that high-grade gneiss domains might be exposed because they are unique and differ from the crust beneath the rest of the Superior Province.

Subprovince Summaries

Following are summaries of the geology of subprovinces that we will be examining during the field excursion. References in the headings refer to recent review articles from which these summaries have been gleaned and to which the reader is referred for more details and references. The geochronogical data referred to in these summaries is also reviewed by Corfu and Davis (1991).

English River Subprovince (Breaks, 1991)

The English River subprovince is a linear belt up to 50 kilometres wide and at least 800 kilometres long, extending an unknown distance beneath unconformably overlying Paleozoic strata at both ends of its exposed extent. The subprovince is composed predominantly of highly metamorphosed and migmatized clastic metasedimentary rocks and compositionally diverse intermediate to felsic plutonic rocks (Figures 2 and 3). The protolith of the dominant metasedimentary component is interpreted to be wacke, siltstone and mudstone originally deposited in a submarine fan or abyssal environment based on:

- pelitic compositions
- widespread preservation of centimetre to decimeter to, less commonly, metre scale, even, continuous bedding
- rare preservation of internal laminations, cross-stratification, clastic dikes, load casts, slump balls, rip-ups, ripples, flame structures and graded bedding, and
- the great extent of the metasedimentary unit.

Minor widely distributed components of the metasedimentary sequence include polymictic conglomerate, calc-silicate rock, ironstone and mafic sediment/tuff beds. The age of sedimentation is not well constrained but the presence of detrital zircons as young as 2698 Ma indicates that it is, at least in part, younger than this age.

Metavolcanic rocks are a minor component in the English River subprovince, abundant accumulations being restricted to the Melchett Lake greenstone belt in the extreme eastern portion of the subprovince. This greenstone belt is somewhat unusual for Superior Province greenstone belts in terms of the dominance of intermediate to felsic over mafic volcanic rocks.

Intermediate to felsic plutonic rocks include:

- Older (~3.0 Ga) tonalitic plutons (Chamberlain Narrows batholith) straddling the English River-Winnipeg River subprovincial boundary and possibly in tectonic contact with the metasedimentary sequence.
- Younger (~2698 Ma) tonalitic plutons (Bluffy Lake batholith and Whitewater Lake and Adamhay Lake stocks) intruding (?) the metasedimentary rocks.
- Granitic to granodioritic plutons (2698-2560 Ma) intruding the metasedimentary sequence.
- Peraluminous granitic rocks occurring as mappable units underlying ~10% of the English River subprovince and also widespread as a component of the migmatitic metasedimentary rocks. These rocks typically have a distinctive mafic and accessory mineralogy that includes biotite and muscovite and one or more of garnet, cordierite, sillimanite and tourmaline that imparts a distinctive peraluminous character to these rocks.

Minor amounts of mafic to ultramafic and anorthositic rocks are noteworthy locally.

The deformational history can be generalized as consisting of earlier recumbent thrusts and folds overprinted by pervasive, upright folding. Ductile to brittle deformation related to transcurrent motion on major faults localized near the northern boundary of the subprovince represent the last major deformational phase.

Metamorphic grade varies regionally, tending to be lower adjacent to the Uchi subprovince and culminating in a series of granulite grade nodes extending along the south-central part of the subprovince. Generally, upper amphibolite to granulite facies metamorphic conditions prevail, accompanied by widespread partial melting. Peak metamorphic conditions of 750 C and 0.54 GPa outlasted the regional upright folding event.

Bird River Subprovince

The Bird River subprovince is a volcanoplutonic domain lying between the English River and Winnipeg River subprovinces at the western edge of the exposed Superior Province. It is among the smallest of subprovinces but it widens to the west where it's extent beneath Paleozoic rocks of the Williston basin remains uncertain. The subprovince includes the Bird River greenstone belt in Manitoba and the Separation Lake greenstone belt in Ontario together with a narrow mafic volcanic septum connecting the two. Minor, disrupted mafic volcanic blocks lying along the English River-Winnipeg River boundary up to several hundred kilometres east of the Separation Lake belt may be related. Though small, the elevation of this entity to subprovince status (Card and Ciesielski, 1986) appears to be warranted based on precedence of subprovincial subdivision elsewhere in the Superior Province (Beakhouse, 1991).

The Bird River greenstone is lithologically diverse, consisting of mafic to felsic metavolcanic rocks, several discrete metasedimentary units and a distinctive chromite enriched layered mafic intrusion (Bird River sill) (Cerny et al., 1981). The Separation Lake greenstone belt is narrower and less lithologically diverse, consisting mainly of mafic metavolcanic rocks with subordinate intermediate to felsic volcanic rocks and clastic metasedimentary rocks. The two greenstone belts are physically connected by a narrow mafic volcanic septum.

It is difficult to unambiguously assign adjacent granitoid plutons to this subprovince due to its small size. Highly geochemically evolved pegmatitic granites and related rare element enriched pegmatites (Cerny et al., 1981; Breaks, 1993) comprise a distinctive component within the Bird River subprovince. One of these, the Tanco pegmatite within the Bird River greenstone belt, is a world class Ta deposit that also has significant production and reserves of Cs, Li and a wide variety of other metals and industrial minerals. The recent recognition of rare-element enriched pegmatites in the Separation Lake area (Breaks, 1993) has expanded the potential for discovery of additional deposits of this type.

The Bird River greenstone belt is structurally complex, consisting of a number of fault-bounded segments, broadly corresponding to stratigraphic units for which contrasting structural styles are reported (Cerny et al., 1981). The southern margin of the Bird River subprovince is mostly an intrusive contact with young granitic plutons, but zones of intense cataclasis occur locally (Cerny et al., 1981; Beakhouse, 1993). Amphibolite grade metamorphic mineral assemblages predominate except for the eastern end of the Bird River greenstone belt. At this latter locality, there is a regional zonation from greenschist facies metamorphism in the southern and western portions of the belt to upper amphibolite facies metamorphism accompanied by partial melting in the northeastern portion of the belt where the greenstone belt is transitional into the English River subprovince metasedimentary migmatite assemblage.

Winnipeg River Subprovince (Beakhouse, 1991)

The Winnipeg River subprovince is a plutonic domain up to 70 km wide and with an exposed strike length of 400 kilometres. The subprovince extends an unknown distance beneath Paleozoic rocks of the Williston Basin in the west. To the east, the subprovince is portrayed as pinching out west of Savant Lake, but potentially correlative rocks occur 100 kilometres to the east at Caribou Lake. The subprovince is composed of diverse plutonic rocks and a volumetrically minor amount of supracrustal rock.

The most widespread supracrustal component occurs as disrupted remnants within and intruded by gneissic tonalite. Massive and pillowed mafic volcanic rocks of subalkalic, tholeiitic affinity predominate. Chert - magnetite ironstone units are locally interlayered with the basaltic rocks. Possible examples of other rock types (clastic metasedimentary rocks, intermediate to felsic metavolcanic rocks) are rare and disputable. The origin of these rocks is commonly enigmatic due to tectonic and intrusive disruption and upper amphibolite-granulite grade metamorphism but primary structures indicative of a supracrustal origin are rarely preserved. The age of these supracrustal remnants is indirectly constrained by crosscutting tonalites ranging in age from 2830 to 3170 Ma.

A second, volumetrically minor supracrustal rocks association is restricted to the Clay Lake area where highly metamorphosed and migmatized wacke-siltstone (like that occurring in the English River subprovince and Sioux Lookout domain of the Wabigoon subprovince) occurs in the core of domes, apparently structurally overlain by gneissic tonalite.

Plutonic rocks underlie >95% of the Winnipeg River subprovince (Figures 2 and 3) and include the following suites:

- Gneissic tonalite (including leuco-tonalite and quartz diorite) occurs in linear to arcuate belts that include supracrustal remnants, have been metamorphosed and partially melted under upper amphibolite to granulite facies metamorphic conditions and are intruded by younger granitic plutons.
- Foliated tonalite, leuco-tonalite and granodiorite is transitional into gneissic tonalite. Both tonalitic suites are locally cut by mafic dikes and are among the oldest rocks in the Superior Province with ages ranging from 2830 Ma to 3170 Ma. They are interpreted to originate from the partial melting of tholeiitic basalt at mantle or lower crustal depths.
- Volumetrically minor diorite-quartz diorite-tonalite-granodiorite plutons occur as discrete younger plutons.
- Voluminous granite-granodiorite plutons range in age from 2665 Ma to 2709 Ma and have geochemical/isotopic signatures indicating they were derived from the partial melting of the earlier tonalitic suites.

Early deformation (e.g., tonalites post-dating folding in supracrustal remnants) is recognized locally but its regional extent and significance is not understood. Small and large scale folding with moderately dipping to recumbent axial surfaces and associated strong transposition is, in part, responsible for strong fabric development and imparting a distinctive banded character to some of the tonalitic plutonic rocks and supracrustal remnants. The timing of one such large scale fold located northeast of the town of Kenora is constrained by the ages of deformed and undeformed plutons to being 2707 +/- 8 Ma. The fold axes and fabrics related to the earlier deformation were subsequently deformed in large scale domical structures, many of which are cored by more massive granitoid rocks.

The upper amphibolite to granulite facies metamorphic events that affected supracrustal and tonalitic suites are cotemporal with the emplacement of granitic to granodioritic plutons. Estimates of P-T conditions range up to 750 C and 0.6 GPa resulting in localized partial melting of tonalitic and supracrustal rocks.

Wabigoon Subprovince (Blackburn et al., 1991)

The Wabigoon subprovince is a 150 kilometre wide volcanoplutonic domain that has an exposed strike extent of 700 kilometres, extending an unknown distance beneath Paleozoic strata at either end. The subprovince is further divisible into other regions or domains (Figure 3). The different regions are based, in the first instance, on contrasting lithological proportions with a dominantly plutonic region (central Wabigoon region) bisecting the subprovince which is otherwise characterized by subequal abundances of metavolcanic and plutonic rock. The Sioux Lookout and Beardmore-Geraldton domains have anomalous abundance of metasedimentary rock with respect to the remainder of the Wabigoon subprovince and differ in several other important respects as well. Their distinction and characterization is more interpretive and controversial and much of the discussion of their significance is consequently deferred to the guide portion of this volume. The approach adopted here is to first discuss the more completely mapped and better understood western Wabigoon region and follow with discussion of the contrasting characteristics of the other regions and domains.

Western Wabigoon Region

The western Wabigoon region is characterized by interconnected, arcuate, metavolcanic dominated 'greenstone belts' surrounding large elliptical batholiths. The metavolcanic component of greenstone belts includes minor ultramafic (komatiitic), through abundant mafic (tholeiitic, calc-alkalic and minor alkalic) to felsic (mostly calc-alkalic) varieties. Except locally, metasedimentary rocks are volumetrically minor but diverse including turbiditic, volcaniclastic deposits, alluvial fan- fluvial deposits and chemical (magnetite ironstone and chert) deposits. Stratigraphic sequences generally comprise a basal, laterally extensive, mafic metavolcanic sequences overlain by laterally limited, diverse mafic to felsic sequences. Minor clastic metasedimentary deposits are associated with some of the intermediate to felsic volcanism. Very locally, coarse clastic-dominated metasedimentary sequences with subordinate chemically distinct metavolcanic rocks unconformably overlie the diverse volcanic sequences. The principal exception to the generalized lithologic proportions outlined above occurs in an area north of the Wabigoon fault in the Dryden Sioux Lookout area where metasedimentary rocks predominate. A more detailed review of the stratigraphy and geochemistry of western Wabigoon greenstone belts is presented in the guide portion of this volume.

Granitoid rocks within the western Wabigoon region include large elliptical to multi-lobate batholiths that define the architecture of the greenstone belts as well as smaller stocks. Most of the large batholiths (Aulneau, Atikwa, Sabaskong) range compositionally from ultramafic to granitic but are predominantly tonalitic to granodioritic. These are closely associated petrogenetically and temporally with the metavolcanic rocks of the greenstone belts and are interpreted to represent sub-volcanic chambers that have risen into their own volcanic ejecta. Exceptions to this generalization include the Dryberry and Ghost Lake batholiths which are younger than and compositionally distinct from metavolcanic rocks in this area. The smaller stocks are predominantly late- to post-tectonic and range compositionally from diorite to granite and syenite. Minor, late alkalic intrusions (e.g., Sturgeon Narrows) occur in the Sturgeon Lake area.

The deformational style of much of the western Wabigoon region, and particularly that portion lying to the south of the Wabigoon fault (Figure 3), is dominated by structural domes cored by large batholithic masses giving rise to apparent synclinal keels of greenstone belts surrounding the batholiths. In detail, it is not possible to correlate units on either side of the apparent 'synclinal axes' and these zones of opposing stratigraphic facing correspond, in part, to faults that have juxtaposed segments of volcanic rock of contrasting ages. Laterally continuous deformation zones exhibiting complex kinematics typically occur along the central axis of the greenstone belts where greenstone sequences face one another and may be related to this faulting. The northern portion (north of the Wabigoon fault) of the subprovince has a distinct structural style reflected in linear, fault bounded panels trending parallel to the subprovincial boundary that contrasts with that of the remainder of the western Wabigoon region. Here there is evidence for early recumbent folding and thrust faulting as well as a later phase of dextral, transcurrent shear.

Greenschist-grade regional metamorphic mineral assemblages characterize much of the greenstone belts. The principle exceptions to this generalization are narrow amphibolite-grade zones that occur at the contact with granitoid batholiths and at subprovince boundaries. A particularly noteworthy exception occurs in the Dryden area where there is widespread evidence for in situ partial melting of pelitic metasedimentary rocks.

U-Pb geochronological constraints indicate that metavolcanic rocks were deposited between 2775 and 2711 Ma, and much of this in the narrow interval of time between 2740 and 2720 Ma. Large granitoid batholiths occurring to the south of the Wabigoon fault were emplaced synchronously with adjacent volcanic rocks, whereas those to the north of the fault tend to be younger than 2710 Ma. Small, post-tectonic plutons were emplaced over a 15 Ma interval commencing at ~2699Ma.

Central Wabigoon Region

The central Wabigoon region is distinguished, in the first instance, on the basis of a much lower proportion of metavolcanic rocks with respect to plutonic rocks. As a consequence of the paucity of greenstone belt lithologies and the perceived lower mineral potential, this area has not been investigated in comparable detail to the western Wabigoon region. Scattered age determinations are, with the exception of several areas occurring along the northern and southern margins of the region, within the range of those in other portions of the subprovince. These observations suggest that the contrasting lithologic proportions of the central Wabigoon region may be either random or systematic (e.g., deeper crustal level) variations within a fundamentally similar tectonic setting.

Other observations suggest that the differences between the western and central regions of the Wabigoon subprovince may be more fundamental:

- Near the northern and southern limits of the central Wabigoon region, some tonalitic plutons cut by mafic dikes as well as unconformably overlying metavolcanic sequences are significantly older (~3000 Ma) than those occurring elsewhere in the Wabigoon subprovince.
- Some greenstone belts are older (~2.9 - 3.0 Ga) than western Wabigoon greenstone belts and contain a distinctive metasedimentary component (quartz arenites and stromatolitic carbonates) and have a higher proportion of komatiitic metavolcanic rocks.
- Several large granitoid plutons appear as prominent magnetic highs and have granodioritic to granitic compositions, similar to Winnipeg River subprovince granitic suite plutons and the Dryberry batholith and unlike the tonalitic to granodioritic syn-volcanic batholiths (e.g., Aulneau, Atikwa).

These observations suggest that the central Wabigoon subprovince is, at least in part, comprised of an older volcanoplutonic terrane, the relationship of which to the younger volcanoplutonic terranes of the Wabigoon subprovince remains enigmatic.

Eastern Wabigoon Region

The eastern Wabigoon region, lying to the east of Lake Nipigon, is similar in many respects to the western Wabigoon region. Most of this region is characterized by arcuate greenstone septa wrapping around ovoid to multi-lobate granitoid batholiths. The southern portion of the region, adjacent to the Quetico subprovince, has a higher proportion of metasedimentary rocks and a more linear geometry reflecting laterally continuous, fault-bounded panels of alternating metavolcanic and metasedimentary units. This portion of the Wabigoon subprovince will not be examined as part of this field excursion.

Quetico Subprovince (Williams, 1991)

The Quetico subprovince is a linear belt, 15-60 kilometres wide and at least 800 kilometres long. To the west it extends a further unknown distance beneath poorly exposed areas and Paleozoic cover. The Opatica belt, to the east of the Kapuskasing structural zone, may be in part equivalent.

The Quetico subprovince is similar in many respects to the English River subprovince. Variously migmatized metasedimentary rocks, originally consisting of wacke and siltstone, predominate (Figure 2). Minor ironstone and conglomerate are also components of this supracrustal assemblage. Primary sedimentary structures, including even, continuous metre to decimetre thick bedding, graded bedding, scour and rarer ripple marks, cross-stratification and dewatering structures, are preserved in many areas, particularly those characterized by lower grades of metamorphism and less pervasive migmatization. The compositions and sedimentary structures have been interpreted to indicate that the supracrustal assemblage represents deposition of immature detritus from turbidity flows in a submarine basin of great lateral extent.

Granitoid rocks include late, massive to foliated granodiorite to granite and minor, earlier, foliated to gneissic tonalite. The granitic varieties are further subdivided into a mildly peraluminous biotitic granite and a more strongly peraluminous, muscovite-bearing variety. The latter locally contains cordierite, garnet and/or sillmanite and, in the Georgia Lake area, is spatially associated with rare metal-enriched granitic pegmatites.

The earliest tectonic deformation consists of layer-parallel shearing and associated folding which is related to regional scale fabric formation. This fabric, as well as migmatitic layering is subsequently deformed in a second phase of more upright folding. Later structures include small-scale shear zones as well as major faults. The major faults include those which are oriented parallel to, and lie near, the subprovince boundary (Quetico, Seine River) as well as others which transect the subprovince (Gravel River, Kapuskasing Structural Zone).

Regional metamorphism occurred synchronously with the waning stages of deformation and was accompanied by partial melting in higher grade portions of the subprovince. In general, there is a progression from lower grades near the margins to higher grades (typically 620°C and 0.3 GPa) in the central portions of the subprovince.

Wawa Subprovince (Williams et al., 1991)

The Wawa subprovince has an east-west extent of at least 600 kilometres, being truncated by the Kapuskasing structural zone in the east and extending an unknown distance beneath poorly exposed areas and Phanerozoic cover in northern Minnesota. The southern extent of the subprovince is obscured by the waters of Lake Superior and unconformably overlying Proterozoic strata of the Animikie Basin.

The subprovince comprises isolated, arcuate to linear greenstone belts with large intervening masses of granitoid rock. The stratigraphy characterizing the greenstone belts is highly variable but typically consists of one or more mafic or ultramafic to felsic volcanic cycles. Subordinate clastic metasedimentary rocks are generally either a lateral facies equivalent of the intermediate to felsic portion of the volcanic sequence or overlie the metavolcanic rocks. One distinctive assemblage (Shebandowan), characterized by shoshonitic metavolcanic rocks, coarse conglomerate and subordinate wacke and mudstone, is interpreted to unconformably overlie the lower assemblages. Many assemblages are interpreted to be tectonically juxtaposed.

Granitoid rocks internal to the greenstone belts include diverse syn-volcanic to post-tectonic stocks. The bounding granitoid complexes are poorly understood but are dominated by foliated to gneissic tonalitic to granodioritic rocks cut by relatively younger, more massive, granodioritic to granitic plutons.

The earliest structures interpreted or inferred in some of the larger greenstone belts are thrust faults and related small to large scale recumbent fold structures. These structures are overprinted by more upright, belt-parallel folds. Late, transcurrent deformation along ductile shears and faults is widespread. Mineral assemblages within the greenstone belts are indicative of greenschist to mid-amphibolite grade metamorphism with higher metamorphic grades generally restricted to smaller greenstone belts.

An Accretionary Tectonic Model

Uniformitarian accretionary tectonic models for the formation and stabilization of the Archean crust of the Superior Province have emerged as testable, unifying hypotheses. Early proponents (Krogh and Davis, 1971; Langford and Morin, 1976; Blackburn, 1980) based their interpretations on similarities to younger orogenic belts and limited geochronological evidence for southward younging across the Superior Province (the latter since corroborated by many more U-Pb age determinations (Corfu and Davis, 1992). Others have developed the accretionary hypothesis to explain, for example;

- the nature and timing (with respect to tectonic crustal thickening and high-grade metamorphism) of granitoid plutonism in the Winnipeg River and Wabigoon subprovinces (Beakhouse and McNutt, 1986; 1991),
- imbrication of a prograded clastic wedge at the Wabigoon-Quetico interface (Devaney and Williams, 1989; Williams, 1990)

- the nature and timing of regional metamorphism in the Quetico subprovince (Percival, 1989)
- out of sequence stratigraphy (Davis et al., 1988)

Recent review articles (Hoffman, 1989; Card, 1990; Williams et al., 1992) summarize these, and other, interpretations and extend the model to the development of the entire Superior Province.

Although there are differences in detail, all of the accretionary tectonic models developed to date agree in general with the concept that the central portion of the western Superior Province with which this field trip is concerned consists of Late Archean island arcs and associated accretionary prisms together with slivers of Middle Archean crust that were swept together in a collision orogenic (Kenoran) event at approximately 2700 Ma.

Discussion of specific problems and interpretations relevant to how specific segments of the field trip relate to accretionary tectonic models are embedded in the stop description portion of the guidebook. It is our hope that this approach will provoke discussion that will challenge the application or general validity of these models and contribute to the development of new and/or improved models.

Part B
Western Superior Transect: English River - Winnipeg River - Wabigoon Portion

G.P. Beakhouse, C.E. Blackburn, F.W. Breaks and J. Ayer

Day 1. Kenora to Separation Lake

The first day of the field trip is devoted to a transect proceeding north from the town of Kenora. In the morning we will examine exposures relevant to understanding the nature of the Winnipeg River subprovince and its contact relationship with the Wabigoon Subprovince. The afternoon will be spent examining outcrops in the Separation Lake greenstone belt (Bird River subprovince), including an example of one of the rare metal pegmatites discovered recently, as well as one exposure of typical English River subprovince metasedimentary migmatite.

Proceed to the junction of highways 17 and 596 in the town of Norman (between Kenora and Keewatin) and proceed north on highway 596 for 1.5 kilometres, turn right at the junction and proceed a further 0.2 kilometres to the Perch Bay Resort turnoff (Figure 4). We will park the vehicles here to examine exposures of the marginal granodiorite and its contact with mafic metavolcanic rocks of the Wabigoon subprovince.

The Marginal Granodiorite- *The marginal granodiorite is a distinctive granitic suite pluton that derives its name from its position along the Winnipeg River - Wabigoon subprovincial boundary. The pluton has a tabular form, being at least 30 kilometres long with a width ranging from 100 to 700 metres but typically approximately 200 metres wide.*

The tectonic significance of this pluton derives from the interpretation that it is the oldest common element between two temporally distinct domains and hence can be thought of as a 'stitching pluton'. The Winnipeg River subprovince (dominantly 2830 Ma to 3170 Ma tonalites and 2665 Ma to 2709 Ma granites) was apparently inactive during a period of widespread volcanoplutonic activity (2711 Ma to 2775 Ma) in the Wabigoon subprovince. The age of the Marginal Granodiorite (2709 Ma; Corfu, 1988) is closely comparable to that of large scale recumbent folding (2708 Ma; Beakhouse and McNutt, 1986,1991) that has been related to collision of an extinct (Winnipeg River) and an active (Wabigoon) magmatic arc. This collision brought subduction-related magmatic activity in the latter to an end and resulted in tectonic thickening that triggered high grade metamorphism and intracrustal melting in the Winnipeg River subprovince.

Most of the alkali feldspar component of the Marginal Granodiorite occurs as zoned, late-magmatic, microcline megacrysts that range from undeformed, euhedral to subhedral varieties to lenticular augen (the latter relating to a variously developed cataclastic texture). Locally, both varieties of megacryst occur together suggesting megacryst development took place during deformation. Gower et al.(1983) suggest that the development of the Marginal Granodiorite may

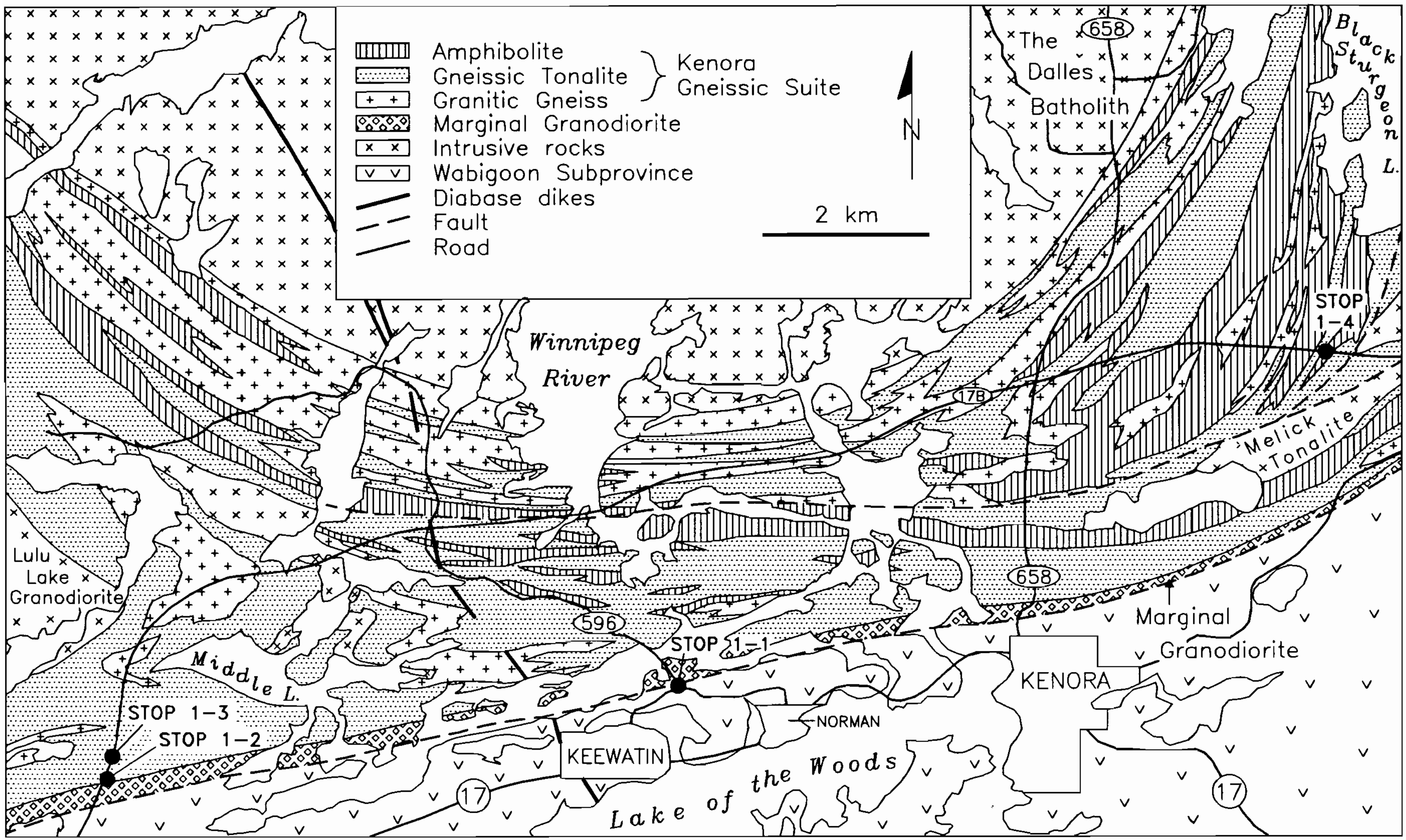


Figure 4. Geological map of the Kenora area with location of stops 1-1 to 1-4 (modified from Gower, 1978).

be related to "frictional reconstitution" of tonalitic gneiss arising from movement along a fault located at the subprovincial boundary.

Stop 1-1 Contact between Marginal Granodiorite and Wabigoon Metavolcanic Rocks

Refer to Figure 4 for location.

The exposures between the Perch Bay Resort turnoff and the Darlington Bay bridge to the north consist of variously cataclastic, microcline-megacrystic granodiorite (Marginal Granodiorite) with minor amphibolitic mafic enclaves and and weak (cataclastic?) banding. The roadcut at the top of the hill approximately 100 metres back along the road exposes a sharp intrusive contact between the Marginal Granodiorite and highly deformed mafic metavolcanic rocks of the Wabigoon subprovince.

Proceed north along highway 596 across the Darlington Bay bridge and proceed a further 3.5 km to the junction of highway 596 and the Kenora Bypass.

The waters of Darlington Bay conceal the transitional contact between the Kenora gneissic suite of the Winnipeg River subprovince and the Marginal Granodiorite that we will examine at the next stop. North of the Darlington Bay bridge are exposures of the Kenora gneissic suite.

Proceed west on the Kenora Bypass from the junction with highway 596 for approximately 5.8 kilometres. Stop just before (east of) the overpass for the old Minaki Road and railway tracks.

Stop 1-2 Marginal Granodiorite- Kenora Gneissic Suite Transition

Refer to Figure 4 for location.

This outcrop spans a transitional contact from the Marginal granodiorite into the Kenora gneissic suite. The south end of the outcrop (nearest the overpass) consists of intensely sheared, microcline megacrystic granodiorite containing a few flattened amphibolitic enclaves and highly deformed, discordant leucotonalite dikes. The north end of the outcrop is banded gneiss consisting of gneissic tonalite to quartz diorite and amphibolite together with variously deformed pegmatitic dikes. There is no clear contact, but rather the composition and texture changes subtly along with an increasing proportion of amphibolitic material and variously deformed granitic veins to the north. One narrow, disrupted mafic dike occurs at the north end of the outcrop but the origin of most of the amphibolitic material in this outcrop is enigmatic.

Kenora Gneissic Suite- *The Kenora gneissic suite is a heterogeneous assemblage consisting of amphibolite, gneissic tonalite and granitic gneiss components. The origin of the amphibolitic component is rarely unambiguous. Commonly, the amphibolite is intruded by tonalite and in some of these cases can be interpreted to represent a mafic metavolcanic rock (especially where pillow structures or cherty or ferruginous interflow sedimentary units are preserved. Rarely, the amphibolitic component is interpreted to represent deformed mafic dikes that cut the tonalite. In most cases, the origin of these rocks is indeterminate.*

The gneissic tonalite component of the Kenora Gneissic Suite is among the youngest of the tonalitic units in the Winnipeg River subprovince with U-Pb zircon ages ranging from 2830-2875 Ma (Beakhouse, 1983; Corfu, 1988). Other ages of tonalitic rocks in the Winnipeg River subprovince include those at ~3.0 Ga and 3.17Ga (Krogh et al., 1976; Davis et al., 1988; Corfu, 1988). It is not clear whether the supracrustal remnants intruded by the various ages of tonalite represent a single old supracrustal sequence or different supracrustal sequences more closely temporally associated with the enclosing tonalitic rocks.

The granitic gneiss component of the Kenora gneissic assemblage is comprised of texturally diverse, variously strained, commonly pegmatitic, granodioritic to granitic, tabular bodies that are commonly concordant to subtly discordant with respect to the other layering in the rocks. In some cases, this component occurs in 'concordant' zones where it diffusely 'soaks' the tonalitic and amphibolitic components rather than forming discrete, sharply bounded sheets.

The proportions of tonalitic gneiss, amphibolite and granitic gneiss vary widely but tends to be relatively consistent along strike and change abruptly across strike giving rise to a gneissic stratigraphy that can be mapped out.

Return to vehicles and proceed east on the bypass for 0.5 km to the next major road cut.

Stop 1-3 Kenora Gneissic Suite

Refer to Figure 4 for location.

This outcrop consists of complexly deformed gneissic tonalite with fine to coarse grained, amphibolitic mafic enclaves and variously deformed pegmatite dikes. This outcrop is typical of the components within the Kenora gneissic suite although the proportions of tonalite, amphibolite and pegmatoid rock vary widely. The origin of the amphibolitic enclaves in this, the previous outcrop and much of the Kenora gneissic suite is uncertain.

This outcrop affords good examples of the well developed banding, secondary gneissosity development, complex folding and disruption of layers that characterizes the Kenora gneissic suite. A wide variety of states of deformation of granitic pegmatoid material ranging from undeformed dikes to strongly deformed but still discordant dikes to those subconcordant sheets that may or may not be completely transposed. The Kenora suite is the most conspicuously banded of the gneissic assemblages in the Winnipeg River subprovince. This is thought to reflect a complex interaction of pre-existing planar anisotropy (beginning with bedding and progressing to secondary compositional layering) controlling the development of subsequent fabrics and generation and emplacement of igneous phases as well as intense transposition of early discordant veins and dikes.

Return to vehicles and proceed east on the bypass for approximately 13 kilometres to the junction with highway 658 (Redditt Road). Proceed a further 3.4 kilometres east on the bypass and stop at a roadcut approximately 50 metres west of a powerline crossing the highway.

Stop 1-4 Metavolcanic component of Kenora Gneissic Suite

Refer to Figure 4 for location.

Between here and the last stop, we have passed through numerous units alternately dominated by tonalite as at the previous stop or mafic rocks as in this stop. This alternation, together with across strike variation in the proportion of granitic pegmatoid material, define a gneissic 'stratigraphy'. The variation in this outcrop is a smaller scale manifestation of this mappable 'stratigraphy'.

The prominent mafic component within this outcrop, though lacking diagnostic primary structures, is nevertheless interpreted to be representative of remnants of a pre-tonalite supracrustal sequence based on analogous features and heterogeneity observed elsewhere where such features (pillows, intraflow chemical metasedimentary rocks) are preserved. The mafic component here includes:

- fine grained, banded to heterogeneous amphibolite with calc-silicate clots interpreted to reflect original pillowed or flow-breccia units. The banded character, defined by variations in colour and mineralogical proportions, is though to represent intensely transposed pillow selvages, etc. The calc-silicate clots are interpreted to be analogous to clots in lower grade metavolcanic sequences in the Superior province that are interpreted to be seawater alteration phenomena.

- fine grained, more massive amphibolite interpreted to represent massive flows interlayered with the pillowed varieties.
- medium grained, massive amphibolite representing either thin gabbroic sills or thicker massive flows.

Return to the vehicles and return to the intersection of Highways 17 (Kenora Bypass) and 658 (Redditt Road). Set the odometer to zero and proceed north on highway 658 for approximately 23.8 kilometres to the English River Road.

> Along this route we will pass exposures of the Kenora gneissic suite (0 to 3.6 km), the Dalles batholith (3.6 to 12.0 km) and back into the Kenora gneissic suite in an area with numerous small granitic stocks (12.0 to 19.0 km). Beyond this point (approximately the Ena Lake road turnoff) we will traverse the Lount Lake batholith, a large (~2000 km^2) granitic suite batholith occupying the core of the Winnipeg River subprovince.

Turn left on to the Engish River Road and proceed north for a further three kilometres. Turn right on to a gravel road. Proceed 0.6 km to the east and turn left (north) on to another gravel road and proceed 0.6 km north to exposures on the west shore of Norway Lake.

The granitic suite of the Winnipeg River subprovince*- The granitic suite plutons of the Winnipeg River subprovince are interpreted to be the plutonic signature of collisional orogeny. This suite is significantly younger (2665 to 2709 Ma) than the voluminous, dominantly tonalitic, plutonic activity in both the Winnipeg River (mostly >2.83 Ga) and western Wabigoon (2711 to 2775 Ma) subprovinces. It post-dates large scale recumbent folding and the juxtaposition of the two subprovinces (as dated by the Marginal Granodiorite) and is broadly synchronous with the timing of high-grade regional metamorphism in the Winnipeg River subprovince.*

The granodioritic to granitic spectrum of the granitic suite contrasts with the dominantly tonalitic to granodioritic spectrum for the other voluminous plutonic events referred to above. They are also geochemically distinct, even if the comparison is restricted to texturally and mineralogically similar granodiorites as are exposed in the next stop. Relative to plutonic rocks of tonalitic affinity, these rocks are characterized by high K2O, Rb, Y, REE and prominent negative Eu anomalies. They also have a crustal Nd and Sr isotopic signature that compares favourably with that which would be expected to have characterized the older tonalitic plutons of the Winnipeg River subprovince at the time of granitic suite emplacement. Consequently, these rocks have been interpreted to reflect intracrustal melting of the older tonalitic rocks and implying that virtually the entirety of the Winnipeg River subprovince was extracted from the mantle prior to 2.83 Ga and represented a buoyant, microcontinental mass at the time of Wabigoon subprovince igneous activity. The *emplacement of the granitic suite was synchronous with high grade regional metamorphism, both of which may be a response to tectonic thickening of the crust associated with the collision of an active (Wabigoon) and an extinct (Winnipeg River) magmatic arc (Beakhouse and McNutt, 1991).*

Stop 1-5 Reddit Granodiorite at Norway Lake

Refer to Figure 3 for location.

The dominant rock type around Norway Lake is a microcline-megacrystic, biotite granodiorite representative a regionally extensive phase of the granitic suite Lount Lake batholith. The microcline megacrysts are interpreted to be a late magmatic phenomena rather than phenocrysts or porphyroblasts.
A sample collected 200 metres south of this location has a U-Pb zircon age of 2702 Ma.

Retrace route back to the English River Road and proceed north for 43.1 km to the Separation Narrows bridge (English River). Continue a further 3.1 km to the Umfreville Road and then a further 1.1 kilometres to stop 1-6.

Separation Lake Greenstone Belt- *Metavolcanic and related rocks of the Separation Lake greenstone belt lie between migmatitic rocks of the English River subprovince to the north and granitic rocks of the Winnipeg River subprovince to the south. Debate continues as to assignment of the belt to either the Bird River subprovince implicit in the terminology of Card and Ciesielski (1986), followed by Beakhouse (1991) and assigned here, or to the Winnipeg River subprovince, as interpreted by Sanborn-Barrie (1988) and Blackburn and Young (1993). Amphibolite facies conditions prevail in the belt. Mafic rocks predominate, both as subaqueously extruded metavolcanics and as concordant gabbroic sills. A thin, but laterally extensive felsic pyroclastic unit occurs at the top of the sequence. The gabbros, along with intraformational magnetite iron formation units, define large folds (Figure 5), but the predominant and overall facing direction is to the north, so that the sequence faces toward the migmatites of the English River subprovince.*
The southern contact of the belt is clearly intrusive in that granitic suite plutons of the Winnipeg River subprovince intrude and break up the mafic sequence, to the extent that a separated portion of the mafic sequence lies southeast of Fiord Bay: structural trends clearly connect it with the main belt (Figure 5). There is no evidence of the cataclasis encountered elsewhere to the west along the southern contact of the Bird River greenstone belt (Cerny et al 1981).
Relationships along the northern contact of the belt are less clearly defined, and are the subject of

Stop 1-6. Between the felsic metavolcanics at the top of the sequence and the migmatites lies a mixed polymictic conglomerate and sandstone unit that contains clasts of the underlying greenstone belt and granitoid rocks to the south. Placement of a subprovincial boundary must take into account that

- *the conglomerate/sandstone unit lies directly on top of the felsic pyroclastics without apparent marked unconformity*
- *the conglomerate/sandstone unit shows evidence of onset of migmatization*
- *the felsic pyroclastic unit shows most evidence of deformation by pure shear*
- *other polymictic conglomerate remnants occur well within the migmatites to the north*

Stop 1-6 Section across the Separation Lake greenstone-English River migmatite contact.
Refer to figure 5 for location

We will walk up-section along the English River Road (figure 6), through from mafic metavolcanics with an intercalated felsic porphyry flow into the felsic pyroclastic unit. We will then traverse from the road northward across the felsic pyroclastic unit into the mixed polymictic conglomerate/sandstone unit, and thence to the transition to migmatites of the English River subprovince. On the return we will cross the road to look at intense deformation within the mafic metavolcanics.

Locality A - a comparatively little-deformed felsic porphyry unit, 10m wide, is intercalated within amphibolitic mafic metavolcanic rocks. Walking up-section, we traverse banded amphibolites derived by tectonism of mafic, commonly pillowed, metavolcanics: no flow structures remain at this state of deformation. Cross the unexposed contact between the banded amphibolites and the felsic pyroclastic unit.

Locality B - within the felsic pyroclastic unit a variety of fabrics is observed, including ghost-like remnants of lapilli-size quartz-feldspar porphyry clasts, tectonic layering superimposed on thin to medium bedding, steep to vertical dip of this layering in which lineation plunges steeply to vertical, late fractures at high angle to foliation, ptygmatic folding of quartz veins. We will walk along outcrops parallel to and along the north side of the road, and then turn to the north, crossing further examples of the felsic unit.

Locality C - polymictic conglomerate lying above the felsic pyroclastic unit. The outcrops are characterised by medium bedded alternation of conglomerate and sandstone beds. Tight chevron folding obscures bedding relationships. Mafic volcanic clasts are strongly deformed (flattened and folded) whereas most granitoid clasts have retained their rounded shape. Granitic leucosome intrudes parallel to foliation and has itself been folded, indicating continued deformation

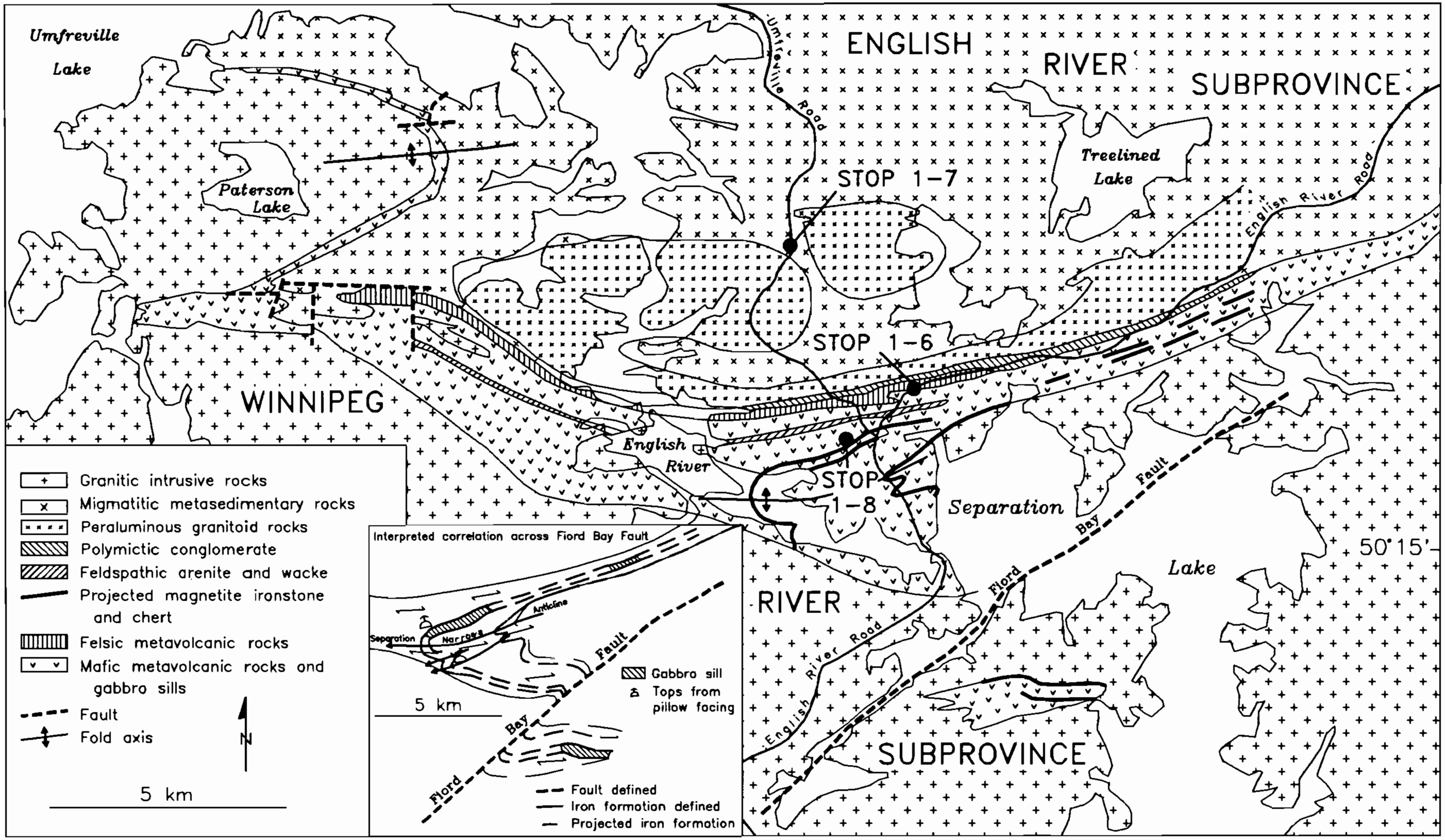


Figure 5. Geological map of the Separation Lake greenstone belt with location of stops 1-6 to 1-8. The inset map illustrates the inferred correlation of gabbro and ironstone units across the Fiord Bay fault.

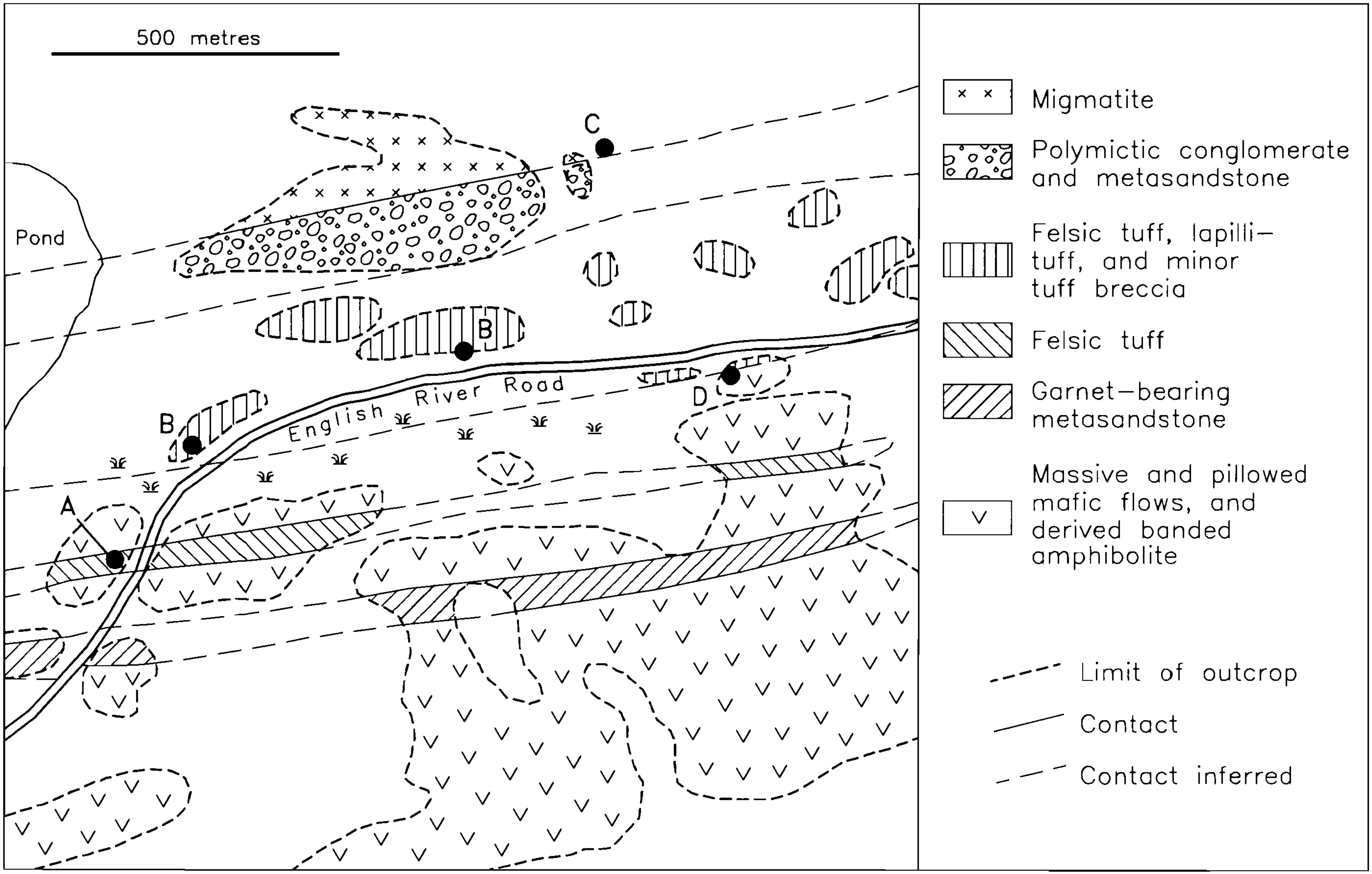


Figure 6. Sketch map illustrating the geology of stop 1-6.

subsequent to onset of migmatization. At the north side of the outcrop there is an abrupt transition to a granitic phase of the English River migmatites. Time permitting we will examine other outcrops close by to further inspect folding in the conglomerate/sandstone sequence and evidence of rotation of granitoid clasts.

Return south across the sequence to the road, and proceed to

Locality D - on the south side of the English River Road intensely sheared mafic metavolcanics and possibly also felsic volcanics are so strongly tectonized as to constitute a mylonite zone. Production of cataclastically-induced melting giving rise to pseudotachylite is seen here. This zone appears to be one of strongest deformation in the vicinity of the greenstone/migmatite boundary zone.

Return to the intersection of the English River Road and Umfreville Road, reset odometer to zero and proceed northwest along the Umfreville Road. At 0.35 km we pass a lumber haulage track on the west side of the road to which we will return. The next stop is in a cut-over area at 6.4 km.

Stop 1-7 English River metasedimentary gneiss and peraluminous granitoid rocks

Refer to Figure 5 for location.

A number of outcrops in a recently cut over area provide an opportunity to examine metasedimentary gneiss and associated peraluminous granitoid rocks. These rocks are typical of the metasedimentary gneiss assemblage of the English River subprovince although the proportions of peraluminous granitoid rocks can vary widely, as is seen on a smaller scale in this outcrop.

The metasedimentary component consists of plagioclase-quartz-biotite-garnet schists and gneisses that locally has a layered structure defined by alternating layers of finer grained, granoblastic gneiss and coarser grained, porphyroblastic gneiss. The latter is richer in biotite and garnet and has more abundant, diffuse pegmatitic material. Although not particularly well preserved at this locality, the alternation of the granoblastic and porphyroblastic gneiss is interpreted to represent original interbedding of wacke and more pelitic (siltstone/shale) beds, respectively.

The peraluminous granitoid rocks here include both widely distributed, thin, stromatitic, pegmatoid lit and a thicker, medium grained, equigranular to coarsely pegmatitic granitic sill. The lit are generally parallel to inferred primary layering but are locally discordant and parallel to the axial surface of minor folds. They are preferentially developed in the more pelitic layers sometimes contain septa or marginal selvages of biotite-garnet rich material. A distinctive feature of the thicker sill is the presence of abundant clots composed enriched in garnet, biotite and, locally, cordierite.
The peraluminous granitic material is interpreted to be the leucosome and the concentrations of biotite, garnet and cordierite in septa, selvages and clots the melanosome originating from the in situ (or nearly so), vapor-present partial melting of the metasedimentary gneiss.

Proceed back towards the English River road for approximately 6 km to the lumber haulage road. Proceed west on this road for 0.6 km where we will park for a walk into an exposure of rare element enriched pegmatite

Granitic pegmatites- *This field trip will present an opportunity to briefly examine one rare-element pegmatite, an example of a mineralization type whose distribution strongly correlates with those subprovince boundary zones that involve, in part, a metasedimentary-rich domain (Fyon et al. 1992, p.1118).*
Investigation of such mineral deposit types is relevant for several reasons:

- *They provide an example of the control of large scale tectonic processes on the localization of economic mineral deposits, e.g., Dryden Pegmatite Field (Breaks and Janes 1991) and Winnipeg River-Cat Lake Pegmatite Field (Cerny and Meintzer 1988).*
- *They exhibit a close association with highly evolved, peraluminous granitic plutons (e.g., Separation Rapids pluton) that were generated by intracrustal melting of wacke-mudstone rocks under high grade to granulite, Abukuma-type metamorphic conditions. These conditions are plausibly related to collisional orogeny.*
- *Highly fractionated rare-element pegmatites are specifically emplaced in low to medium grade metamorphic zones located down-isograd from the high grade centres marked by extensive partial melting of clastic metasedimentary protoliths.*

The Separation Rapids Pegmatite Group lies entirely within the Separation Lake metavolcanic belt (Figure 7) and represents the most easterly part of the Winnipeg River-Cat Lake Pegmatite Field

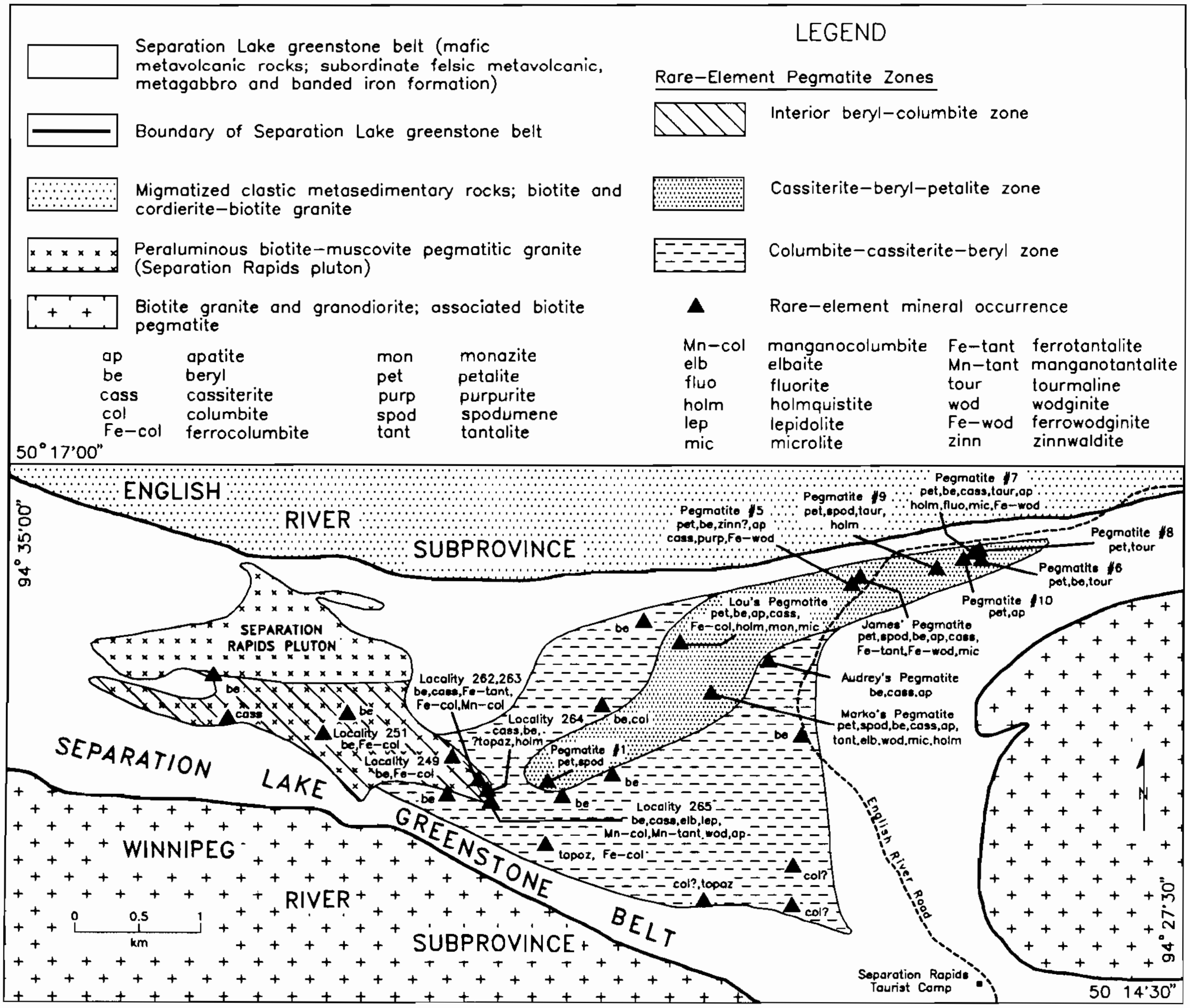


Figure 7. Rare-element pegmatite zones of Separation Rapids Group. Stop 1-8 is at Marko's pegmatite.

which was initially defined only in Manitoba (Cerny et al. 1981). The constituent rare-element pegmatites within this group were derived from the Separation Rapids pluton (Figure 7), a peraluminous granite mass which is dominated by various members of the pegmatitic granite facies of Cerny and Meintzer (1988). The parent granite also is notable in the sporadic occurrence of minerals that also are found in the external pegmatite zones: beryl, lithian tourmaline, ferrocolumbite, manganotantalite, wodginite, stibiomicrolite and cassiterite.

Stop 1-8: Rare-element pegmatite, petalite zone of Separation Rapids Pegmatite Group

Exposed on the ridge to the south is a 8 by 175 meter, white weathering, internally-zoned, petalite-bearing pegmatite dyke which represents the largest yet discovered in the petalite zone of the Separation Rapids Pegmatite Group. Recent stripping and cleansing of this pegmatite allows for detailed examination of the geological and mineralogical features of a pegmatite mass comparable in extent of magmatic fractionation to rare-element pegmatites of the Bernic Lake Pegmatite Group of Manitoba, which contains the famous Tanco Pegmatite.

Primary Pegmatite Units

The detailed geology of Marko's Pegmatite is given in Figure 8 where two primary zones compose most of the dyke:

1. Wall Zone
2. Petalite-Rich Core Zone.

Wall Zone

The Wall Zone comprises a 0.5-1 meter-thick, beryl-muscovite-albite-quartz assemblage which is absent of petalite and K-feldspar. This quartz-rich unit contains albite in coarse blocky habit which typically occurs in aggregates. In some parts of the zone, quartz is euhedral to subhedral and completely enveloped by albite, possibly indicative of early crystallization for the quartz. Accessory cassiterite, wodginite, green tourmaline, apatite and rare stibiomicrolite occur sporadically.

Petalite-Rich Core Zone

The wall zone lies in fairly sharp contact with a 0.5-6 meter-thick core zone that contains exceptional quantities of coarse, white petalite (80-95%). The amount of petalite is locally variable especially in the westernmost mass, where coarse blocky K-feldspar up to 2 meters across apparently crystallized prior to the petalite masses as it fills the interstices between the euhedral K-feldspar crystals. Scarce quartz masses occur in a similar interstitial manner. At the western end of the largest outcrop (50 meters length), the wall zone encloses the petalite-rich core zone and appears to represent the rarely observed apex of a rare-element pegmatite dyke

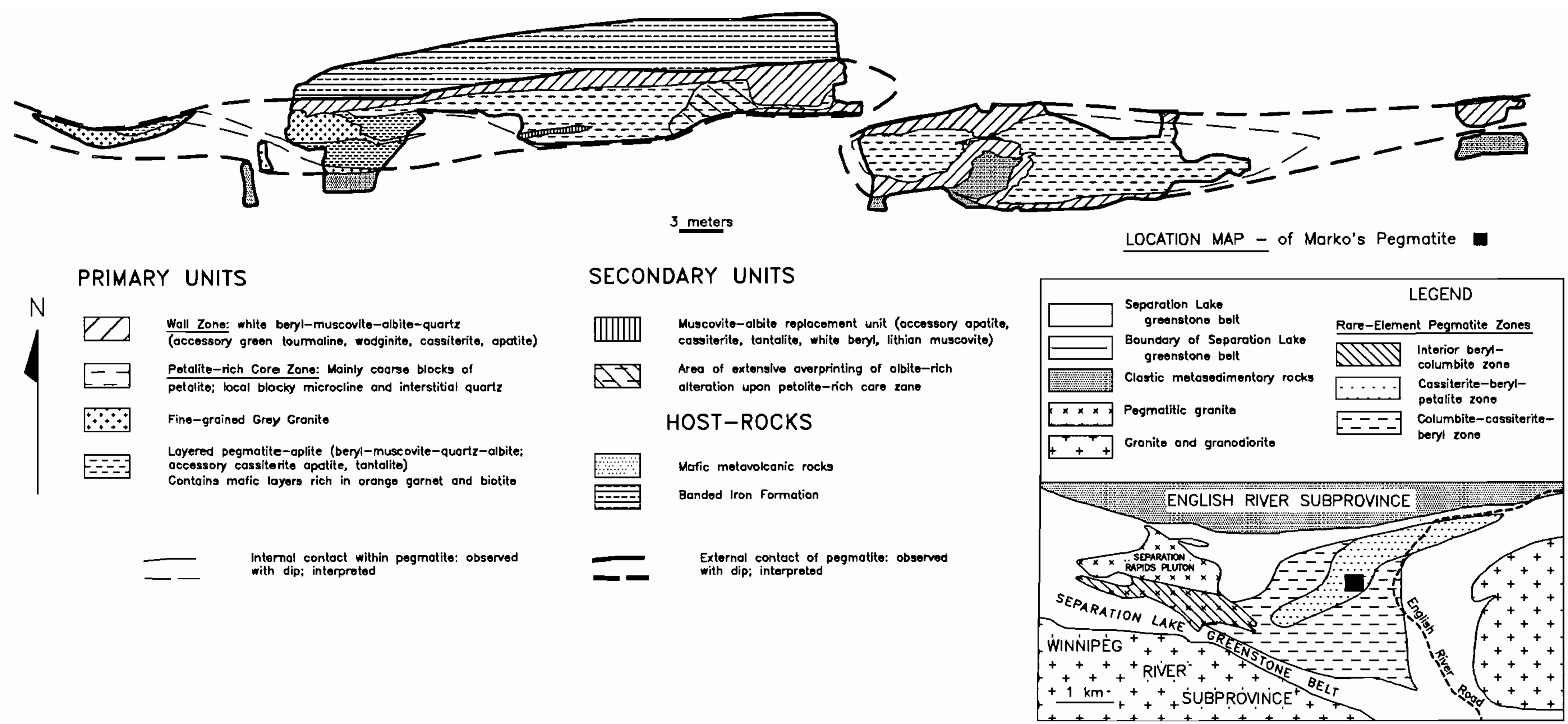


Figure 8. Detailed geology and internal zones in Marko's Pegmatite.

system.

Other Primary Units

At the eastern end of the same outcrop area, the petalite-rich core zone grades into a layered sequence composed of fine to medium grained, grey, muscovite granite and coarser, light pink, beryl-muscovite-quartz-albite rock. The latter unit contains curious, dark layers, up to 8 cm thick, which are richer in quartz (30%), orange garnet (60%), and biotite (10%) than its albite-rich host. Fine grained black oxide minerals (cassiterite, tantalite and wodginite) tend to be more common in the albite-rich unit or along the contact with the orange garnet-rich rock.

A short walk of 30 meters east partially through low thick bush reveals the end of the pegmatite at a point where it has narrowed to about 0.5 meter. Here intense metasomatic reaction with the banded iron formation host-rock is marked by the high concentration of biotite and coarse masses of garnet. The pegmatite here is dominated by albite with local masses of light yellow-brown, altered pollucite. This rare mineral is known only at three other localities in Ontario. Sporadic brown cassiterite up to 0.5 cm diameter also occurs here.

Secondary, Replacement Units

Late stage replacement processes have effected several mineralogical changes, most notably to the petalite-rich core zone. About 5-10% of this zone has been affected by:

1. two stages of albitization
2. breakdown of petalite into "microsqui" veinlets (Squi is an acronym for spodumene-quartz intergrowth)
3. green muscovite replacement of K-feldspar.

The first stage of albite-replacement involves development of orange low albite porphyroblasts and small veins (up to 5 cm thick by 1.5 meters length) in petalite. This alteration type is transected by veinlets of "microsqui" which comprise fine intergrowths of secondary spodumene+quartz that developed by thermal breakdown of petalite. This alteration is, in turn, overprinted by the second stage of albite replacement which is significant in containing concentrations of economically important oxide minerals (wodginite, tantalite, microlite). This alteration occurs as bulbous masses of albite (70-80%), lithian muscovite (10-20%) and lesser white beryl and apatite which crosscut the orange albite and microsqui veinlets. Accessory amounts of cassiterite, tantalite, microlite, wodginite, zircon and monazite occur in these replacement masses.

Return to the English River Road and proceed south on to highway 658 and continue south to Kenora.

Day 2 Tannis Lake - High Lake - Rush Bay Area

Today we will proceed west of Kenora and briefly examine another aspect of the Wabigoon - Winnipeg River boundary. The majority of the day will be devoted to examination of the Keewatin Series metavolcanic rocks as well as a late, unconformable sequence that overlies both the Keewatin Series and a porphyry intrusion.

Proceed west from Kenora on Highway 17 to the Gundy Lake Road (several kilometres east of the provincial border). Reset odometer to zero and proceed north on the Gundy Lake Road

0.0 km Gundy Lake road and highway 17 junction

3.1 km fork in road- turn left

4.2 km junction of Gundy Lake Road and Piute Lake Road. Turn left on to the Piute Lake Road and proceed a further 1.25 kilometres and take the track to the left and park vehicles on the large outcrop overlooking Piute Lake.

Winnipeg River - Wabigoon subprovinces relationship- *The Winnipeg River and Wabigoon subprovinces represent discrete crust-forming events with the former separated, directly or indirectly, from the mantle between 2830 and 3170 Ma and being inactive during the interval 2710 to 2775 Ma when the Wabigoon crust developed (Beakhouse and McNutt, 1991). The recognition of these two adjacent crustal segments of different age with no clear shared geological history prior to late granitic plutonism (see discussion above regarding the marginal granodiorite) is permissive of the contact between the two being either an unconformity or a fault. Proponents of the former interpretation point to the absence of tonalitic dikes in the adjacent Wabigoon volcanic sequence that faces away from the contact, minor metasedimentary units along the inferred unconformity surface and mafic dikes cutting the inferred basement rocks (Clark et al., 1981). Others point to evidence for ensimatic development of the Wabigoon subprovince, paucity of 2710-2775 Ma activity in Winnipeg River subprovince and timing of tectonic thickening and regional high-grade metamorphism and advocate tectonic juxtaposition of two terranes by collisional orogeny (Davis et al., 1988; Beakhouse and McNutt, 1991).*

Critical in this regard is the age and significance of mafic dikes that cut the marginal portions of the Winnipeg River subprovince in several locations (Tannis Lake area, Sioux Lookout area) adjacent to the Wabigoon subprovince. These mafic dikes exhibit broad textural (equigranular and subordinate plagioclase-phyric varieties) and compositional (sub-alkalic, tholeiitic basalt) similarities to many of the mafic metavolcanic sequences in the Wabigoon but these characteristics are not uncommon in Archean mafic volcanic sequences and the similarities remain equivocal. In the absence of additional work, the age and significance of these mafic dikes remain unclear. It is possible that they represent feeder dikes to Wabigoon volcanism, but also could be significantly older and related to either volcanism in younger Winnipeg River gneissic assemblages (e.g., 2830-2880 Ma. Kenora gneissic assemblage) or rifting of a more extensive pre-Wabigoon sialic crust.

Stop 2-1 Tannis Lake Tonalite cut by Mafic Dikes/ Contact with Wabigoon Subprovince at Piute Lake

Refer to Figure 9 for location.

A series of outcrops to the south of Piute Lake (Figure 10) expose highly deformed mafic metavolcanic rocks along with minor cherty and ferruginous interflow metasedimentary rocks of the Wabigoon subprovince and tonalite together with crosscutting mafic dikes of the Winnipeg River subprovince. The contact itself is not exposed but occurs within a narrow unexposed interval marked here, as well as for at least several kilometres to the east, by the presence of several chemical metasedimentary units in the Wabigoon.

The Wabigoon metavolcanic rocks here are correlated with the Lower Keewatin Supergroup of the northern Lake of the Woods region (Figure 11) for which a U-Pb age of 2738 Ma has been obtained (Table 1). The mafic metavolcanic rocks exposed here are highly strained and, although some of the compositional banding observed locally may represent transposed pillow selvage material, little can be determined concerning the primary depositional characteristics of these rocks.

The tonalite here, and throughout much of the Tannis Lake area, is strongly lineated but only weakly gneissic, contrasting with the Kenora gneissic suite to the east. A U-Pb investigation carried out in similar rocks several kilometres northeast of this location indicates that the tonalite was emplaced at 3051 Ma and the mafic dikes were metamorphosed at 2701 Ma (Davis et al., 1988).

Return to Highway 17 and proceed west for 3.75 km. Proceed to the east end of a long outcrop on the north side of Highway 17.

Lake of the Woods greenstone belt- *Regional strain intensity is generally greatest along the margins of the Western Wabigoon subprovince (Blackburn et al. 1991). In the Lake of the Woods greenstone belt, located along the northern margin of the Wabigoon subprovince adjacent to the Winnipeg River subprovince, deformation is manifest as 3 or more episodes of regional folding and faulting. These deformation events have greatly complicated stratigraphic interpretation by repetition of stratigraphic units and it has only been since the early 1990s that a stratigraphic model has been erected based on systematic detailed and regional mapping, structural studies, and precise U-Pb geochronology. Table 1 illustrates the main components of the stratigraphy, lithologies, deposition environments and U-Pb ages. Briefly, the stratigraphic model consists of lowermost tholeiitic mafic volcanic groups comprising the Lower Keewatin supergroup, the upper part of which*

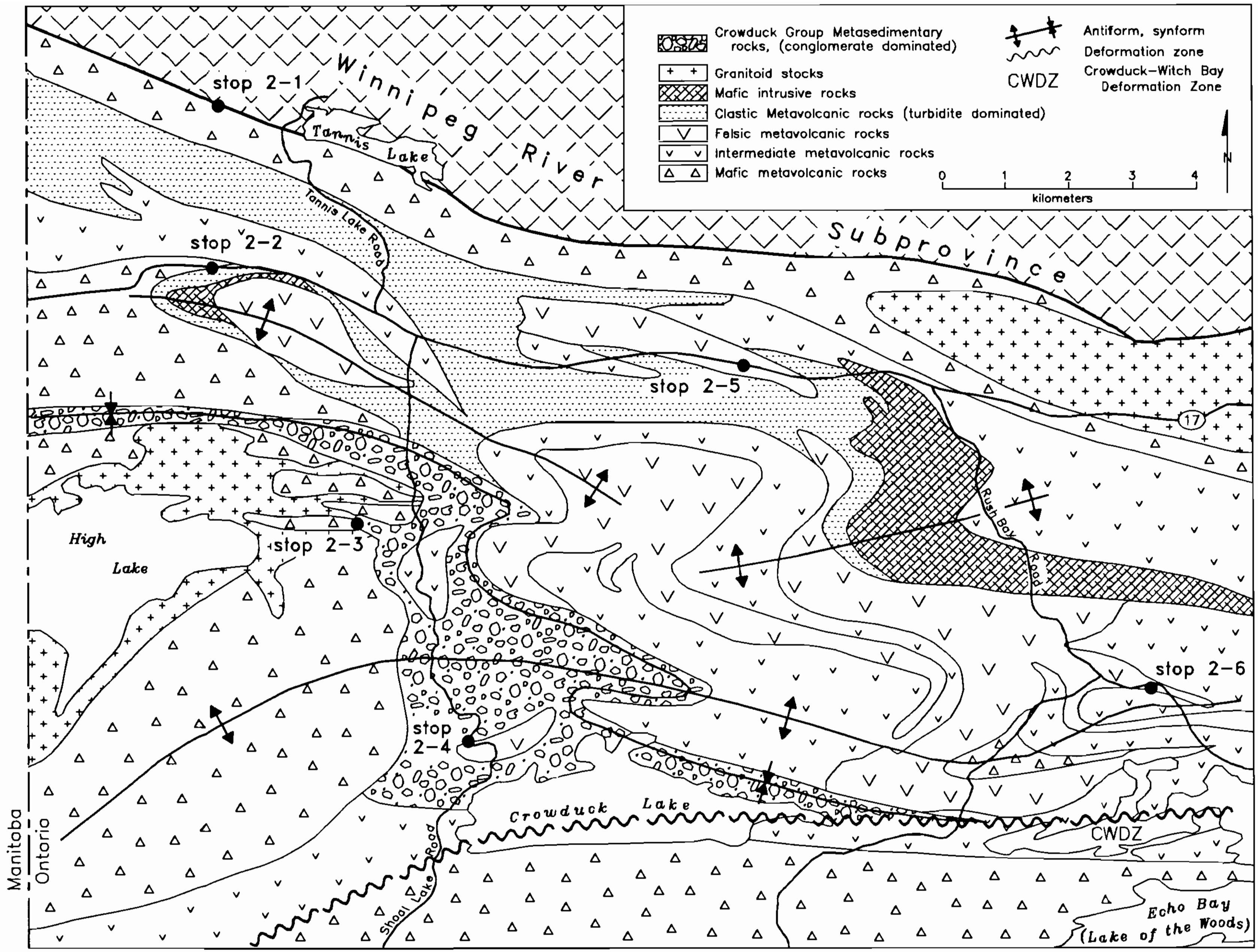


Figure 9. Geological map of the High Lake - Rush Bay area (modified after Davies, 1965) illustrating the location of day 2 stops.

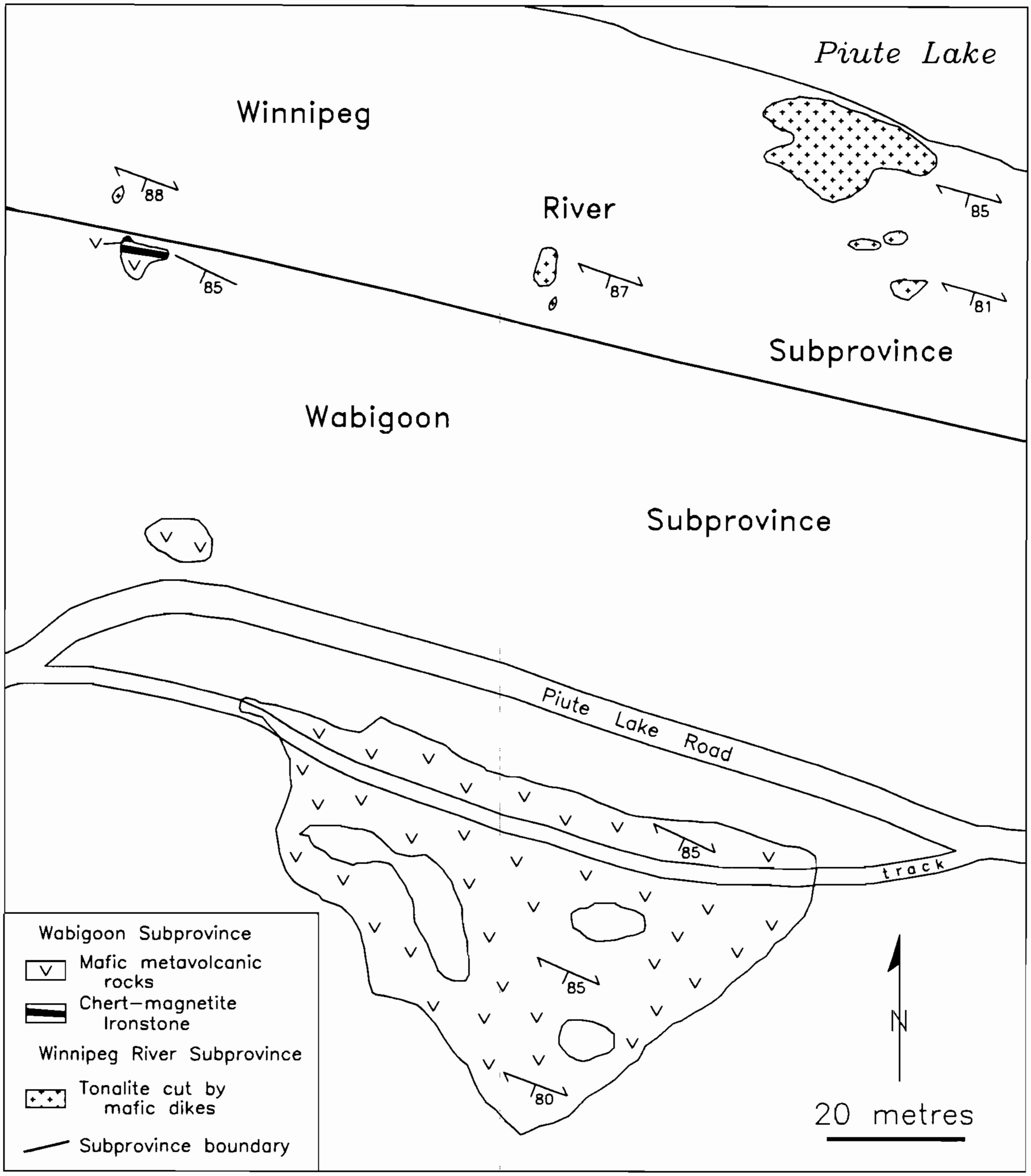


Figure 10. Generalized outcrop map of the Winnipeg River - Wabigoon contact area at Piute Lake (Stop 2-1).

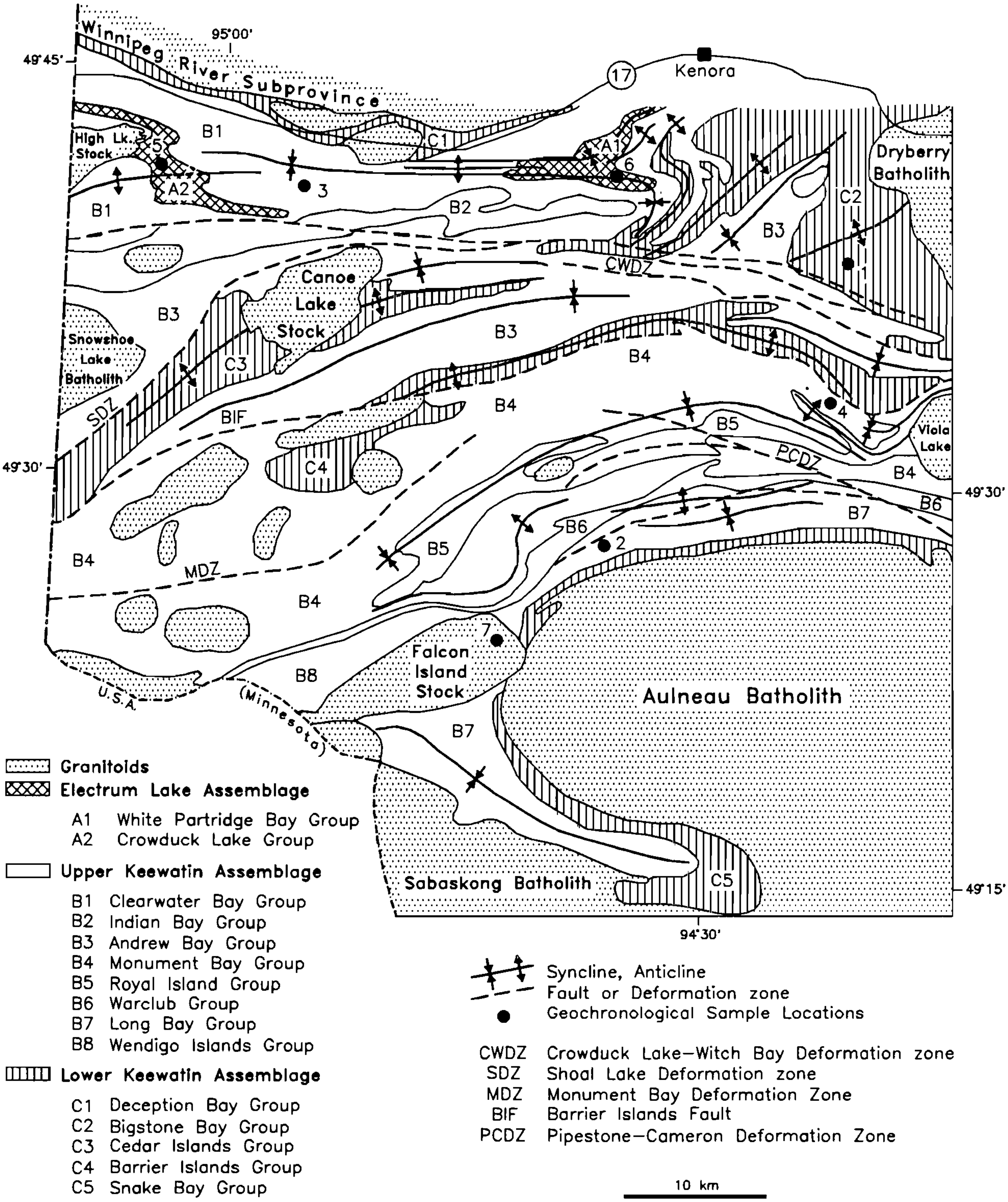


Figure 11. Stratigraphy of the Lake of the Woods area.

Table I. Lake of the Woods Greenstone Belt Lithostratigraphic Supergroups and Groups (Geochronology by D. Davis see [1]Blackburn et al. 1991, [2]Ayer, in press).

Stratigraphic Unit	Dominant Lithologies	Subordinate Lithologies	Deposition Environment	U-Pb age (Ma)
ELECTRUM LAKE SUPERGROUP				
Crowduck Lake Group	Conglomerates, arenites, wackes, siltstones.	Calc-alkaline mafic to felsic flows and pyroclastics.	Proximal subaerial to submarine.	<2699[2]
White Partridge Bay Group	Turbidites, conglomerates, mudstones, siltstones.	Felsic pyroclastics.	Distal to proximal submarine.	<2709[2]
UPPER KEEWATIN SUPERGROUP				
Clearwater Bay Group	Calc-alkaline mafic to felsic flows & pyroclastics.	Clastic & chemical sediments, tholeiitic mafic flows.	Shallow submarine to subaerial.	2719[2]
Indian Bay Group	Tholeiitic mafic flows	Calc-alkaline felsic volcanics.	Shallow submarine	
Andrew Bay Group	Calc-alkaline mafic to felsic volcanics.	Clastic & chemical sediments & tholeiitic mafic flows.	Shallow submarine	2723[1]
Monument Bay Group	Calc-alkaline mafic to felsic volcanics.	Tholeiitic to komatiitic flows, conglomerates, turbidites, & chemical sediments.	Shallow submarine to subaerial	
Royal Island Group	Turbidites.	Conglomerates & mafic to felsic volcanics.	Distal to proximal submarine.	
Warclub Group	Turbidites.		Distal to proximal submarine.	2715[1]
Long Bay Group	Calc-alkaline mafic to felsic volcanics.	Tholeiitic to komatiitic flows & clastic sediments.	Shallow submarine	2719[2]
Windigo Islands Group	Calc-alkaline mafic to felsic volcanics.	Clastic and chemical sediments.	Shallow submarine.	
LOWER KEEWATIN SUPERGROUP				
Deception Bay Group	Tholeiitic mafic flows.		Deep submarine.	
Bigstone Bay Group	Tholeiitic mafic flows.		Deep submarine.	2738[2]
Cedar Island Group	Tholeiitic mafic flows.	Komatiitic flows & calc-alkaline felsic volcanics.	Deep submarine	
Barrier Islands Group	Tholeiitic mafic flows.	Felsic volcanics.	Deep submarine	
Snake Bay Group	Tholeiitic mafic flows.		Deep submarine	>2732[1]

has an age of 2738 Ma. The Upper Keewatin supergroup contains highly diverse calc-alkaline and tholeiitic to komatiitic volcanic rocks with ages which range from 2714 to 2723 Ma. Sedimentary units are a minor part of the Upper Keewatin supergroup and are typically turbidites intercalated within the volcanic rocks. Uppermost in the stratigraphy is the Electrum supergroup consisting of clastic sedimentary rocks and minor volcanic rocks with detrital ages indicating deposition later than about 2700 Ma.

Stop 2-2: Clearwater group lithologies and contact relationships within the Upper Keewatin supergroup.

Refer to Figure 9 for location.

Normally graded wacke to siltstone range from 10 cm to 3 m in thickness. Bedding strikes about $170^0/75^0$ with tops to the west. Flame structures and detached ball and pillow structures can be recognized within the silty tops of beds. Light gray and dark gray spots occur in the upper silty portions of graded beds and may be either concretions or reduction spots. A predeformational origin is indicated by their flattening parallel to the cleavage. The high angle of foliation to bedding indicates proximity to the nose of the Baubee Lake anticline (Figure 9).

Walk 200 m west to the next outcrop on the north side of Highway 17.

The stratigraphic position of the tholeiitic mafic volcanic rocks of the High Lake formation is enigmatic in that previous interpretations have placed this unit within the Lower Keewatin supergroup (Davies 1965, Smith et al. 1988). The revised interpretation positions these mafic volcanic rocks within the Upper Keewatin supergroup. Synoptic and geochemical investigations indicate there are at least 5 different formations of tholeiitic to komatiitic composition within the Upper Keewatin supergroup and these rocks are generally compositionally more primitive than the tholeiites of the Lower Keewatin supergroup (Figures 12 and 13). This outcrop may provide evidence that the mafic volcanic rocks of the High Lake formation overlie calc-alkaline volcanic and sedimentary rocks of the Clearwater group (Table 1, Figure 11). The rocks of this outcrop have experienced considerable strain with a foliation roughly parallel to the contacts and an alternate interpretation might be that the sediment-volcanic contact marks a fault along which underlying mafic volcanics of the Lower Keewatin supergroup have been thrust over younger sedimentary rocks. However, the generally more primitive geochemical nature of the metabasalts of the High Lake formation and the interdigitation of volcanics and sediments at the contact favour the former hypothesis.

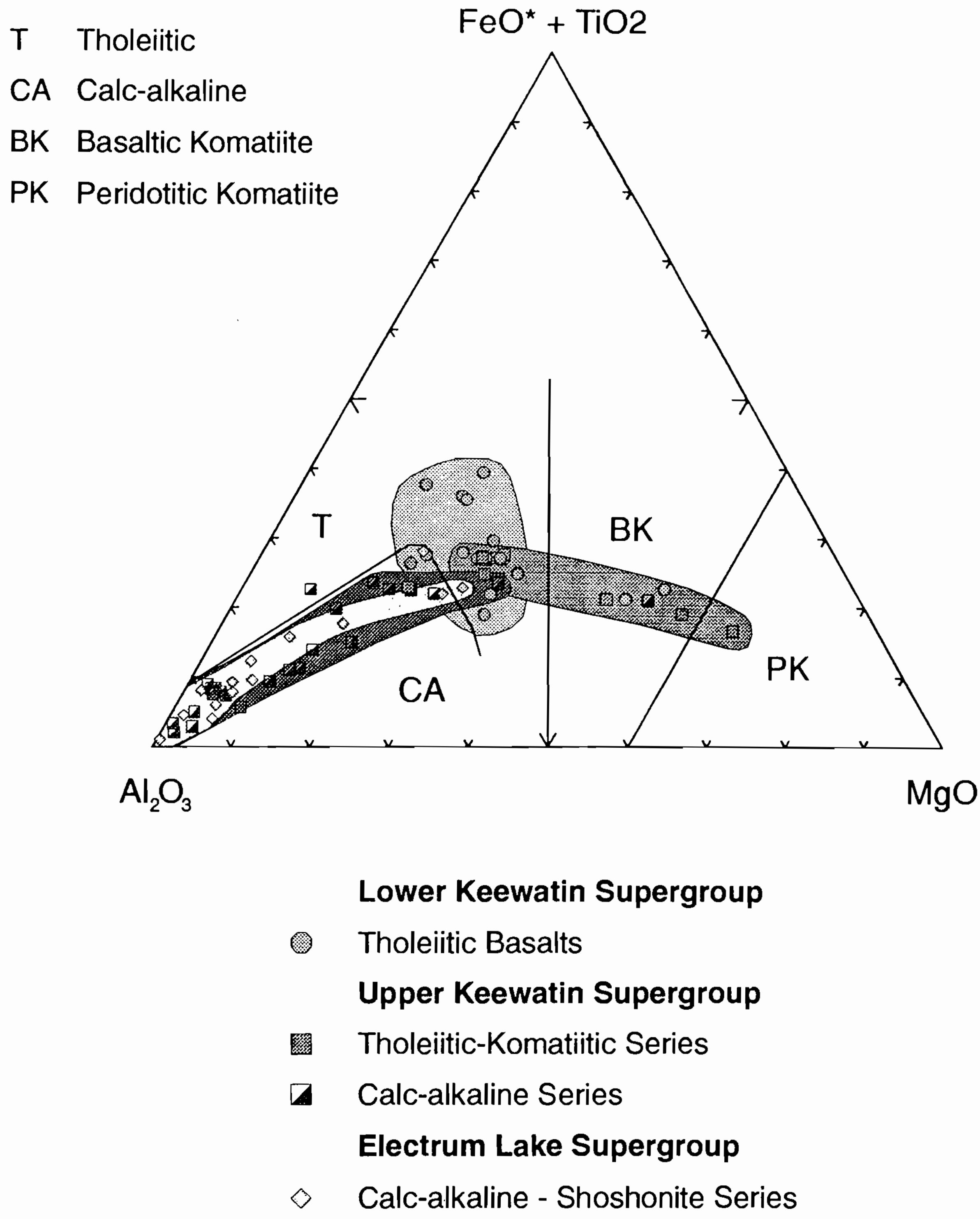


Figure 12. Major element geochemistry of Lake of the Woods area volcanic rocks.

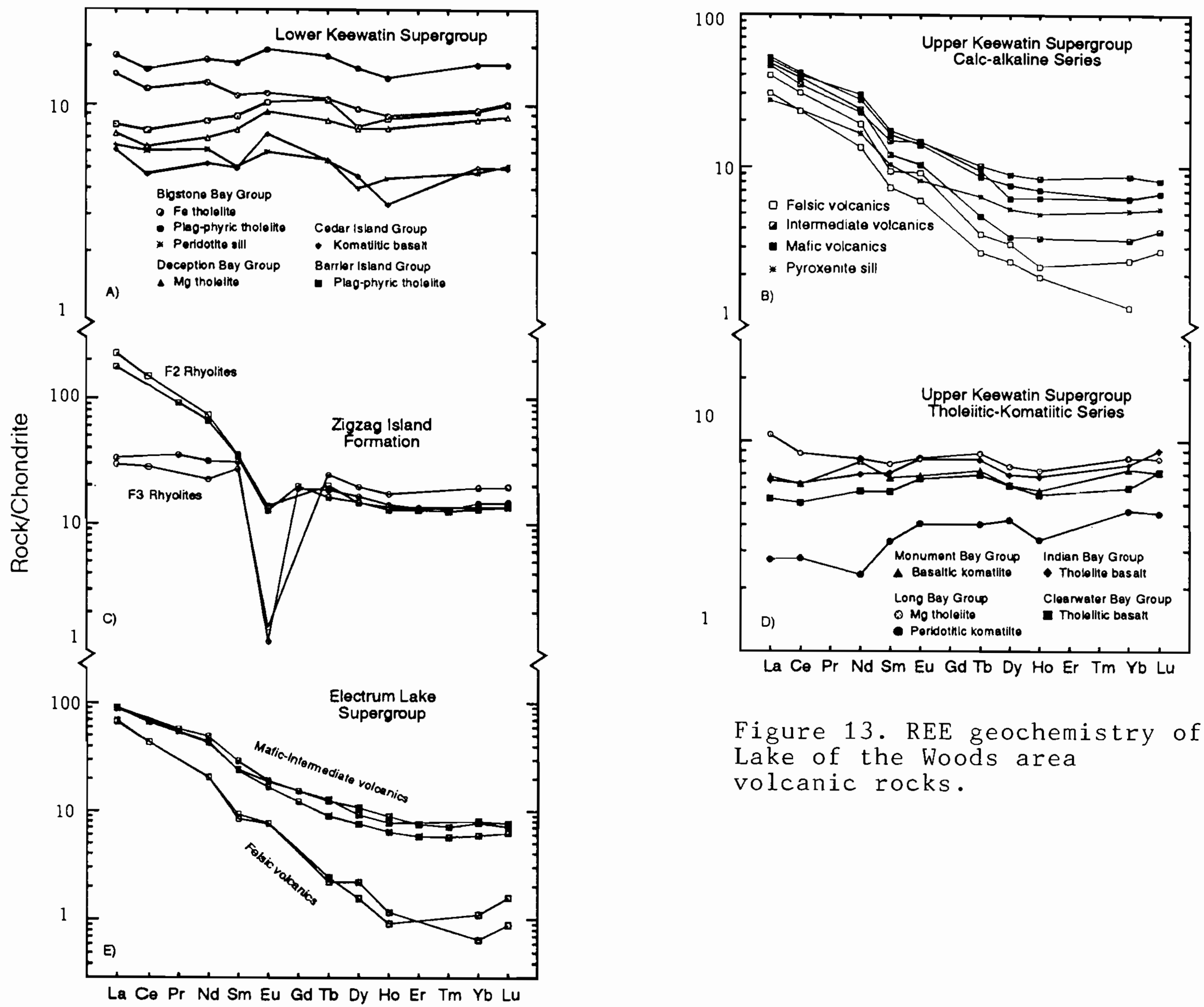


Figure 13. REE geochemistry of Lake of the Woods area volcanic rocks.

On the east side of the outcrop, thickly bedded wacke grades westerly into thinly bedded siltstones that have abundant Z-folds. In the central part of the outcrop the contact with the overlying mafic volcanics is interdigitated with thinly bedded rusty siltstone and plagioclase-phyric mafic flows. This contact trends southeasterly. Mafic flows along the contact range from fine-grained plagioclase-phyric varieties with subhedral phenocrysts up to 1 mm in size to medium-grained varieties. At the western end of the outcrop the mafic flows are fine-grained, aphyric varieties with possible pillow selvages.

Retrace our route eastward on highway 17 for 4.5 km to the Shoal Lake road. Proceed south on the Shoal Lake road for 2.9 km and turn west (right) on to a narrow road and proceed a further 0.8 km. The stop is located approximately 30 metres south of the trail at this point.

Late Unconformable Sedimentary Sequences- *Metasedimentary-dominated supracrustal sequences, often referred to as Temiskaming-type, that unconformably overlie deformed volcanic rocks and some plutons and were themselves subsequently deformed and metamorphosed are a minor, but significant, component of Archean greenstone belts in the Superior Province. Some of these sequences contain a component of alkalic or shoshonitic volcanism (see discussion later in guidebook). These sequences are widely distributed but tend to occur in proximity to major tectonic boundaries. Indirect age constraints based on the age of unconformably overlain rocks and detrital zircons in metasedimentary rocks for several such sequences (Ament Bay Group, Crowduck Group) located near the Wabigoon - Winnipeg River interface indicate that these sequences were deposited subsequent to 2709 Ma (Davis et. al., 1988; Davis and Smith, 1991). These sequences are interpreted to be the sedimentary signature of collisional orogeny related to compressional uplift related to collision and/or the development of transtensional basins.*
Two such units (the Crowduck and White Partridge groups) are situated in the northern part of the LWGB and have been assigned to the Electrum supergroup. Detrital zircons indicate minimum depositional ages of 2700 Ma, but also ages which range from 2785 to 2798 Ma, considerably older than any recognized igneous activity in the Lake of the Woods greenstone belt. These detrital zircons may have come from the Winnipeg River terrane but are somewhat problematical as there are no known equivalent Winnipeg River subprovince plutonic ages, only ones which are either older (i.e. 2830-3170 Ma) or younger (2665-2709 Ma).

Stop 2-3 Keewatin Series -Crowduck Group Unconformity at High Lake
Refer to Figure 9 for location.

In this outcrop, metabasalts of the Keewatin series are cut by porphyry dikes related to the High Lake stock and both are unconformably overlain by metasedimentary rocks of the Crowduck Group. Pillowed and massive metabasalts and the porphyry dikes are exposed on the western edge of the outcrop and the Crowduck Group faces towards the east (Figure 14).

In this outcrop, the Crowduck Group is divisible into 4 units that reflect a general coarsening upwards trend although stratigraphic relationships are complex because many units were eroded prior to, or concomitant with, deposition of overlying units. These scouring relationships are well exposed in the western portion of the outcrop. Here, as elsewhere along the unconformity, the local provenance is reflected in a close correspondence between clast population and immediately underlying lithologies. In this outcrop, clasts of the porphyry are conspicuously larger (up to 1 metre² in cross-sectional area) and predominate over other clasts which include relatively abundant mafic to felsic volcanic clasts and minor chert, ironstone, massive sulphide and vein quartz clasts. Other noteworthy features of the Crowduck Lake Group are well developed soft-sediment structures (slump folds, load casts and flame structures) in the fine grained units at the base of the sequence. A regolith is developed beneath the Crowduck Group and is most apparent in the porphyry unit.

It is not clear how much of the irregular nature of the unconformity can be attributed to pre-Crowduck topography. The large clast sizes in the conglomeratic units requires at least local steep gradients to have been present. A possible example of a syn-depositional growth fault rooted in a porphyry-basalt contact occurs in the southwest portion of the outcrop (Figure 14). Here, thinly bedded basal Crowduck Group sediments overlie the porphyry and butt up against the metabasalt whereas similar, stratigraphically higher sediments drape both underlying units. The geometry of the unconformity here also mimics the pattern displayed by regional folding in this area (Davies, 1965). The orientation of the foliation in this outcrop is approximately parallel to the axial surface of these regional folds and is markedly discordant to the unconformity and bedding in the Crowduck Group. Consequently, at least some of the irregularity of the unconformity surface may be attributable to post-Crowduck deformation.

Volcanic rocks are relatively rare in the Electrum supergroup. In the Crowduck group they consist of rhyolite flows in a lens-shaped dome about 2 km long by 500 m thick near the southern stratigraphic base of the predominantly conglomeratic Crowduck group. Clinopyroxene-phyric, mafic to intermediate, massive, amygdaloidal flows and pyroclastics occur in a similar stratigraphic position within the Crowduck group about 3 kilometres north of stop 2-4. Although volcanic rocks of the Electrum supergroup fall within a similar range to the calc-alkaline series of the Upper

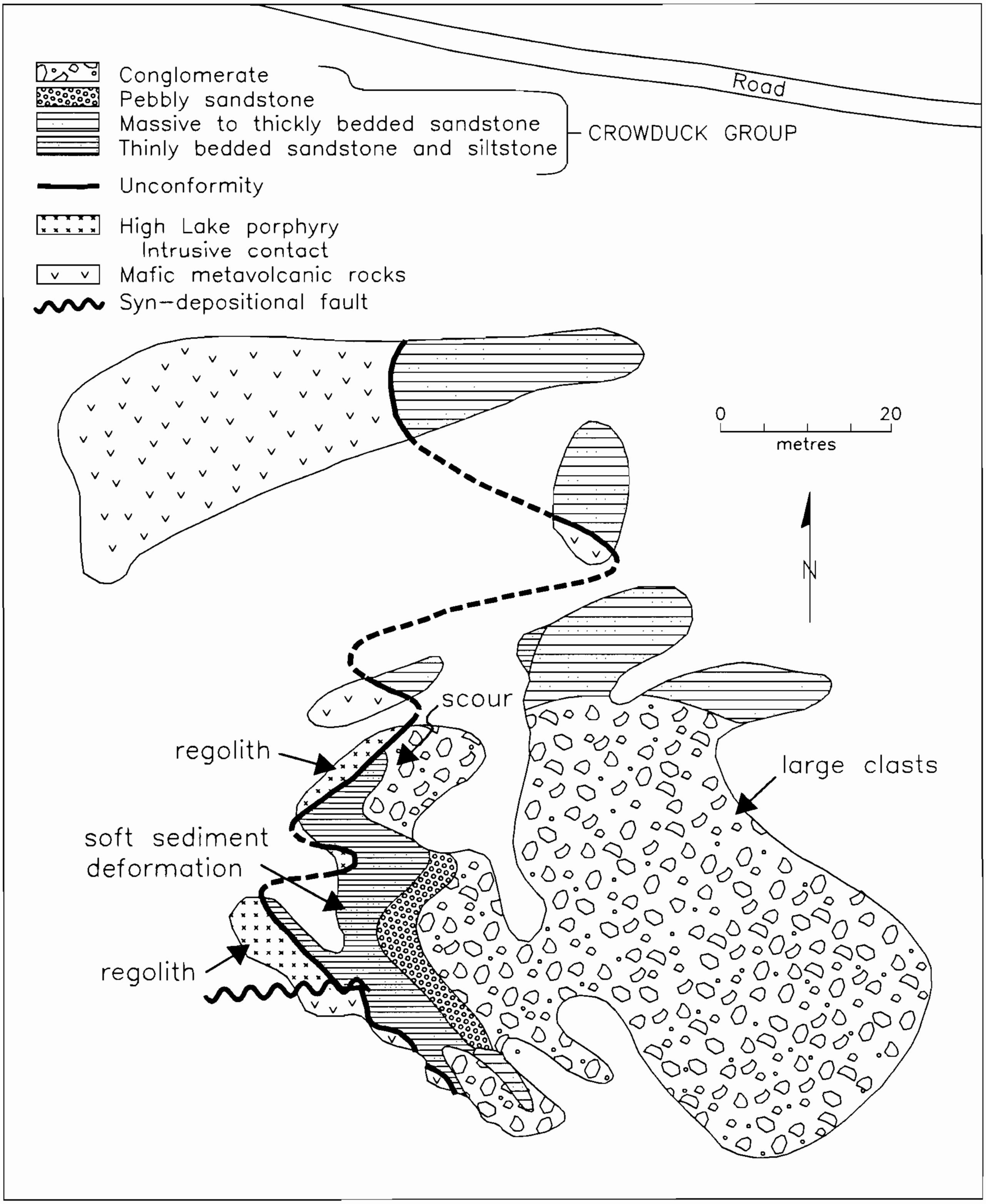


Figure 14. Generalized outcrop map of the Crowduck Group unconformity near High Lake (Stop 2-3).

Keewatin supergroup on figure 12. On the basis of chondrite-normalized REE plots they are more evolved with higher absolute REE values in mafic and intermediate compositions than in the felsic compositions (Figure 13) and are interpreted to be of shoshonitic affinity.

Proceed back to the Shoal Lake Road, turn right and proceed south for 7.8 km to a small flat outcrop on the east side of the road.

Stop 2-4: A felsic volcanic dome within the Crowduck group

Refer to Figure 9 for location.

At this stop the felsic metavolcanic rock is brecciated with angular, light-green coloured aphyric fragments in a similar matrix. Many of the clasts have highly contorted flow laminations defined by zones of coalescing devitrification spherulites. The brecciation appears to be restricted to the upper, northern portion of the dome. Underlying these brecciated flows the felsic volcanics are massive to flow laminated but unbrecciated (visible in the outcrop 100 m to the south). These features suggest the felsic unit represents an exogenous felsic dome with an upper brecciated carapace.

Return to Highway 17, turn right and proceed for 8.5 km to a long outcrop on the north side of Highway 17 near the east end of Moth Lake.

Stop 2-5: Felsic debris flows and pyroclastic flows within the Upper Keewatin supergroup.

Refer to Figure 9 for location.

A heterolithic debris flow bed with an 8 cm reversely graded zone along the basal contact occurs near the east end of outcrop. Clasts are moderately sorted, subangular to angular and matrix supported. Clasts vary from fine-grained and aphyric to feldspar phyric intermediate volcanics. The foliation is at a high angle to bedding and this has resulted in scalloped contacts between beds. Underlying the debris flow is a 6 m thick pyroclastic flow containing angular felsic quartz-feldspar porphyry clasts and subround mafic clasts in an aphanitic matrix. Clasts range up to 10 cm in size.

This pyroclastic flow unit is underlain by bedded tuffs and thickly bedded pyroclastic flows. In the central part of the outcrop variation in matrix content

and grading define several debris flow units capped by a rusty, tectonized, argillaceous horizon. This is overlain by quartz-feldspar lithic lapilli-tuff pyroclastic flow.
The west end of outcrop comprises fine-grained thinly to thickly laminated tuff overlain by more lapilli tuff.

Return to vehicles and continue driving 3.0 km east to the Rush Bay road. Turn right and proceed south for 6.1 km to the junction with the Clytie Bay road. Keep to the left fork and proceed a further 0.95 km to a large outcrop on the right side of the Rush Bay road.

Geochemistry of Lake of the Woods Felsic Metavolcanic Rocks- *Geochemically unusual felsic volcanic rocks occurs in the Zigzag formation within the Clearwater group. The formation is about 50 km long and 0.1-1 km thick. The felsic volcanics are dominantly pyroclastic and locally welded suggesting shallow submarine or subaerial eruption. The formation consists of high silica F3 type rhyolites which have elevated and unfractionated HREE and significant Eu depletion anomalies (Figure 13), in distinct contrast to the highly fractionated REE patterns of the typical calc-alkaline rhyolites (Figure 13) of the Upper Keewatin supergroup. F3 rhyolites are an important mineral exploration target in that they are often found closely associated with volcanogenic massive sulphide deposits. The genetic nature of this correlation is as yet uncertain, but speculation is that they are a product of partial melting of underlying basaltic crust (e.g. The Lower Keewatin supergroup) at shallow depths and are thus indicative of localized upper crustal extension and high heat flow (Lesher et al. 1986, Barrie et al. 1993).*

Stop 2-6: Zigzag formation, a proximal volcanic environment consisting of felsic flows and coarse debris flows.

Refer to Figure 9 for location.

The southern part of the outcrop comprises a massive, quartz-feldspar porphyry with subhedral to euhedral phenocrysts up to 3-4 mm in a very fine-grained matrix. The contact between the porphyry and the overlying debris flow is rubbly with clasts of porphyry decreasing away from the contact. On the east side of the outcrop lateral to the porphyry is a train of porphyry clasts. The porphyry has been interpreted to be a lava flow overriding its own autoclastic breccia. Immediately below a megaclast described later, is a series of podiform quartz feldspar porphyries from 1 to several metres thick found along the same stratigraphic horizon. There may be several interpretations for this feature, some of which are; podiform flows, erosion of the same flow, or tongues of the same flow.

The debris flows are poorly sorted and vary in thickness with local evidence for normal grading of clasts within a matrix of ash size crystals and rock fragments. The subangular to angular clasts are up to several metres in size. The most noticeable and predominant clast type is a mottled texture felsic volcanic with quartz and feldspar phenocrysts in an aphanitic matrix. Many of these clasts exhibit flow banding that has been highlighted by devitrification spherulites. Of note is an extremely large clast (podiform flow?) of similar material about 5 m x 3 m in size resting on the top of a quartz feldspar porphyry flow. Minor clasts within the debris flows are black graphitic slate and dark green vesicular mafic volcanic. Within the thick massive debris flows lenticular beds of cut and fill channel deposits are up to 7 m by 1 m. They consist of normally graded, heterolithic, lapilli-tuff to tuff. These lenses are interpreted to be channel infillings within the thicker debris flows.

A tectonic fabric is defined by the flattening of clasts in a ratio of 2:1 striking $080^0/90^0$. There are small sinistral fault offsets along the same strike and could be responsible for the interdigitation of coarse and fine debris flows.

Return to highway 17 and proceed east to Kenora. We will continue east on highway 17 through Vermilion Bay to the Dryden area. En route to Vermilion Bay, we will pass exposures of the Dryberry Batholith and the Vermilion Bay greenstone belt (Figures 2 and 3). East of Vermilion Bay we pass into an area underlain by metasedimentary and granitoid rocks such as will be examined on Day 4.

Day 3 Manitou Lakes Area

The third day of the trip will provide an opportunity to examine geological relationships typical of the central portion of the Western Wabigoon subprovince.

From in Dryden follow the road signs to Highway 594. At the traffic lights just before crossing the Wabigoon River, reset odometer, and continue 6.6 km west along Highway 594 to the turn off onto Highway 502, the Manitou Highway. Turn south (left) onto Highway 502. Reset odometer. All distances south along the Manitou Highway will be from this point, for a distance of 147 km.

Outcrops at the side of the road at this intersection are pelites and wackes of the Warclub group, metamorphosed to low grade (greenschist facies). A low area of no outcrop marks the trace of the Wabigoon fault. Another 0.3 km on are mafic metavolcanics of the Upper Wabigoon group: we have crossed out of the Sioux Lookout domain into the Atikwa-Manitou volcano plutonic domain. Proximity to the fault is indicated in the metavolcanics at this outcrop: they are severely

deformed, but remnant pillow structures can be seen.
Continuing on, the highway traverses a north-facing section through, in descending order, Upper Wabigoon volcanics (mafic), Lower Wabigoon volcanics (mixed mafic to felsic, mostly flows) and Eagle Lake volcanics (mafic), and then into granitic rocks of the Atikwa batholith after about 10 km.

The Atikwa - Manitou Volcanoplutonic Domain - *Relationships between volcanic and plutonic rocks centred on and including the Atikwa batholith are typical of those occurring in other volcanoplutonic centres (Aulneau, Sabaskong) in the western Wabigoon subprovince. Extensive geochronological investigations suggest that many of the metavolcanic rocks together with both intermediate and mafic plutonic rocks are coeval, and perhaps comagmatic (Davis and Edwards, 1985, 1986; Davis e⁴ al., 1982). The plutons are envisaged to be the subvolcanic magma chambers that have risen into their own volcanic ejecta.*
The Aulneau, Sabaskong and Atikwa batholiths, that are interpreted to represent the plutonic root to much of the volcanism in the western Wabigoon subprovince, are dominated by rocks within the compositional spectrum tonalite - quartz diorite - granodiorite. The absence of inherited zircons (Davis et al., 1988) and equivocal nature of field evidence for an unconformable relationship with potential underlying sialic crust, suggest development in an ensimatic regime. These rocks have distinctive geochemical characteristics including low K and Rb and high Ca, Sr and Na/K, moderately fractionated REE with HREE at 1-4x chondrite, LREE at 30-60x chondrite and negligible Eu anomalies and mantle-type radiogenic (Sr and Nd) isotopic signatures. These characteristics have lead to the interpretation that many of these rocks were derived from the partial melting of tholeiitic basalt at mantle or lower crustal depths (Davis and Edwards, 1985; Beakhouse and McNutt, 1991). The extent to which a component derived from direct melting of LILE-enriched mantle may be present in these plutons remains a question for further study. These rocks, chemically high-Mg andesites and originally referred to as sanukitoid, have many geochemically overlapping characteristics with basalt-derived tonalite but differ importantly in having high-Mg as well as Ni and Cr (e.g., Stern et al., 1989; Sutcliffe et al., 1990). An important difficulty in distinguishing these two origins, however, arises from the probability that even limited fractionation will obscure some of the diagnostic characteristics of primitive sanukitoid magmas.

23.7 km south on highway from intersection of 594 and 502 (near end of a long straight stretch and just before a right hand curve in the road).

Stop 3-1 Atikwa batholith

Refer to Figure 15 for location.

This outcrop is in the Dore Lake lobe of the Atikwa batholith. The main rock type in the outcrop is a medium grained, massive to weakly foliated, relatively homogeneous tonalite containing widespread, minor mafic enclaves. The tonalite is cut by several finer grained, equigranular to porphyritic dikes ranging in

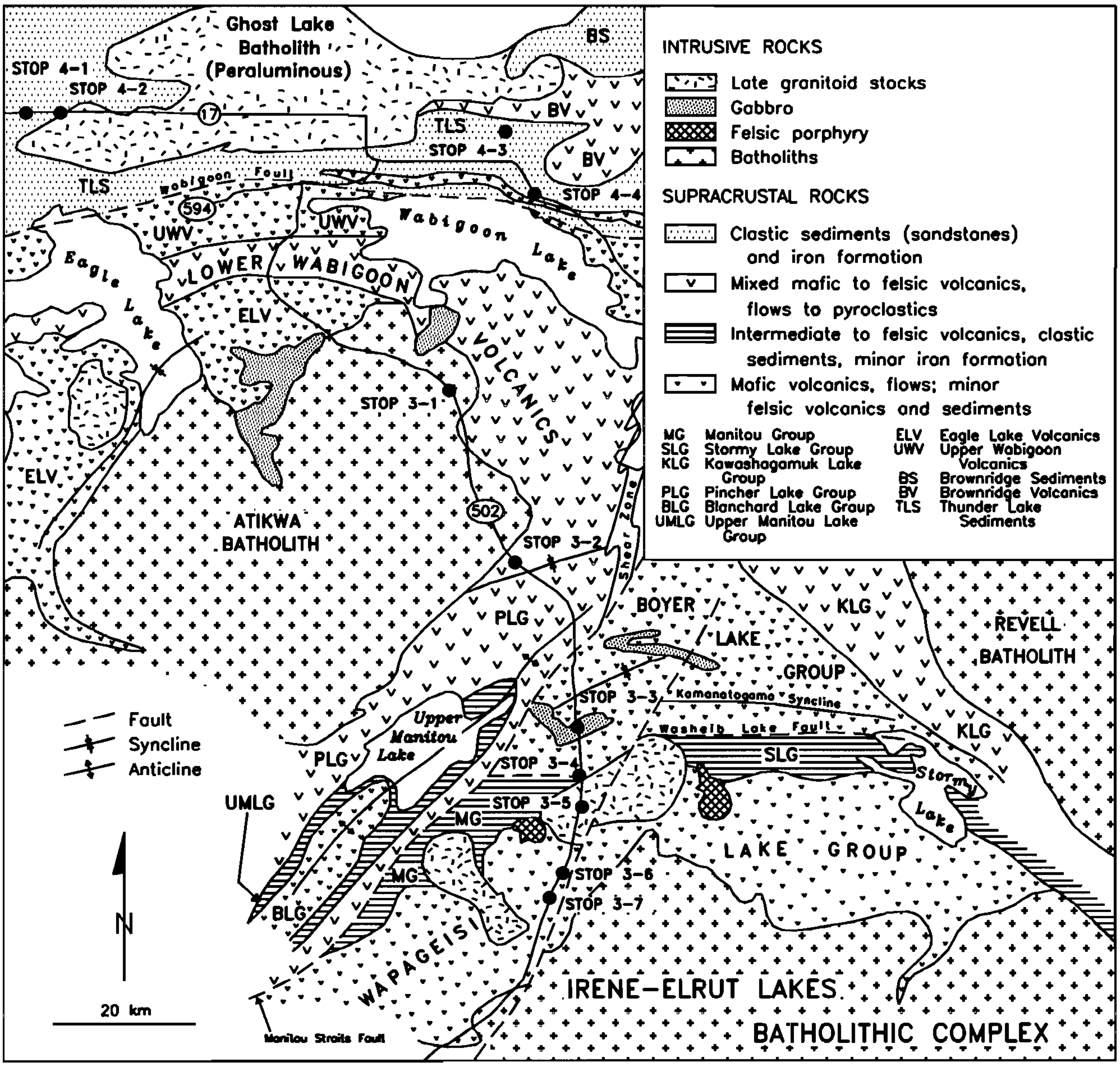


Figure 15. Geological map of the Dryden-Atikwa-Manitou area illustrating the location of stops on days 3 and 4.

composition from mafic to felsic. In some cases, the dikes appear to be composites of mafic and felsic phases but relative age relationships are not clear. Such dikes are common along the eastern margin of the Dore Lake lobe where they form sharply discordant dikes. In the interior of the lobe, the dikes are less abundant and, in some instances, form less continuous, locally amoeboid masses, possibly representing co-mingled magmas. The predominance of finer grained, sharply discordant dikes phases along the eastern contact may be indicative of the cooled carapace to an active magma chamber.

Return to vehicles and continue south on Highway 502 (Manitou Highway). At about 38 km granitic rocks of the Dore Lake lobe pass through a series of diorites into mafic metavolcanics of the Pincher Lake group of the Manitou-Stormy lakes greenstone belt.

Manitou-Stormy Lakes greenstone belt- *Stratigraphic and structural relationships in this greenstone belt, supported by age dating (Davis et al 1982; Davis 1990; Parker et al 1989: summarised in Blackburn et al 1991), suggest a history in which two early mafic tholeiitic sequences, the Wapageisi Lake group (not dated directly, but in the range 2745 to 2730 Ma) and the Boyer Lake group (not dated directly, but > 2722 Ma) have been juxtaposed against an intervening sequence of later calc-alkalic pyroclastics and overlying clastic metasedimentary rocks (Manitou-Stormy groups: <2706 Ma and >2696Ma). To the northwest of the Manitou Straits fault, supracrustal units (Upper Manitou Lake, Pincher Lake, Lower Wabigoon volcanics: >2732 Ma:) that extend northward to the Wabigoon fault are examples of volcanic ejecta that are invaded by their own magma chamber (Atikwa batholith, Dore Lake lobe: 2732 Ma). The early mafic suites are distinctly different, and may represent the ensimatic mafic plane within which the emergent pyroclastic edifices rose in island arc-related environments. The mafic plane analogue is supported by the presence within Wapageisi volcanics of two 200 m thick, laterally extensive (>10 km strike length) plagioclase-phyric flow units, 1500 m apart.*
Differing lines of evidence from two tabular gabbro bodies (Gabbro Lake, Mountdew Lake) within the Boyer Lake group lead to different interpretations of their relative age of emplacement. Presence of layering in the Gabbro Lake body suggests that they are horizontally-emplaced sills, and regional cross-cutting relationships suggest their emplacement subsequent to major folding. This argument led to the suggestion (Blackburn 1980) that they were emplaced subsequent to overthrusting of the Boyer Lake volcanics. However, an age of 2722 Ma later obtained (Davis et al 1982) from the same sill is <u>*not*</u> *consistent with a post-thrusting emplacement. These problems remain unresolved.*

Drive to a road cut at 39.9 km.

Stop 3-2. Giant pillows of the Pincher Lake group.
Refer to figure 15 for location.

The compositionally variable Pincher Lake metavolcanics are predominantly calc alkaline, and range from basalt to rhyolite. Pillow shape and size is much more variable than that in early tholeiitic mafic sequences. One such variety is seen at this stop. Giant pillows that extend the height of the road cut on the east side of the highway are at least 5 m in exposed long dimension, and 2 m thick. Pillow selvages are narrow, and a few small vesicular pillows occur intercalated between the giant pillows.

Gold in the Manitou-Stormy Lakes greenstone belt- *Past-producing mines at Gold Rock at the northeast end of Upper Manitou Lake, from which 12 thousand ounces of gold were intermittently produced during the period 1900 to 1948, were located in a broad zone of carbonatization and shearing within the Pincher Lake group. The zone is located along a splay of the Manitou Straits fault, a regional northeast-striking sericitic schist zone, centrally located within the greenstone belt, that effectively divides the belt into two contrasting structural entities. The splay is similar to others off this centrally located fault zone, such as that 80 km to the east that hosts Nuinsco Resources Ltd's Cameron Lake mine where over 1/2 million ounces of gold have been identified. Age of epigenetic gold mineralization elsewhere in the Manitou-Stormy lakes belt is later than a felsic flow dated at 2703 Ma, and probably younger than the post-tectonic Taylor Lake felsic stock at 2696 Ma (Parker et al 1989).*

Continue south on the Manitou Highway.

At 43.9 km on the left side (east) of the road shallow blasting in pillowed basalt was done by a local prospector on a narrow quartz vein that yielded spectacular coarse gold. Drilling beneath the vein and trenching along strike failed to find extension to any depth or length. This is on an extension of the Gold Rock splay. Between about 47 and 48 km the highway follows a swampy depression that is the trace of the Manitou Straits fault. From this point on we traverse Boyer Lake group rocks, mostly mafic tholeiites, the overthrust equivalents of the Wapageisi group.

We will stop at 54 km, at the foot of a down-hill stretch, stop on the shoulder of the road. USE

CAUTION AT THIS STOP - THE HIGHWAY IS NARROW, WITH VERY LITTLE SHOULDER.

Stop 3-3 Mountdew gabbro.

Refer to figure 15 for location.

The outcrop on the east side of the road is in gabbros of the Mountdew sill, close to its southern side. Only gross layering has been observed in this body, though phases similar to those in the Gabbro Lake sill are present: leucocratic olivine-pyroxene-plagioclase gabbro; diabasic pyroxene-plagioclase gabbro; pyroxenite (+/-plagioclase); pegmatitic pyroxene-plagioclase gabbro. At this stop we will see two phases: a dark green gabbro with large, dendritic amphibole; and a quartz-eye gabbro phase. Magnetite content is variable. Pyrite occurs in late, hair-line fractures. Apparent lack of differentiation and diagnostic structures precludes certainty as to horizontal, sill-like emplacement. The body lies predominantly in the south limb of the Kamanatogama syncline, but appears to cross cut the axial plane.

Proceed to Rattlesnake Creek, at 57.9 km, stopping on the south side of the valley.

Manitou and Stormy Lake groups - calc-alkaline pyroclastic edifices, alkaline flows, and marginal epiclastic facies- *The correlatable Manitou and Stormy Lake groups are typical of upper, emergent, chemically and texturally diverse, predominantly calc-alkaline sequences that provided fill for marginal sedimentary basins. In other greenstone belts these have been termed "Temiskaming" type. The base of the sequence unconformably (in places with high angle) rests on Wapageisi Lake group tholeiites. Calc-alkaline pyroclastics are typically coarse, and intruded by co-magmatic, dacitic, subvolcanic porphyry stocks, one of which has been dated at 2699 Ma (Don Davis, personal communication, 1989). At the top of the pyroclastic sequence lies a mafic alkaline (trachybasalt) flow unit (Sunshine Lake formation) that is unique in the western portion of the Wabigoon subprovince. Rubble at the top of the unit marks the passage into the epiclastic suite, in which there is vertical and lateral facies variation from alluvial fan and fluvial into resedimented conglomerates, sandstones and mudstones. The resedimented facies show all the characteristics of turbidite sequences: coarse heterolithic conglomerates, graded wacke to mudstone beds, and loading textures (eg. flame structure) (Teal and Walker, 1977; Blackburn 1981).*

The field trip route crosses a portion of the belt where a major fault, combined with intrusion of a post-tectonic stock, has cut out most of the above-described sequence: only the alkaline flows can be viewed.

Stop 3-4 Sunshine Lake alkaline flows, and Mosher Bay-Washeibemaga (MBW) fault.

Refer to figure 15 for location.

The river valley follows the east-striking MBW fault, along which the Boyer Lake volcanics may have been thrust over the Manitou and Stormy Lake groups. In the outcrop on the east side of the highway alkaline volcanics of the Manitou group are intruded by granitic rocks of the late tectonic Taylor Lake stock. Sigmoidal shears in the alkaline volcanics may be related to movement on the MBW fault. To the north across the valley is a narrow gabbro unit, similar to those at the last stop, emplaced into the Boyer Lake volcanics.

Although not evident at this stop, the MBW fault is cut out for a portion of its length by the Taylor Lake stock, and has in turn been sinistrally offset along the north northeast-striking Taylor Lake fault.

Proceed to 58.8 km, well within the Taylor Lake stock, on a straight stretch of highway passing into a right-hand turn.

***Stop 3-5 Taylor Lake Stock** (Optional Stop)*

Refer to Figure 15 for location.

The Taylor Lake stock is one of the volumetrically minor, but widely distributed, late- to post-tectonic stocks that mark the end of significant additions to the crust in the Wabigoon subprovince. The stock is inhomogeneous and ranges in composition from granodiorite to monzodiorite (Blackburn, 1981; Pichette, 1976) although the highway transects the comparatively homogeneous, granodioritic, western portion of the stock. Distinctive features of the granodiorite in the western portion of the stock include the presence of subhedral, late-magmatic microcline megacrysts as well as minor, widely distributed, dioritic enclaves. The stock has a U-Pb zircon age of 2695 Ma (Davis et al., 1982).

Continue south, passing into Wapageisi tholeiitic basalts, and stop in a road cut at 65.6 km. At this locality, at a left-hand bend in the highway, there is a high, steeply sloping rock face on the west side of the road, and lower, knobbly outcrops on the east side.

Stop 3-6. Pillowed lavas in Wapageisi volcanics.

Refer to figure 15 for location.

CAMERA OPPORTUNITY (brief stop).

A very good example of typical pillowed basalts of the lower, tholeiitic, mafic plane sequence. In the west face, pillow shape and packing indicates tops to the northwest. Good examples of bun-shaped to mattress-shaped to budding forms can be seen. Sparse feldspar phenocrysts are scattered throughout. Narrow selvages and lack of vesicles suggest deep-water emplacement but the presence of carbonate-filled gas cavities appears to contradict this. Size of pillows ranges from 20 cm or less to about 1 m. On the east side of the highway, excavation for the road cut, followed by winter frost heave, has exposed exceptionally well preserved three-dimensional pillow shapes. In one specimen, the budding neck of the pillow can clearly be seen in three dimensions.

Continue south to 67 km, still in Wapageisi volcanics, stopping at the Meggisi Road turnoff to the east. We will walk up-section, from this point to the Uphill Road turnoff to the west, a distance of about 400 m.

Stop 3-7. Variety of structures and textures characterizing the Wapagesi volcanics.

Refer to figure 15 for location.

The section commences in plagioclase-phyric pillowed basalt, typical of the two laterally extensive, plagioclase-phyric flow units that provide stratigraphic markers in the Wapagesi volcanics. Two overlying flows are crossed as we walk up-section:

- massive aphyric basalt passes into pillowed to pillow-brecciated basalt at the top of the first flow
- massive basalt with large ("eyebrow" structure) gas cavities passes into the pillowed top of the second flow.

At the Uphill Road turnoff, overlying massive basalt of a succeeding flow is seen.

Various planar structures, including flow contacts, "eyebrow" structures, and narrow (<0.5 cm) cherty beds, indicate a strike of 220°, dipping 60° to the northwest.

This is the final stop on the section through the Manitou-Stormy lakes greenstone belt. The next optional stop is described for the benefit of those who may in the future use this guide informally as it involves a further 80 kilometers of driving to the south.

As we continue south on Highway 502 we pass a commemorative plaque (opening of the Manitou Highway) on the right hand side at about 73 km, where the road passes the southern end of Scattergood Lake. From this point on, there is a shallow dipping foliation to the northwest in the mafic volcanics at the southern edge of the greenstone belt. Passing out of the belt, a drive of about 25 km takes us through the western end of the Irene-Eltrut lakes batholithic complex. The complex is part of the central Wabigoon region. After passing through another greenstone belt septum (Otukamamoan Lake belt), and more granitic rocks, stop at 147 km, where an east-trending arm (Crowrock Inlet) of Rainy Lake marks the trace of the Quetico fault.

Quetico Fault- *Most workers consider the Quetico fault to be a major structural feature in Superior Province, extending in an east-west direction a distance in excess of 200 km, from the southern end of Lake of the Woods in the west to north of Thunder Bay in the east. Over much of its distance the fault parallels and in part defines the boundary between the Wabigoon and Quetico subprovinces. Movement along the fault has been variably estimated by different workers, most suggesting it to be overwhelmingly dextral, and some suggesting it to be in excess of 100 km (eg. Mackasey et al 1974). Estimates have been based for example on correlation of similar, and unique, ultramafic pyroclastic units on either side of the fault at Atikokan and Rainy Lake (Schaeffer and Morton 1981). The fault differs from major faults along central portions of Wabigoon subprovince greenstone belts in that 1) it cuts across supracrustal belts and major batholithic domes (eg. Rainy Lake batholithic complex) 2) the fault is straight 3) mylonitic rocks are characteristic in the fault zone. The fault was operative over a protracted time period, into the Paleoproterozoic (Peterman and Day 1989). The fault zone takes on differing characteristics within the variety of rock types that it transects. In the Rainy Lake area, a wide zone containing mylonitic fabrics occurs in granitoid rocks. In the Atikokan area, the fault has either been recognized as a series of splays in metavolcanic rocks (Fumerton, 1982) or even discounted as a major feature (Stone et al., 1982).*

Stop 3-8 Quetico Fault at Crowrock Inlet.

Refer to figure 3 for location.

Steep road-cut faces on the west side of the highway, just north of the causeway over Crowrock Inlet, expose rocks within the fault zone. At this locality the zone is on the order of 500 m wide, the width of the inlet. Porphyroclastic to mylonitic bands, with development of augen gneiss, within otherwise little deformed granitic rocks are indicative of variable strain rates within the zone.

This ends stops for day 3. We will retrace our route back to Dryden.

Day 4 Dryden Area

The fourth day of the trip will provide an opportunity to examine geological relationships in the northern portion of the Western Wabigoon subprovince in the Dryden area. Some aspects of the general geological relationships differ from those we examined on days 2 and 3 of the trip. The significance of these differences can be debated during this portion of the trip.

Sioux Lookout Domain *- The northern portion of the western Wabigoon region differs from that portion examined on day 3 of the fieldtrip in several important respects:*

- *Metasedimentary rocks are anomalously abundant with respect to their abundance elsewhere in the Wabigoon subprovince although certain segments (e.g., Lake of the Woods area) are still volcanic dominated.*
- *Metamorphic grade ranges from greenschist facies to amphibolite facies with conditions locally sufficient to induce partial melting of metasedimentary rocks (Dryden area).*
- *Granitoid plutonism is distinctive. The strongly peraluminous Ghost Lake batholith is interpreted to be produced by in situ partial melting of the metasedimentary rocks. The Dryberry batholith is strongly magnetic (like similar batholiths in the Winnipeg River subprovince), is not associated with coeval volcanic rocks and truncates, rather than folds, stratigraphy. This batholith, as well as the Snowshoe Bay batholith to the west (Davis and Smith, 1991), may post-date collision and be derived through intracrustal melting of underthrust Winnipeg River type older crust in much the same manner as the granitic suite of the Winnipeg River subprovince (see earlier discussion).*
- *Structural style is dominated by linear trends of shear zone bounded panels*

parallel to the subprovincial boundary. Early recumbent folds are recognized locally and thrust faults are also inferred from geochronological evidence for out of sequence stratigraphy (Davis et. al., 1988). The domain also incorporates a style of asymmetric minor folding interpreted to reflect a broad zone of dextral transcurrent shear that is not seen to the south of the Wabigoon fault.

The Sioux Lookout domain is metallogenically distinct by virtue of the presence of rare element pegmatite mineralization and, arguably, a higher proportion of pluton hosted Mo-Cu-Au mineralization.

These observations and interpretations have lead to the suggestion (Beakhouse, 1989) that this area is a discrete domain having different tectonic significance than that portion of the Wabigoon subprovince lying to the south of the Wabigoon fault. This hypothesis would hold that the metavolcanic rocks and related volcaniclastic metasedimentary rocks of this domain are allochthonous and were thrust over the Winnipeg River subprovince in a terminal arc - continent collisional event. Late unconformable sequences (Crowduck Group, Ament Bay Group) and high-grade metamorphism/ intracrustally derived granitoid rocks are responses to the uplift and thickening, respectively, associated with collisional orogeny. The area is imperfectly defined but is provisionally interpreted to be 15-40 kilometres wide and to extend for at least 250 kilometres. For much of this length, its northern and southern extent are defined by the contact with the Winnipeg River subprovince and the Wabigoon fault respectively (Figures 2 and 3).

Drive west from Dryden to a outcrop ~2.9 km past the Aubrey Creek bridge. This is located approximately 15 km east from Vermilion Bay along Highway 17 if approaching from the opposite direction.

Stop 4-1: High grade migmatitic metasedimentary rocks immediately west of the Ghost Lake batholith.

Refer to Figure 16 for location.

This stop is intended to illustrate the character of high grade, migmatized, interbedded metawacke/metamudstone rocks that comprise the host-rocks for the Ghost Lake batholith along its western flank (Figure 16). Such rocks likely constituted the protolith from which were derived peraluminous granitic melts and related rare-element (Li, Cs, Rb, Be, Ta and Nb)-enrichment in pegmatitic granite units that comprise the batholith.

The paleosome constituent at this low leucosome-fraction migmatite exposure consists of medium grained, metamorphically coarsened, sillimanite-garnet-cordierite-biotite-K-feldspar metapelite which contrasts with, interlayered, finer-grained garnet-biotite-quartz-plagioclase metawacke. A low percentage of foliation-parallel, 1-10cm-thick, white, biotite and garnet-biotite granite leucosome is apparent.

The migmatitic metasedimentary rocks were intruded by dykes, 3 to 15 meters-

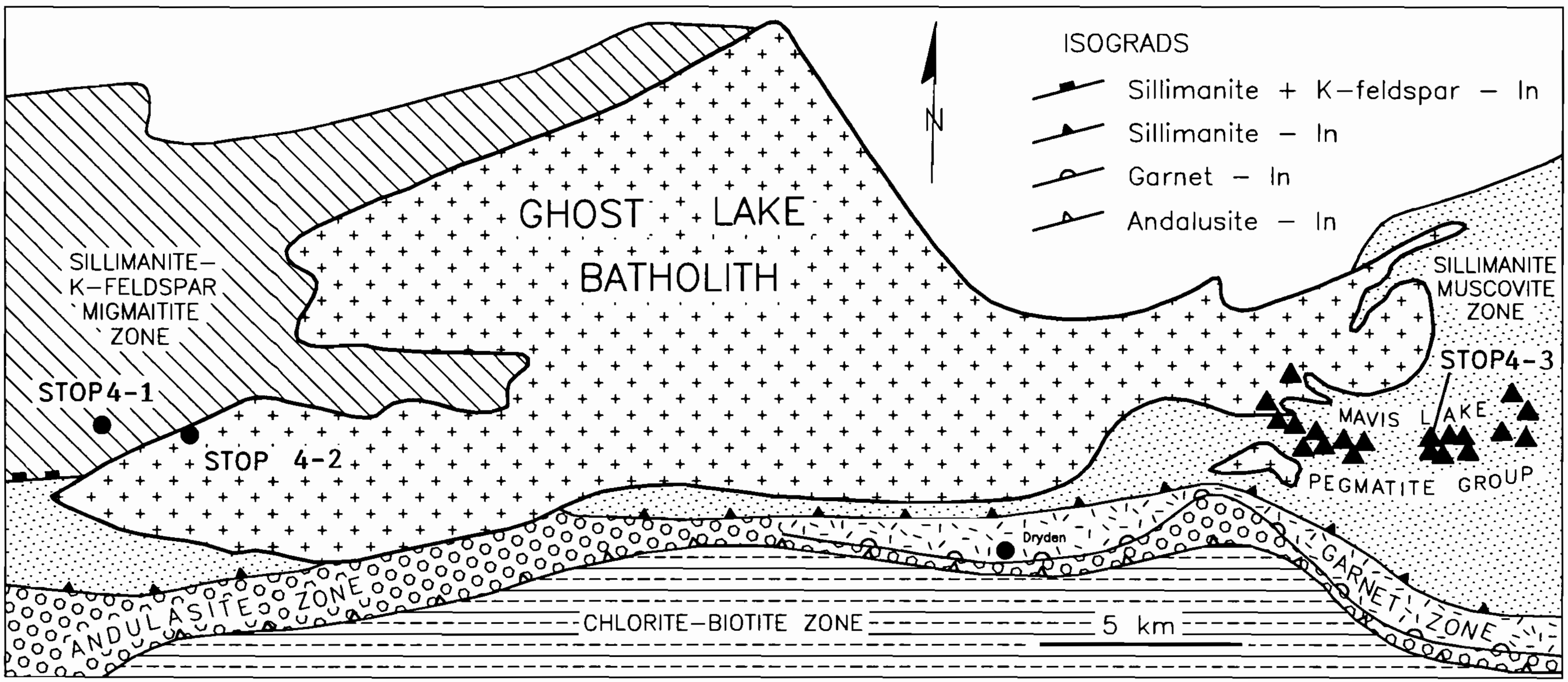


Figure 16. Distribution of regional metamorphic zones and isograds in relation to the Ghost Lake batholith and rare-element pegmatites of the Mavis Lake Group and location for 4-1.

thick, of peraluminous granitic rock quite similar to that found in the western part of the nearby Ghost Lake batholith (unit GLB-1: Breaks and Moore 1992). These inequigranular dykes reveal a large range in grain size (medium grained to pegmatitic) and contain a suite of peraluminous accessory minerals identical to that found in the batholith: biotite, garnet, cordierite, sillimanite, black tourmaline and rare dumortierite. Sea-green apatite and secondary light green muscovite round out the accessory mineral population. Dumortierite occurs as fine violet prisms in some of the graphic K-feldspar-quartz megacrysts. The largest pegmatitic dyke notably contains clots, up to 28 cm across, and irregular veins, up to 10 by 75 cm, of biotite-cordierite-quartz up to 5 by 75 cm. The cordierite is evident as fresh, mauve, commonly subhedral grains comprising about 40% of the clot. Sillimanite occurs in porcellaneous, fine grained masses of the fibrolite variety. Similar clots will be noted in the western part of the Ghost Lake batholith at the next stop.

Drive 2.9 km east on Highway 17 to the outcrop at the Aubrey Creek bridge.

Stop 4-2: Western contact of Ghost Lake batholith with migmatitic metasedimentary rocks

Refer to Figure 17 for location.

This outcrop presents a rarely exposed contact of the western part of the Ghost Lake batholith, a complex described in detail by Breaks and Moore (1992) which developed a rare-element pegmatite swarm adjacent to its most chemically evolved part at the eastern end. This peraluminous granite complex is somewhat unique amongst other known Archean fertile granite masses as it not only contains a highly evolved fertile pegmatitic granite facies (GLB-4 to -8: Figure 17) but also possesses more primitive granite units normally not exposed in smaller fertile granite masses. The latter will be examined in this stop which corresponds to unit GLB-1 of Breaks and Moore (1992).

Revealed along the north side of the road cut is a sharp contact between cordierite-biotite pegmatitic leucogranite and flat-lying, migmatitic, metasedimentary rocks similar to those examined in the previous stop. Flat-lying isoclinal folds in leucosome are locally apparent in the host-rocks. Unit GLB-1 is composed of sillimanite-muscovite-biotite- and cordierite-biotite pegmatitic leucogranite and less coarse garnet-sillimanite-muscovite granite. Most of the muscovite at this outcrop is likely secondary. The pink pegmatitic leucogranite is characterized by irregular to blocky, graphic megacrysts of perthitic, microcline ($Or_{74}Ab_{26}$)+quartz up to 26 by 60 cm and abundant cordierite. The latter mineral occurs as euhedral, square (up to 2 cm) and rectangular (up to 1 by 3 cm) crystals. Cordierite occurs

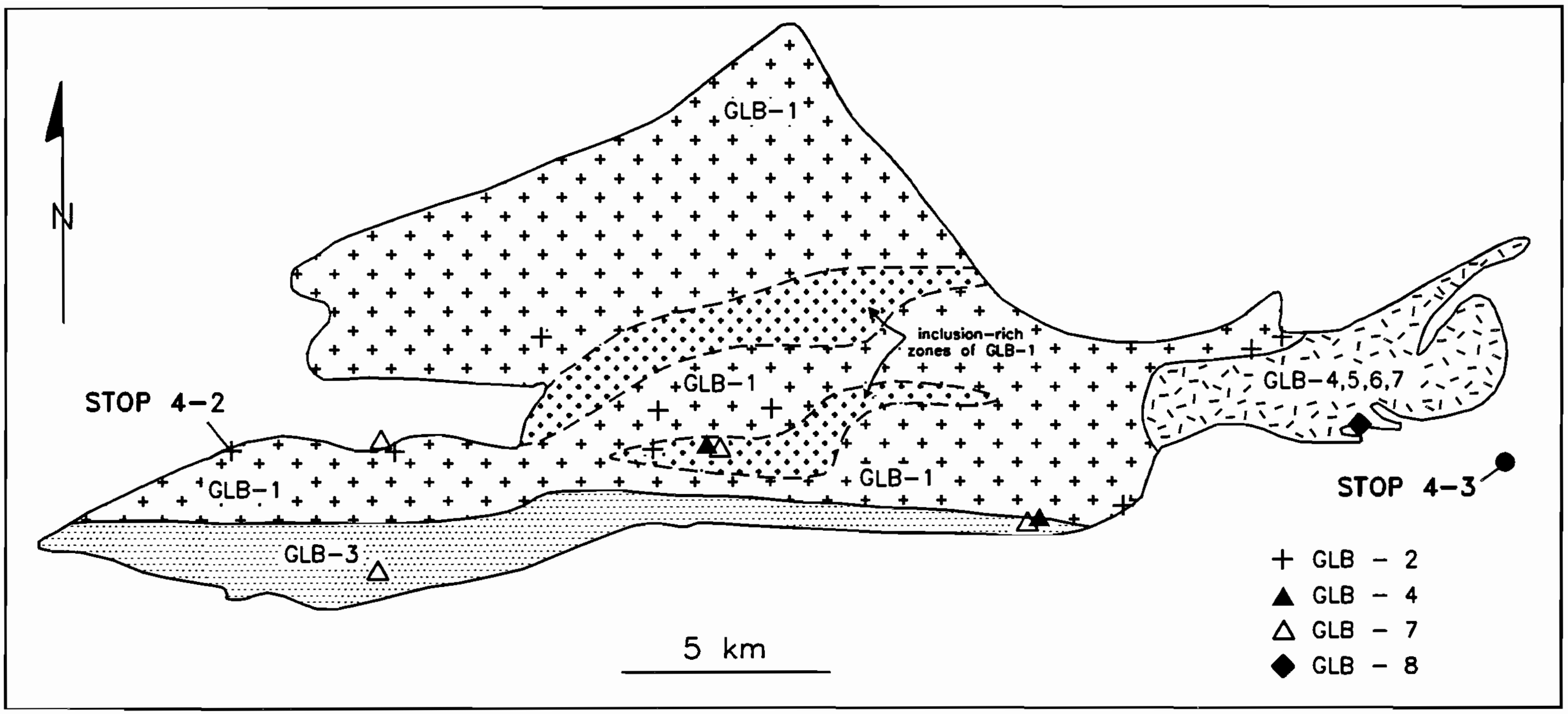


Figure 17. Distribution of internal granitic units in the Ghost Lake batholith and location of stops 4-2 and 4-3.

in the coarse grained matrix to the microcline-quartz graphic megacrysts, but rarely are found as inclusions within these. In addition, cordierite is found within large ovoid intergrowths accompanying approximately equal amounts of quartz; minor plagioclase and garnet may be present. These symplectites are more common along the exposure along the south part of the road nearest to Aubrey Creek. Here one finds the largest single quartz-cordierite intergrowth, apparent as a roughly circular section, up to 1.5 meters across, which contains about 50% cordierite. Most of the cordierite here has been replaced by a randomly oriented mesh of biotite and chlorite.

Black tourmaline occurs sparsely in garnet-biotite pegmatitic leucogranite and within an intergrowth with quartz in a coarse pod hosted in garnet-biotite aplite on the south side of the road. Rare, blue, fine grained dumortierite has also been verified for this exposure.

Anomalous trace levels of some of the rare-elements occur in certain niches of GLB-1 at this outcrop: cordierite-quartz clots (Be =35-47 ppm and Li =108-136 ppm); biotite-sillimanite restite (Li = 32 ppm, Cs = 30 ppm, Ga = 48 ppm, Rb = 371 ppm).

Return to vehicles and travel 34 km east on Highway 17, passing through the Town of Dryden. Turn north on to the Thunder Lake Road and proceed for 1.7 km. Turn north off an east-trending section of this road on to the Mavis Lake Forest Access Road. Travel for 3 km and then park vehicles on road side. Walk for approximately 500 meters east on old drill road.

Stop 4-3: Spodumene-beryl-tantalite pegmatites, Mavis Lake Pegmatite Group

Refer to Figure 17 for location.

The Mavis Lake Pegmatite Group consists of an east-striking, 8 km by 0.8-1.5 km, concentration of rare-element pegmatites and related metasomatic zones. It exhibits a well defined zonation of pegmatite mineral assemblages with increasing distance from the parent Ghost Lake batholith (Figure 18 and Breaks and Moore 1992, p. 847). Twelve spodumene pegmatites are known in this zone and these range in size from 3 by 15 meters to 15 by 280 meters. Most of these lensoidal bodies strike parallel to the foliation in the host mafic metavolcanic rocks, however, dips vary in direction and amount (40-75^0 N and 50-80^{0}S). At this stop we will examine two spodumene pegmatites and associated metasomatic rocks at the western extremity of the spodumene-beryl-tantalite zone.

Fairservice Pegmatite No.1

This 12 by 75 m pegmatite (Figure 19) occurs in foliated and gneissic mafic metavolcanic rocks and minor metawacke. The pegmatite exhibits a vague internal zonation (Figure 20), listed below in order of decreasing quartz content, which is generally lacking in pegmatites of the Mavis Lake Group:

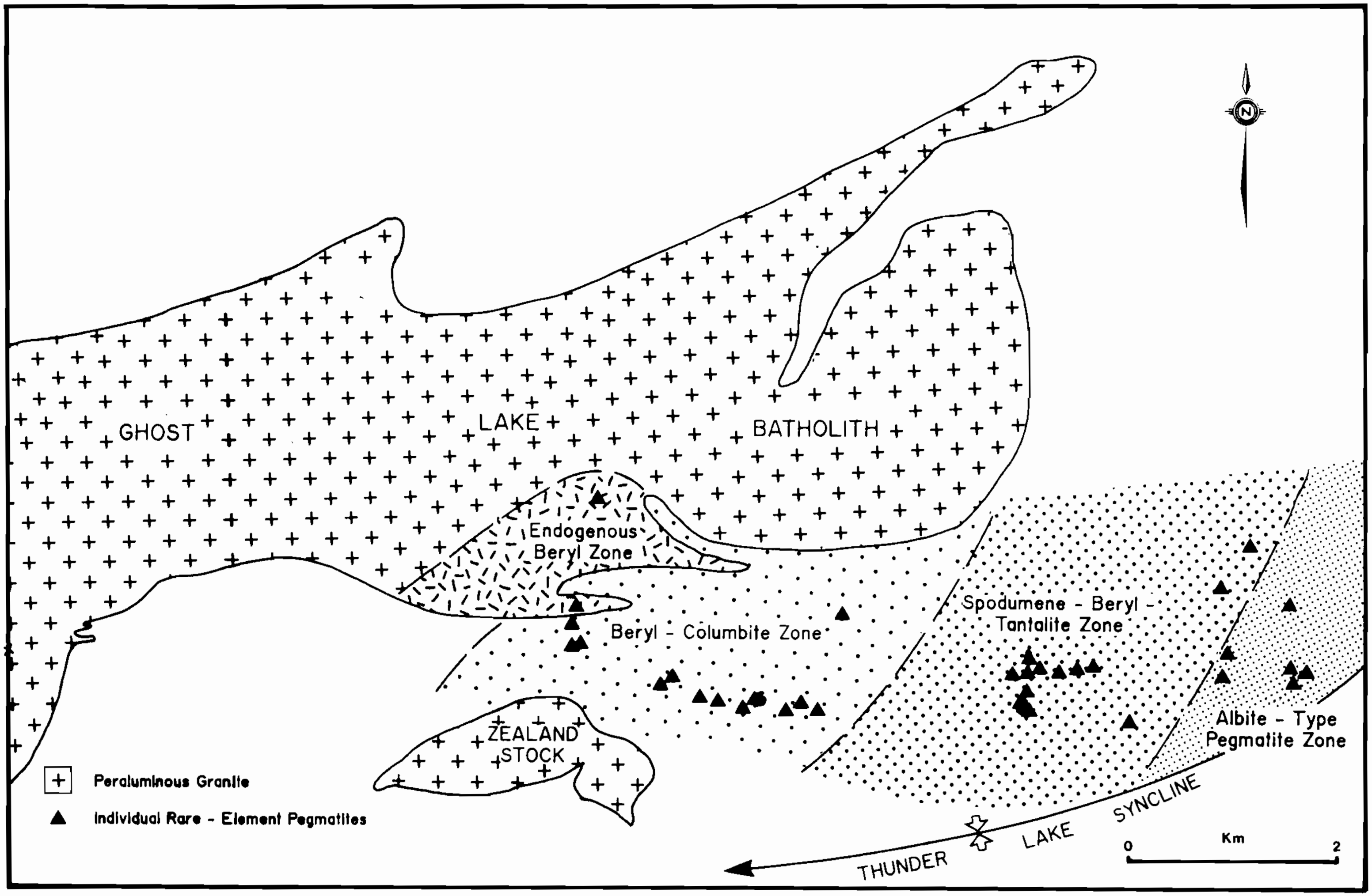


Figure 18. Distribution of rare-element pegmatite zones of the Mavis Lake Group in relation to the Ghost Lake batholith.

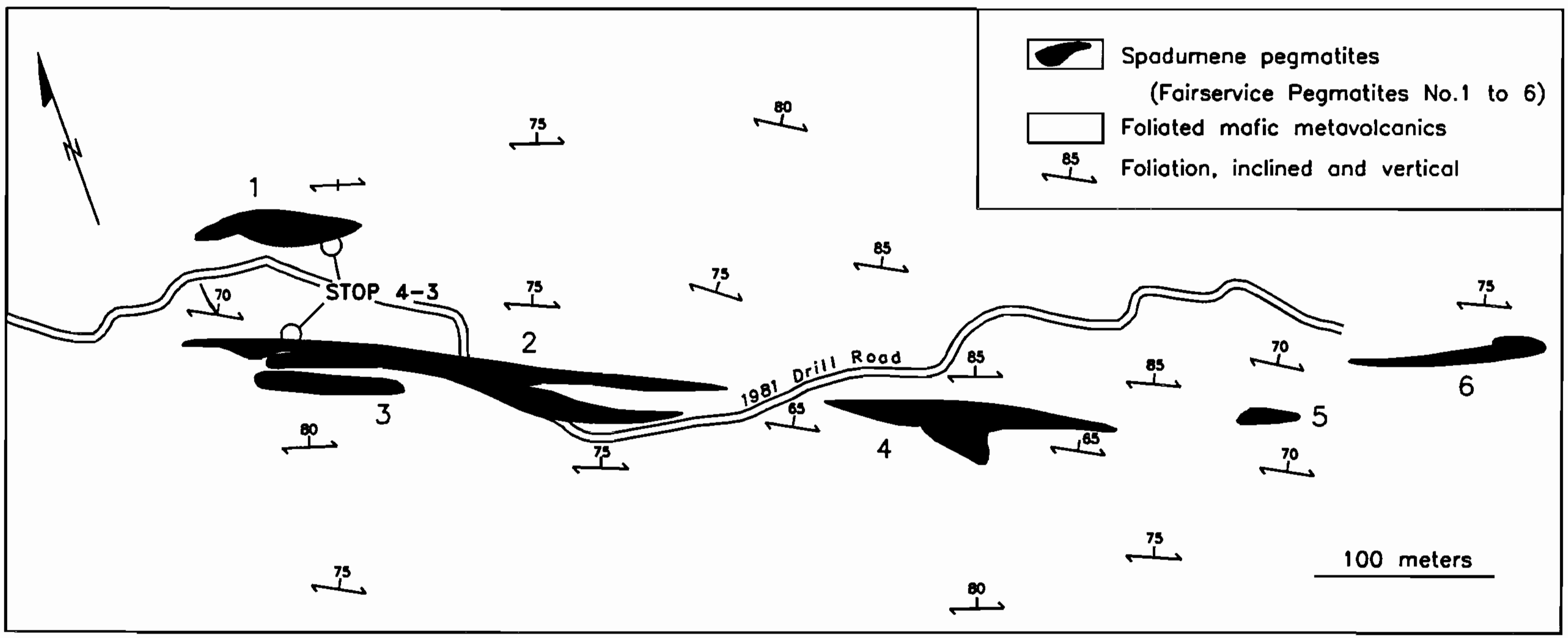


Figure 19. Distribution of spodumene pegmatites at the North Zone of the Fairservice Property and location of stop 4-3.

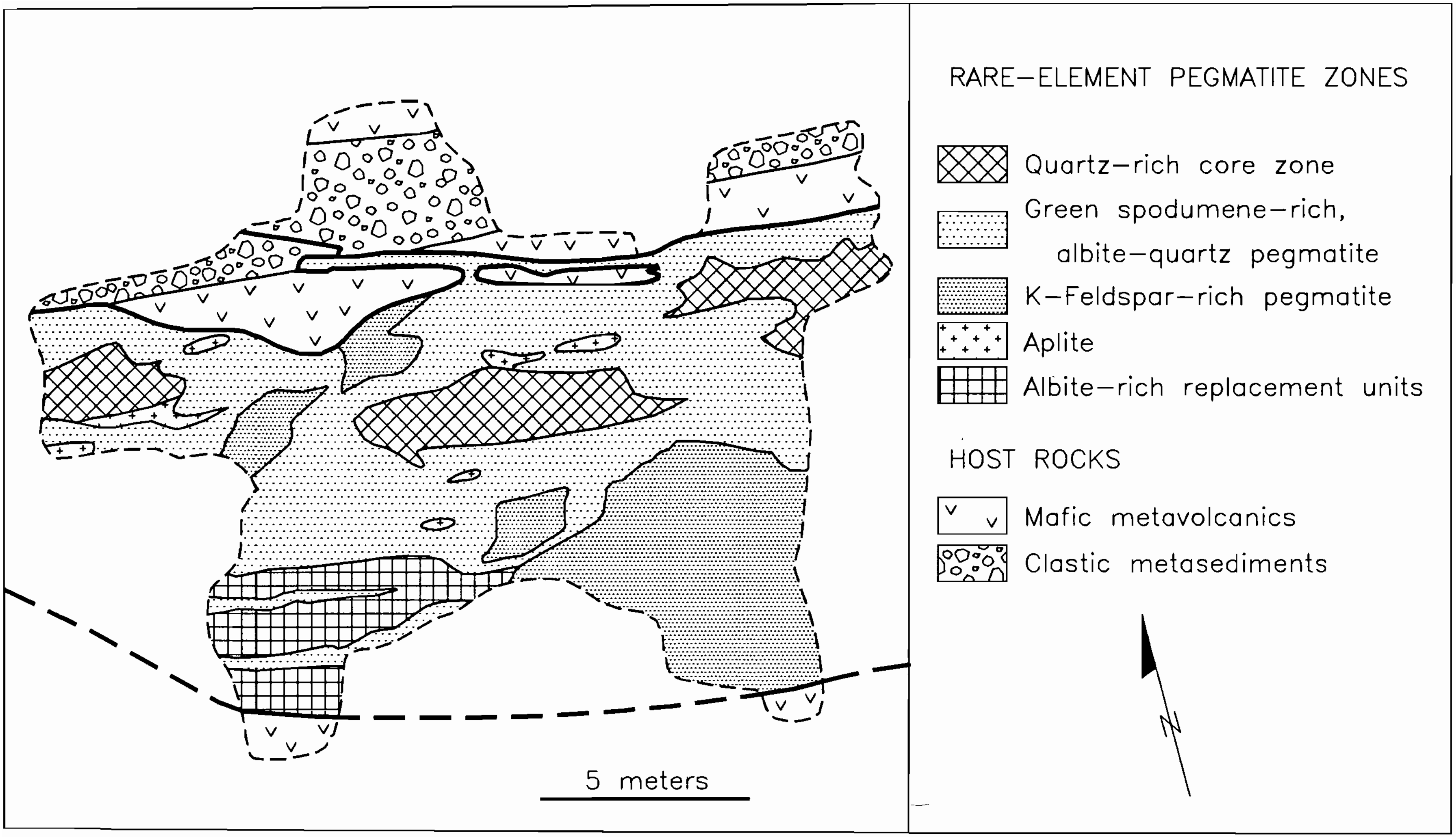


Figure 20. Detailed geology of part of Fairservice Pegmatite No. 1, revealing its crude internal zonation.

1. Quartz-rich core zone
2. Green spodumene-rich, albite-quartz pegmatite, and,
3. K-feldspar-rich pegmatite.

Quartz-rich Core zone

This zone exhibits a discontinuous distribution and occurs as patches between the pegmatite centre and the north wall (Figure 20). These irregular patches, up to 2 by 9 meters, contain conspicuous single crystals and aggregates of blocky microcline ($Or_{74}Ab_{24}$), individually up to 33 by 41 cm, green spodumene, yellow-green muscovite and albite which are immersed in a matrix of 70-80% light grey massive quartz. Minor phases include white to light blue beryl (up to 3 by 7 cm), black tourmaline, blue apatite and orange garnet.

A characteristic texture that occurs solely in the core zone is that of spodumene megacrysts, up to 4 by 8 cm, that are partially to completely rimmed by fine-grained albite, minor muscovite and rare tantalite. Such textures likely originated during late stage albite-replacement processes that in which previously formed phases were attacked in response to changing fluid composition. Similar textures have been documented in the Black Hills area by Spilde and Shearer (1992, p.731).

Green Spodumene-Rich, Albite-Quartz Pegmatite

This zone envelops the quartz-rich core zone (Figure 20) and contains the highest quantity (50%) and largest size (9-12 cm by 1 m) of spodumene observed in any pegmatite of the Mavis Lake Group. High levels of quartz (37-47%) are also notable as is the meagre content of total feldspar (7-14%). Pockets of sodic aplite, up to 1 by 3 m, occur sparsely throughout the zone and these are considered as primary as no discernible replacement of contacting phases such as spodumene is detectible. This aplite is also distinguished from other aplite types by the presence of apatite and brown phosphate minerals, the latter speculated to have been triphylite but which was subsequently altered to heterosite-purpurite group minerals).

K-feldspar-Rich Pegmatite

This zone occurs mainly in the outermost parts of Pegmatite No. 1, particularly along the south wall (Figure 20). The unit is distinguished by a concentration of coarse, coalescence of light pink to white, blocky microcline (55%). Spodumene (7%) is considerably sparse relative to the adjacent spodumene-rich, albite-quartz unit and is restricted to intergrowths with quartz (23%)in the interstices between the blocky microcline crystals. Albite (23%) and green muscovite (3%) comprise the remaining phases. Beryl is quite rare in this pegmatite zone.

Albite-Rich Replacement Unit

Irregular, albite-rich replacement units typify all spodumene pegmatites of the Mavis Lake Group. In Fairservice Pegmatite No.1, about 10% of the dyke consists of intensely albitized rocks in which little vestige of the primary mineralogy has survived. The largest unit at 5 by 9 m is composed of an inequigranular assemblage of milky white beryl-green muscovite-quartz-albite and accessory garnet and tantalite. In other pegmatites of the spodumene-beryl-tantalite zone, the albite replacement units are characterized by lensoidal pseudomorphs of fine grained albite and green mica after spodumene.

Tantalite in Pegmatite No.1 is sporadically distributed and commonly is associated with replacement stage albitization, as in albite-rich rims on spodumene and along ragged, cleavage-controlled, albite-rich indentations into aggregates of blocky microcline (Analysis 406 in Table 1).

Proceed about 60 meters south to spodumene pegmatite dyke exposed along edge of small ridge.

Fairservice Pegmatite No. 2

The westernmost part of this bifurcated, 270 meter-long spodumene pegmatite dyke (Figure 19) mostly consists of the spodumene-bearing, K-feldspar-rich pegmatite unit observed at Pegmatite No.1. However, this pegmatite is distinguished by conspicuous megacrysts of green spodumene which are complexly intergrown with quartz. Slender spodumenes, up to 12-20 cm by 1 meter, are oriented approximately normal to the south-dipping, upper contact of the pegmatite. These megacrysts, which are typically embedded in coarse blocky microcline masses, contain abundant, much finer-grained, equigranular quartz which composes about 30% of the intergrowth. The spodumene-quartz symplectite is then mantled by a 4-5 cm-thick, quartz-green muscovite-albite assemblage.

This pegmatite exhibits local development of a quartz-rich core units which may contain blocky microcline, white euhedral beryl and faint green spodumene. Pockets of saccharoidal sodic aplite bearing phosphate minerals (green apatite and brown, possible former triphylite) are also sporadic and similar to the primary aplites examined at Pegmatite No.1.

Tantalite, although sparse, is more uniformly disseminated in this pegmatite relative to Pegmatite No.1. Maximum levels of Ta (380 ppm) and Nb (310 ppm) for any pegmatite in the Mavis Lake Group were obtained from analysis of channel samples of Pegmatite No. 2.

Late Stage Metasomatic Alteration of Host-Rocks

The mineralogical and chemical expression of the interaction of late-stage fluids/vapour with mafic metavolcanic wall-rocks can be clearly be seen at the small vertical face above the 50^0 south-dipping, upper contact. Here, subtle, dark purple, needles of holmquistite up to 4 cm in length, randomly transected the foliation and hornblende lineation . Other metasomatic phases include black tourmaline and bronze, rare-element-enriched biotite.

Chemical profiles which document the variation of metasomatically-enriched elements derived from the pegmatite are shown in Figure 21.The typical variation

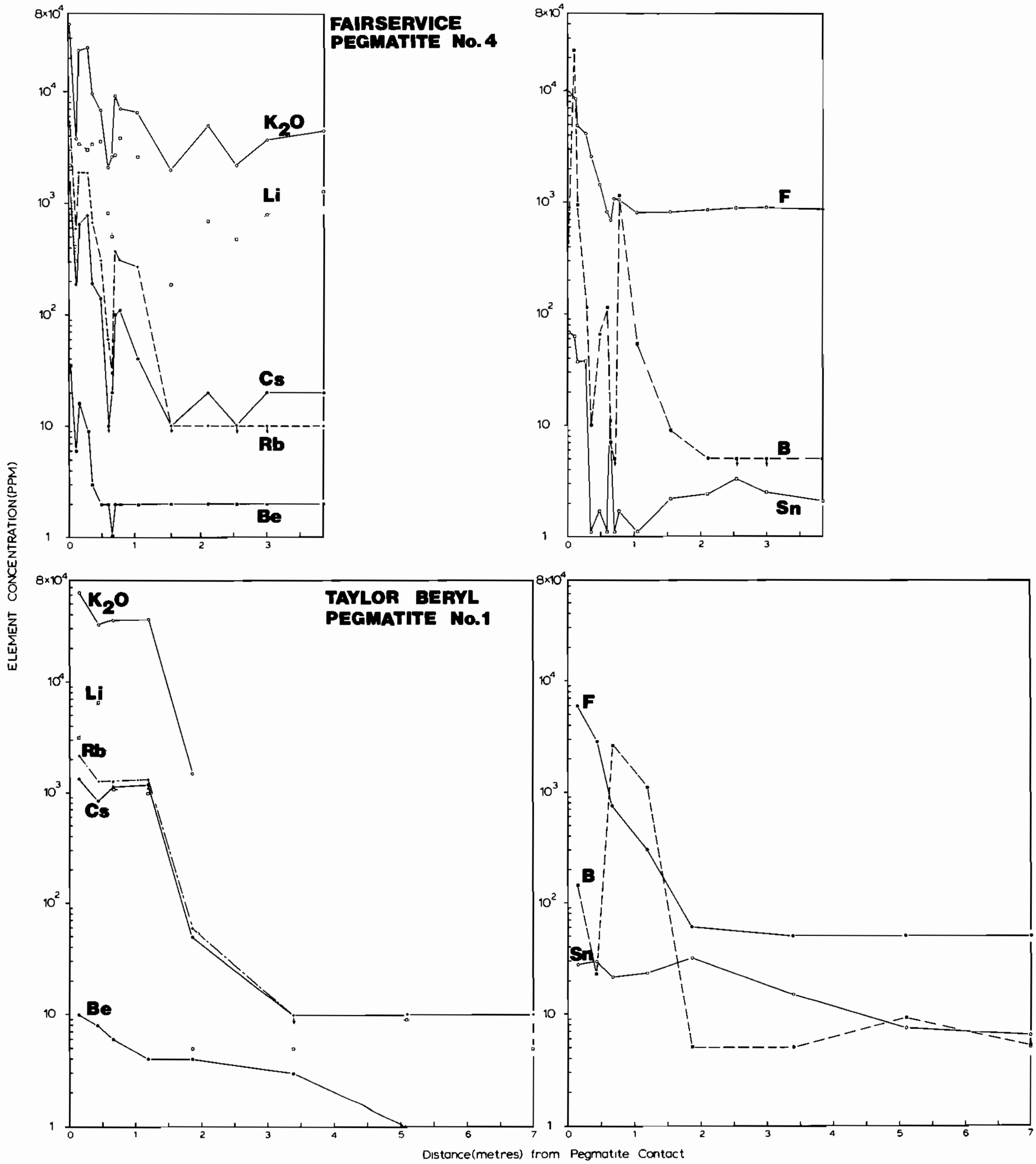


Figure 21. Profiles of concentration of various exomorphic elements in mafic metavolcanic host-rocks against increasing distance from the north contact of Fairservice Pegmatite No. 4.

of K, F, Li, Rb, Cs, Sn and Be in the mafic metavolcanic host-rocks with increasing distance from a spodumene pegmatite is given in Figure 21, for Pegmatite No.4, about meters east of this exposure.
Lithium generally forms the most extensive exomorphic aureole around the Mavis Lake Group spodumene pegmatites which substantiates the work of Beus et al.(1968) who concluded that lithium is one of the most mobile exomorphic elements. A large, elongate lithium anomaly, 120-400 by 3050 meters, marked by levels which exceed 100 ppm, envelopes most spodumene pegmatites of the area. Li_2O attains a maximum concentration in the mafic metavolcanic host-rocks of 1.5%. Boron and fluorine also exhibit extensive exomorphic dispersion (Breaks 1989); smaller aureoles characterize K, Rb, Cs, Sn and Be.

Return to the Mavis Lake Forest Access Road and thence to Highway 17.

Proceed east on highway 17 for approximately 3 kilometres. We will stop at a roadcut on the south side of the highway (Thunder Lake is visible to the north) near the turnoff for the Bell Canada tower. If proceeding directly from Dryden, this outcrop is located 11.6 kilometres east of the traffic lights at the Comfort Inn.

Stop 4-4 Fold style in the Thunder Lake metasediments

Refer to figure 16 for location.

Road-cut outcrops on either side of Highway 71 show style of folding and order of superposition of folding in this portion of Warclub group metasedimentary rocks. On the south side of the highway, on a smooth sloping glaciated surface, two phases of folding can be seen. Transposition of bedding into the plane of schistosity is seen where the limbs of first phase isoclinal folds have been sheared through, producing isolated fold noses. The form of these folds is a tight Z, in which axial planes are parallel to the local east-southeast formational strike. Differentially weathered beds in the metapelites and wackes serve to outline these folds, portions of some having been completely detached by shearing. A second phase of open Z folding refolds the first phase.
Vertical road-cut surfaces on both sides of the highway display steeply-plunging mineral lineation that is parallel to fold axes, probably of both fold phases.
Regional major folds such as the Thunder Lake anticline, that affect Warclub group metasedimentary rocks, including intercalated iron formation units, and metavolcanics of the Neepawa group to the north, have similar style and orientation (northeast-striking axial plane traces) to the second phase, open Z folds at this stop, and elsewhere within the supracrustal rocks north of the Wabigoon

fault. South of the Wabigoon fault, folding is about subhorizontal axes only, in contrast to the steep axes in evidence here.

End of Trip

Part C
Western Superior Transect: Wabigoon - Quetico - Shebandowan Portion

G.M. Stott, D. Stone, C. Farrow and F. Corfu

Day 5. Atikokan area

Day 5 features unique geology spanning 300 My of Archean crustal evolution in the southern Wabigoon and adjacent Quetico Subprovinces. Stops include the early Marmion tonalite batholith, a rare Archean unconformity, the Steep Rock Group (a platform sequence) and contiguous metavolcanic and metasedimentary rocks (Figure 22). An attempt is made to examine the structural relation between these rocks. Most stops are in vicinity of the abandoned Steep Rock Iron Mine from which approximately 100 Mt of ore were removed from 1944 to 1979. The stratigraphic relationships between units, and the locations of the stops are illustrated in figures 23 and 24, respectively.

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

The Marmion batholith *-The Marmion batholith is a large oval tonalitic intrusion extending at least 50 km east of Atikokan. Dated at 3003±5 Ma (Davis and Jackson 1988) it is among oldest felsic intrusions in the western Superior Province and together with the pre 2900 Ma Dashwa Gneiss Complex and Finlayson and Lumby greenstone belts (Davis and Jackson 1988) is part of the enigmatic older volcanoplutonic terrane in the central Wabigoon Subprovince.*
The Marmion batholith formed a mature erosional surface on which the Steep Rock Group was unconformably deposited at an unknown but probably early stage in tectonic evolution of the area. Although rare in central parts of the Marmion batholith, metagabbro dikes are voluminous near the Steep Rock belt; some dikes seem to predate whereas others postdate the Steep Rock Group. Although the tectonic environment at the time of Steep Rock Group deposition is speculative, the dikes imply extensional conditions, possibly in a rifted continental margin setting (Wilks and Nisbet 1988).
Felsic and mafic varieties of tonalite are mapped in the Marmion batholith near the Steep Rock belt.

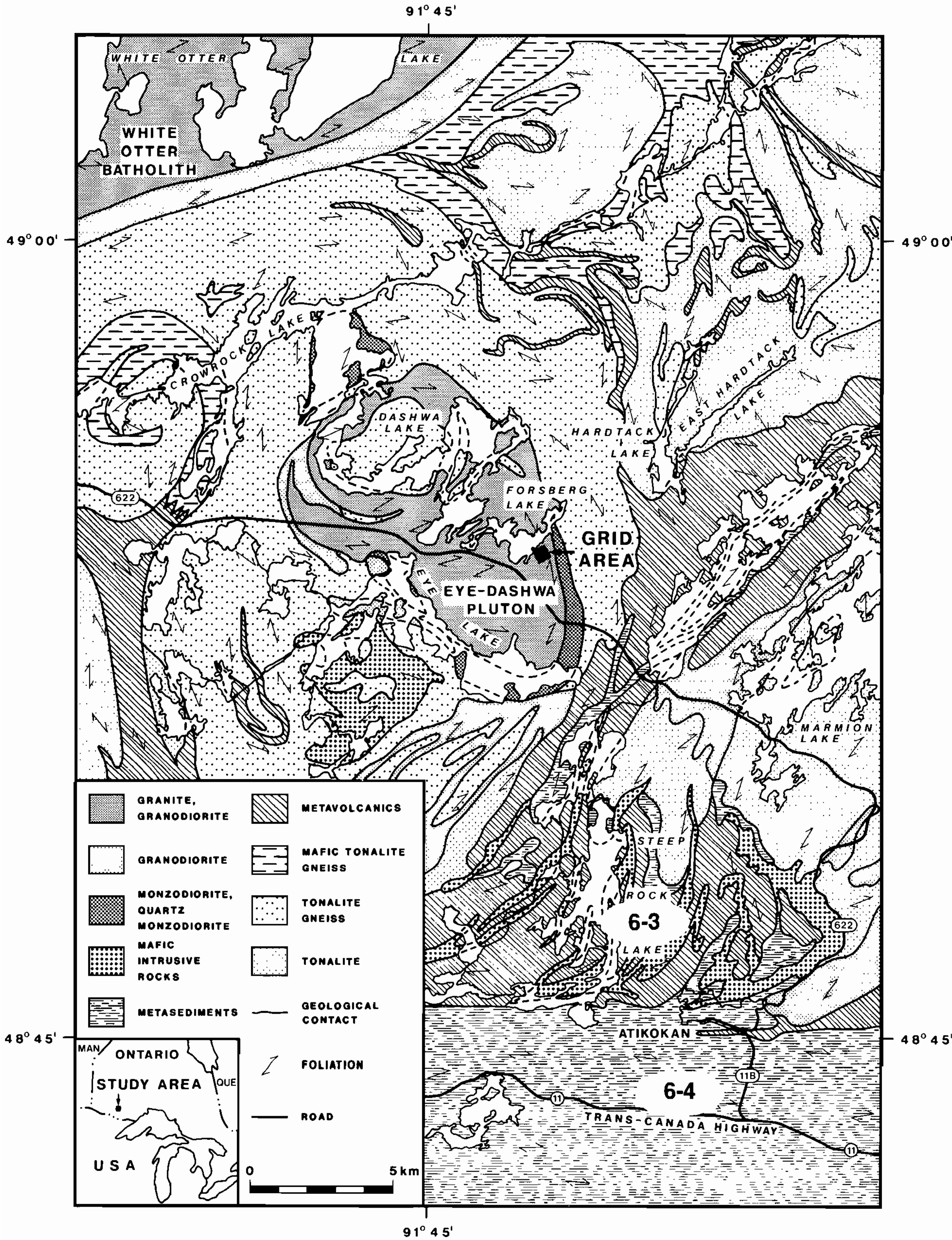


Figure 22. Geologic map of the Atikokan area showing field-trip stops 6-3 and 6-4. See figure 24 for other stops.

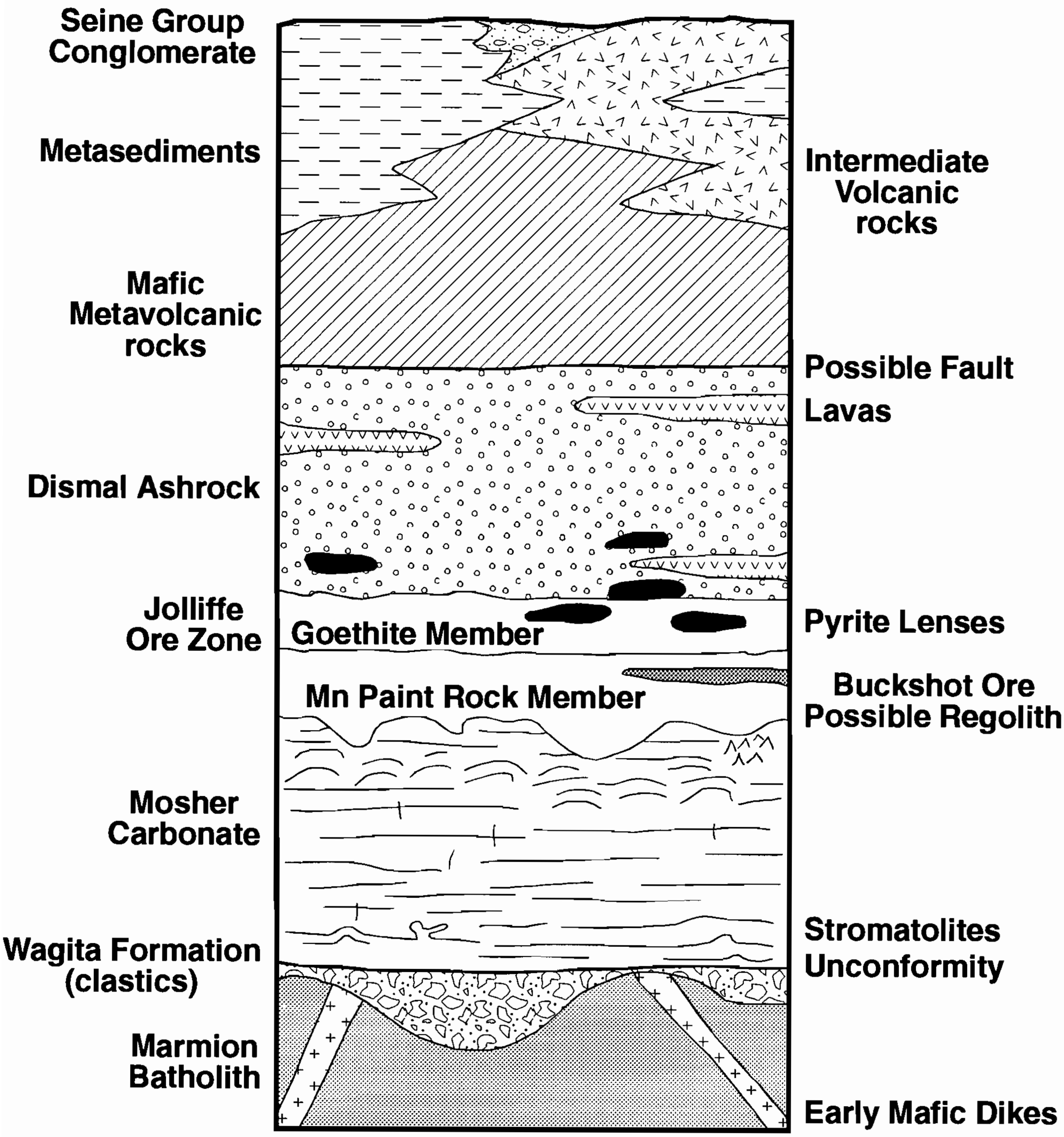


Figure 23. Schematic diagram (after Wilks and Nisbet 1988) showing the Steep Rock Group and contiguous supracrustal rocks of the Steep Rock belt. See legend of Figure 22. The upper ashrock contact is locally faulted and may represent a depositional time-gap.

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

Stop 5-1 The Marmion batholith adjacent to the Steep Rock greenstone belt

Refer to figure 24 for location.

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

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

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

Stop 5-2 Unconformity at the base of the Steep Rock Group

Refer to figure 24 for location.

Outcrops at this locality illustrate the gradational nature of the unconformity at the base of the Steep Rock Group. After walking about 200 m along the torn-up railway line past outcrops of mainly tonalite, you will see a large vertical rock-cut on the east side. Foliated, relatively fine-grained dark metagabbro makes up most of the vertical rock-cut and, at the north end of the vertical rock-cut, is transitional

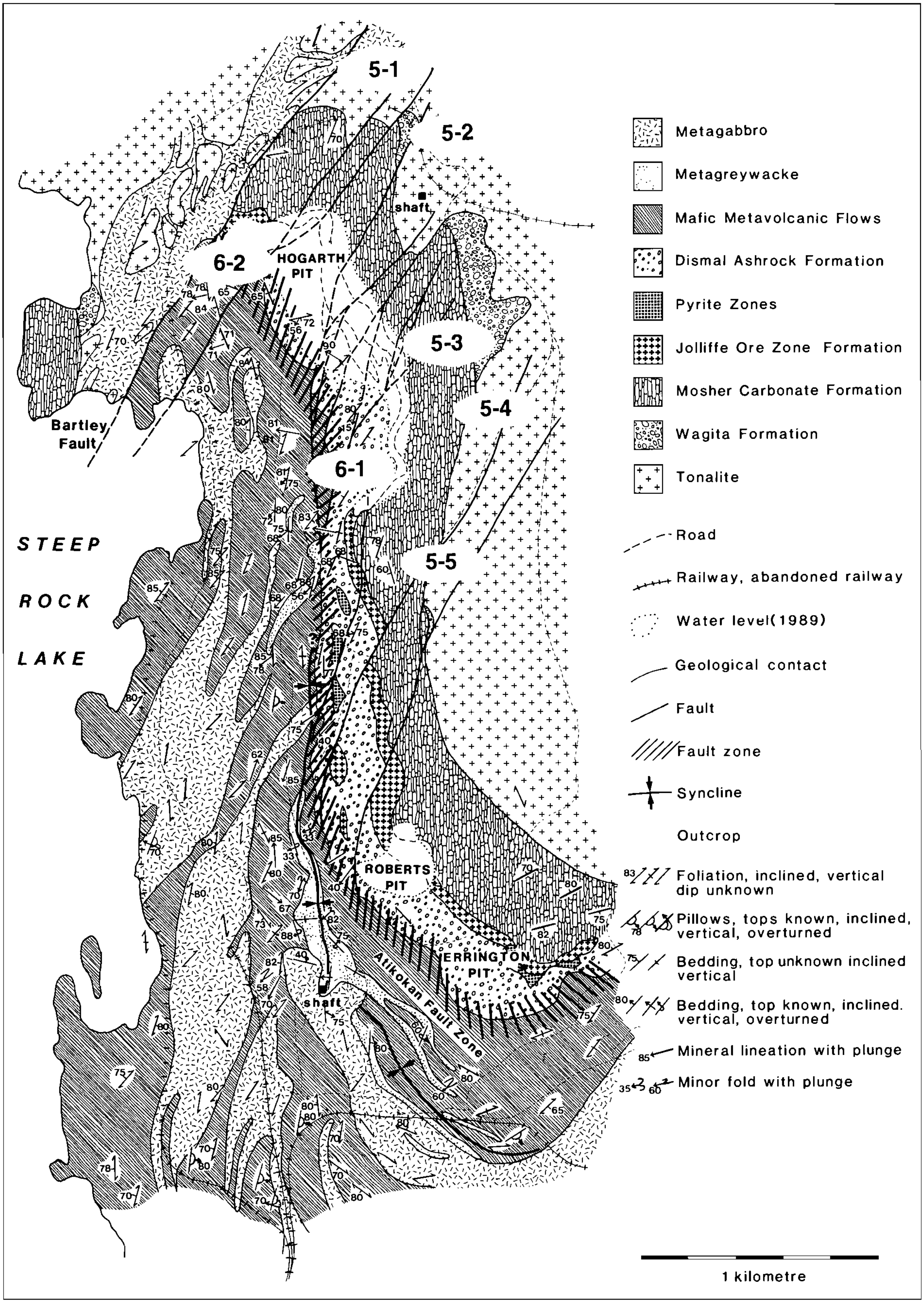


Figure 24. Detailed geologic map of the Steep Rock Mine area showing locations of field-trip stops 5-1 to 5-5, 6-1 and 6-2.

to friable, buff, gritty metasandstone (Wagita Formation). The metasandstone is poorly bedded, but locally shows clasts, and is well foliated, possibly indicating enhanced shearing at the unconformity. The metasandstone can be traced north from the large vertical rock-cut through a series of small exposures to a rounded outcrop of metaconglomerate that lies opposite piles of mine-waste placed on the railway line. Boulders of the conglomerate are mainly tonalite and a dark rock that is probably metagabbro.

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

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

Return to the vehicle and drive 0.9 km south along the gravel perimeter road past remains of the head frame of the Hogarth shaft and a branch road (east) to the former concentrator plant area. Turn right and proceed .2 km down the pit access ramp. Turn right at the first side-road and proceed 0.15 km and park at the outside of a sharp bend overlooking the pit. Walk down the gravel slope following an erosional channel past several rusty, jagged outcrops of tonalite and gabbro. Follow a flagged trail from the base of the gravel slope about 100 m westerly through a wooded area and emerge onto a north-facing slope interspaced with outcrops, boulders and small trees.

The Steep Rock Group *-The Steep Rock Group was extensively studied over the past century (see reviews in Stone et al. 1992) and although early workers offered diverse structural and stratigraphic interpretations, later studies showed the Group to be a homoclinal succession of 4 formations totalling about 1000 m thickness. These include, from the base up, the Wagita Formation (conglomerate, sandstone), Mosher Carbonate Formation (limestone, dolostone), Jolliffe Ore Zone Formation (goethitic iron formation) and Dismal Ashrock Formation (ultramafic pyroclastic rocks) as shown in Figures 23 and 24. The lower boundary of the Steep Rock Group is defined by an unconformity at the base of the Wagita formation. Although sheared locally (e.g. Stops 5-2, 5-3) other exposures of the unconformity (e.g. Stop 5-5) show little deformation and preclude tectonic juxtaposition of the Steep Rock Group onto the Marmion batholith. The upper contact of the Dismal Ashrock Formation is highly deformed in the Steep Rock Mine area and is widely cited as the upper boundary of the Steep Rock Group (e.g. Jolliffe 1955).*

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

Stop 5-3 Marmion tonalite; Wagita Formation; Mosher Carbonate Formation; stromatolites

Refer to figure 24 for location.

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

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

Return to the side-road and drive .1 km up the hill and park in a level place at the observation area. Walk to the west side of the observation area.

Stop 5-4 Scenic lookout and overview of the Steep Rock Group

Refer to figure 24 for location.

The partly flooded Hogarth pit can be seen to the north and the Roberts pit to the extreme south. Looking west, the west arm of Steep Rock Lake can be seen beyond a concrete retaining dam.

This stop provides an overview of the west-facing Steep Rock Group and adjacent, east-facing mafic metavolcanic sequences in the Mine area (Figure 24). The scenic lookout is situated on Marmion tonalite at the base of the Steep Rock Group. The Wagita Formation is not well exposed and can be either absent or else thin at this locality. In either case, the base of the Steep Rock Group extends along the foot of the slope immediately in front of you. Grey outcrops underlying the tree-covered hill 200 to 300 m west, and the rusty jagged bluff to the northwest are Mosher Carbonate Formation. The Jolliffe Ore Zone Formation is almost completely flooded; a few reddish outcrops at the northwest end of the Hogarth pit may be remnants of the ore-zone. Dark grey-greenish outcrops on the west side of the pit near the centre of the mine area are Dismal Ashrock Formation. The sides of large heaps of waste rock on the west side of the mine area also appear to be mainly ashrock. The upper bench road that runs nearly full length of the west side of the mine area approximately marks the contact between Dismal Ashrock Formation and adjacent mafic metavolcanic rocks. East-facing pillow lavas and metagabbro dikes underlie most of the tree-covered ridge beyond the upper bench road. The failed slope at the northwest end of the Hogarth pit is in metagabbro and tonalite.

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

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

Refer to figure 24 for location.

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

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

Since lower levels of the mine are flooded, proceed back up the access ramp and follow the perimeter road returning to Atikokan.

Day 6. Atikokan to Shebandowan

This morning we return to the abandoned Steep Rock Iron Mine by proceeding north on Mercury or O'Brien streets in Atikokan and continue past the Atikokan airport. After passing under the Canadian National Railway overpass, continue about 100 m and then turn right onto a gravel access road. Proceed north through the level, open maintenance yard area and the Errington Shaft and begin to descend into the pit turning left onto the upper (west) bench road.

Proceed north about 1 km on the upper bench road past a large failed cliff of volcanic rocks in the west pit wall. Park and follow the group leader east and down-slope to an outcrop at the top of a high bench.

Stop 6-1 Dismal Ashrock Formation; pyrite lens

Refer to figure 24 for location.

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

Climb back up the slope to the parking area and drive about 0.7 km north on the upper bench road. Turn right and proceed .3 km down a pit-access ramp turning right at an ensuing junction and continue 0.2 km down toward the flooded Hogarth pit. Park by two power poles mounted in cement-filled tires. Walk north along the access road curving up-hill to the west. From the end of the road follow a flagged trail north over heaps of mine waste to the edge of the pit.

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

At the present locality, the boundary between the Dismal Ashrock Formation and contiguous mafic metavolcanic rocks could possibly be an assemblage boundary. Favourable evidence includes the major lithologic and structural transition from the west-facing Steep Rock Group to a thick sequence of east-facing mafic lava flows and gabbro. The boundary is faulted in the mine area (Figure 24; Stop 5-7),which implies that the mafic sequence could have been tectonically juxtaposed with the Steep Rock Group. Since neither the Steep Rock Group or metavolcanic rocks of the Steep Rock belt are dated, the depositional time gap at their mutual contact is unknown.

Stop 6-2 Deformation at the contact between the Dismal Ashrock Formation and mafic metavolcanic rocks

Refer to figure 24 for location. Please note that this stop might be flooded at the time of the trip and an alternative stop might have to be selected.

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

Provided that the slope of waste rock to the north is not flooded, carefully move down the slope. Observe the pale-green to white friable rock that shows good pencil structure in the bench wall to your left. The pale green rock is probably a deformed mafic metavolcanic flow. Moving farther down the slope you pass beside soft, crumbled and crenulated, variably rusty to dark green rock of uncertain type. *Upon reaching the lower bench level, go west along the bench to a point where it is overrun with talus.* The dark-green deformed rocks in the pit wall behind the bench are Dismal Ashrock Formation and metagabbro. A rusty zone, which is possibly a pyrite lense, can be seen high on the pit wall. Several subvertical, north-trending mesoscopic faults cut up through the ashrock and are eroded back into the pit wall.

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

Proceed back up the ramps to the upper bench road and go south on to the paved perimeter road in the direction of Atikokan. Turn right on the road leading to the seaplane base on Steep Rock Lake and continue up a steep hill and around several curves for 0.4 km. Park in a turn-out on the left side of the road and walk east about 50 m through small trees to low outcrops in an abandoned gravel pit.

Stop 6-3 Pillowed mafic metavolcanic flows *(optional)*

Refer to figure 22 for location.

This outcrop lies within the Canadian Chareston Gravel Pit that was strip-mined for iron-ore gravel in 1960 to 1965.
Pillowed mafic metavolcanic flows that underlie the western half of the Steep Rock belt are represented in this outcrop. Pillows typically have a pale-green, homogeneous interior and very-fine-grained, dark-green rims. Several varieties of pillows ranging from large "mattress" to small spherical types occur together. The younging direction derived from pillow shapes in this outcrop and elsewhere in the western Steep Rock belt is east. A massive flow occurs at the extreme west end of the outcrop. The rock is calc-alkaline basalt.

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

Stop 6-4 Quetico metasedimentary rocks *(optional)*

Refer to figure 22 for location.

This outcrop shows several types of bedding and sedimentary features. Thick beds (up to 1 m) of fairly homogeneous metawacke occur at the north side of the outcrop. Bed thickness is generally .05 to .2 m; many beds are graded from a coarse sandy base to a grey, laminated, silty top. Truncated beds and crescentic scour channels filled with coarse material are present. Clasts (up to 0.2 m) of fine-grained, white chert or possibly felsic volcanic material fill some scoured channels. Grainsize gradation and scour channels indicate that the local younging direction is south. The outcrop is cut by quartz veins, fractures and small-displacement faults. The mafic mineral is mainly biotite. Metamorphic grade increases south with the appearance of amphibole and garnet about 1 km distant.

Return to Highway 11 and proceed to the Shebandowan greenstone belt.

From the town of Atikokan, proceed south to the junction of Highway 11; turn eastward (left) onto Highway 11 and drive for approximately 81 km where an *optional* stop can be made at the top of the hill marked by the presence of several microwave towers and a long low roadcut. **Please be aware of the traffic along the highway at all stops during the day!** We can use this opportunity to briefly examine the Quetico metasedimentary gneiss and comment on the structural styles across this metasedimentary subprovince. Alternatively, simply note as you are driving from this point eastward, that there is a decreasing proportion of anatectic migmatite and a clearer definition of greywacke beds as we approach the Shebandowan greenstone belt.

The second half of this day will focus on outcrops in the western part of the Shebandowan greenstone belt. We will spend the afternoon looking at several representative rock units, mineral occurrences and structures in the Shebandowan belt. In relation to the theme of this field trip, we will discuss the regional significance of rock associations and structures observed in the Shebandowan belt and set the geology of this part of the Wawa Subprovince into the broader framework of the tectonomagmatic evolution across the western Superior Province.

Stops 6-5 to 6-11 are shown in relation to the western half of the greenstone belt in Figure 25. Stops 7-1 to 7-7 are located in Figure 26.

Stop 6-5 Quetico metasedimentary belt near the transition from diatexitic granites to metatexites with minor partial melt present *(Optional)*

The outcrops show a typical metatexite dominated by biotite-quartz-plagioclase feldspar greywacke with interlayers of grey granite leucosome bordered with biotite selvages. Bedding is preserved but top determinations are not always reliable in this area. As we approach the Shebandowan greenstone belt, the graded bedding is better preserved and has been used by Sawyer (1983) to interpret the presence of large scale stratigraphic inversions. He showed the Quetico Subprovince is characterized locally by overturned early folds that were refolded about shallowly plunging upright folds. In this region these later folds plunge typically towards the east. The extension fabric parallel to the fold axes can be traced as shallowly-plunging mineral lineations across the Quetico boundary into the northern half of the Shebandowan greenstone belt.

Proceed eastward along Highway 11 for 4.9 km to a basaltic outcrop (Stop 6-6) on the north side of the highway.

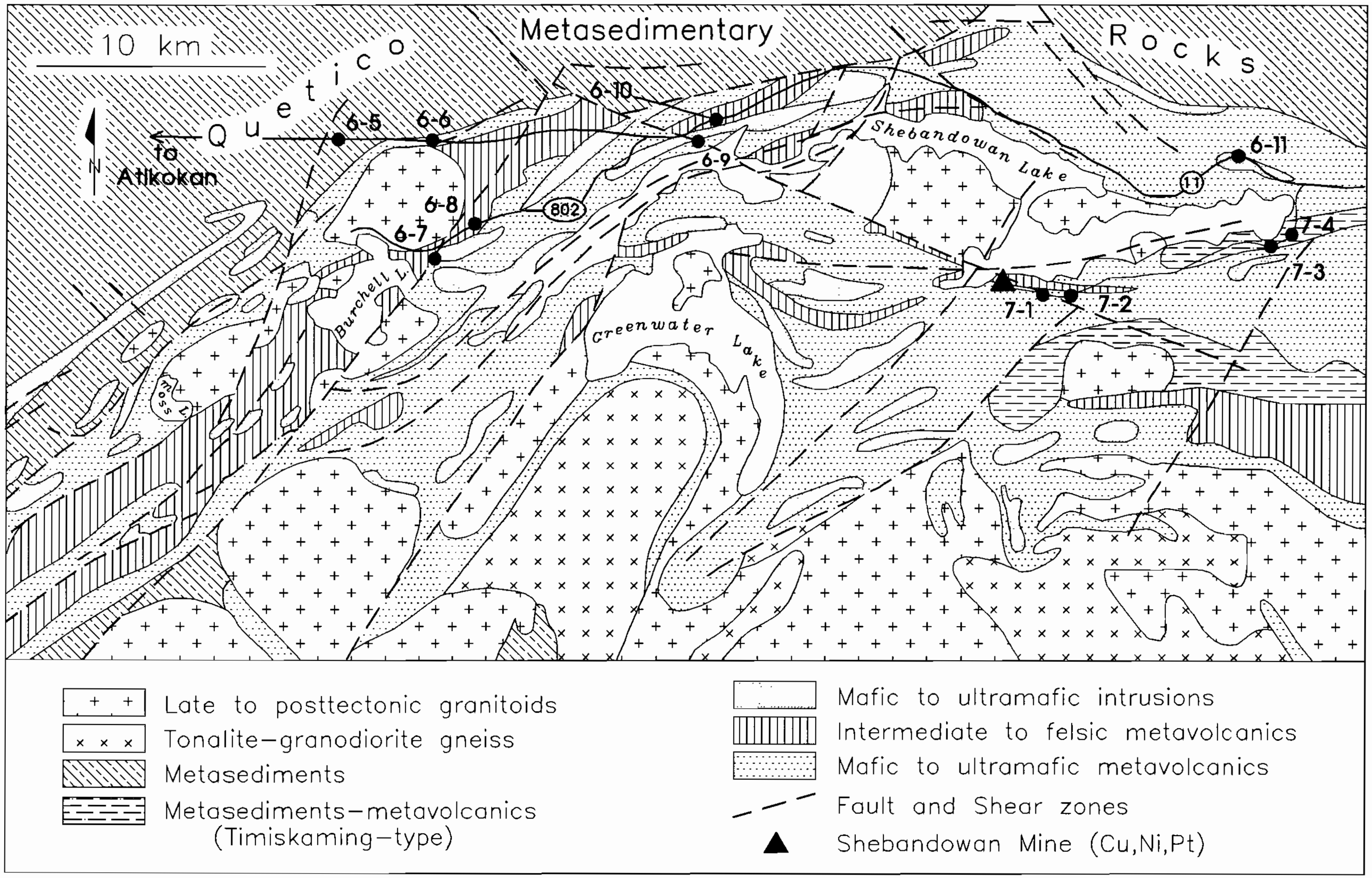


Figure 25. General geology of the western half of the Shebandowan greenstone belt showing stops 6-5 to 6-11 and 7-1 to 7-4.

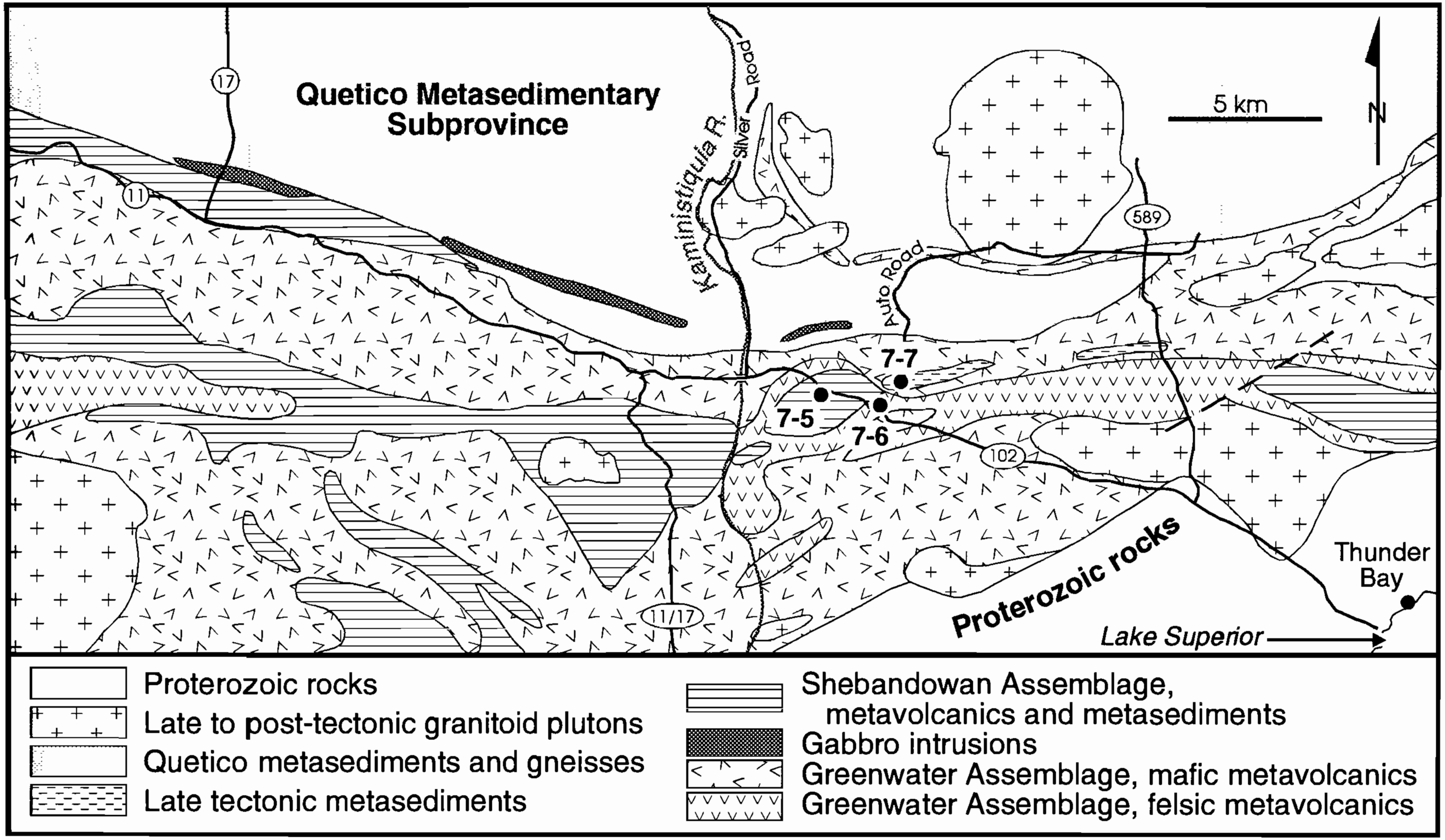


Figure 26. General geology of the eastern half of the Shebandowan greenstone belt showing stops 7-5 to 7-7.

***The Shebandowan greenstone belt** (see Williams et al., 1991) is dominantly composed of two ages (Corfu and Stott, 1986, 1995) of volcanic rocks: 1) a tectonostratigraphic sequence, the Greenwater assemblage, of circa 2720 Ma tholeiitic basalt to rhyolite across the breadth of the belt, and including komatiitic flows and sills in the southern half of the belt; 2) a younger sequence - the Shebandowan assemblage - of circa 2690 Ma calc-alkalic to shoshonitic autobrecciated lavas and pyroclastic rocks, typically reddish to olive-green coloured, and a suite of wacke, siltstone, mudstone, conglomerate and banded iron formation that is locally interbedded with, but typically overlies the volcanics.*

1) The bulk of the Shebandowan greenstone belt is marked by an apparent "cyclicity" of volcanism dominated by subaqueous tholeiitic to calc-alkaline basaltic flows and basaltic komatiite with each "cycle" capped by calc-alkalic to locally tholeiitic, intermediate to felsic volcanics (Osmani, 1996). North of the Shebandowan lakes, stratigraphic top directions generally face northward in contrast with volcanic strata of similar age, south of these lakes where top directions mainly face southward. At one stage we thought that an apparently greater volume of ultramafic rocks in the southern half of the belt, east of Greenwater Lake, plus oppositely facing stratigraphy between north and south halves of the belt suggested the possiblity of two major assemblages as shown on the tectonic assemblages map of Ontario (OGS 1992), underlying the younger Shebandowan assemblage. However, subsequent U-Pb zircon age determinations (Corfu and Stott 1995) coupled with earlier determinations (Corfu and Stott 1986) shown in Figure 27, now suggest a different interpretation: the markedly similar 2718-2721 Ma age of most intermediate to felsic volcanic units and layered gabbroic to anorthositic synvolcanic intrusions suggests that part of this "cyclicity" is attributable to tectonic interleaving and repetition of strata. The implication is that the Shebandowan belt is probably a fold-thrust belt formed during the Kenoran orogeny.

2) The tectonic setting of the distinctive rocks of the Shebandowan assemblage remains a subject of discussion; shoshonites in modern settings typically form in immature oceanic island arcs, or within continental crust associated with wrench or extensional faults during and after plate collision. Geophysical and age evidence suggest an intra-continental setting for the Shebandowan assemblage, but the evidence of extensional fault boundaries to this assemblage is arguable. However, linear aeromagnetic anomalies (notably of iron formation and ultramafic units) of the 2720 Ma strata can be followed for many kilometres and locally traced under the overlying, 2690 Ma Shebandowan assemblage rocks. The general trend of the Shebandowan assemblage units is also locally oblique to the trend of the circa 2720 Ma strata. Further descriptions and discussion on the tectonic setting of the volcanic sequences in this belt will be entertained on the field trip.

Felsic plutons within and south of the Shebandowan belt are generally late to post-tectonic, approximately 2680-2684 Ma; south of the belt, tonalite to granodiorite gneiss (the Northern Light Gneiss Complex) varies in age from at least 2750 to 2707 Ma (Figure 28). Within the belt, minor quartz porphyry intrusions, like the elliptical Shebandowan pluton (max. 2692 Ma), represent magmatic activity similar in age to some felsic volcanism in the northern part of the belt dated at about 2695 Ma (Figure 28); this volcanism is presently considered the oldest component of he Shebandowan assemblage and occurs locally along Highway 11.

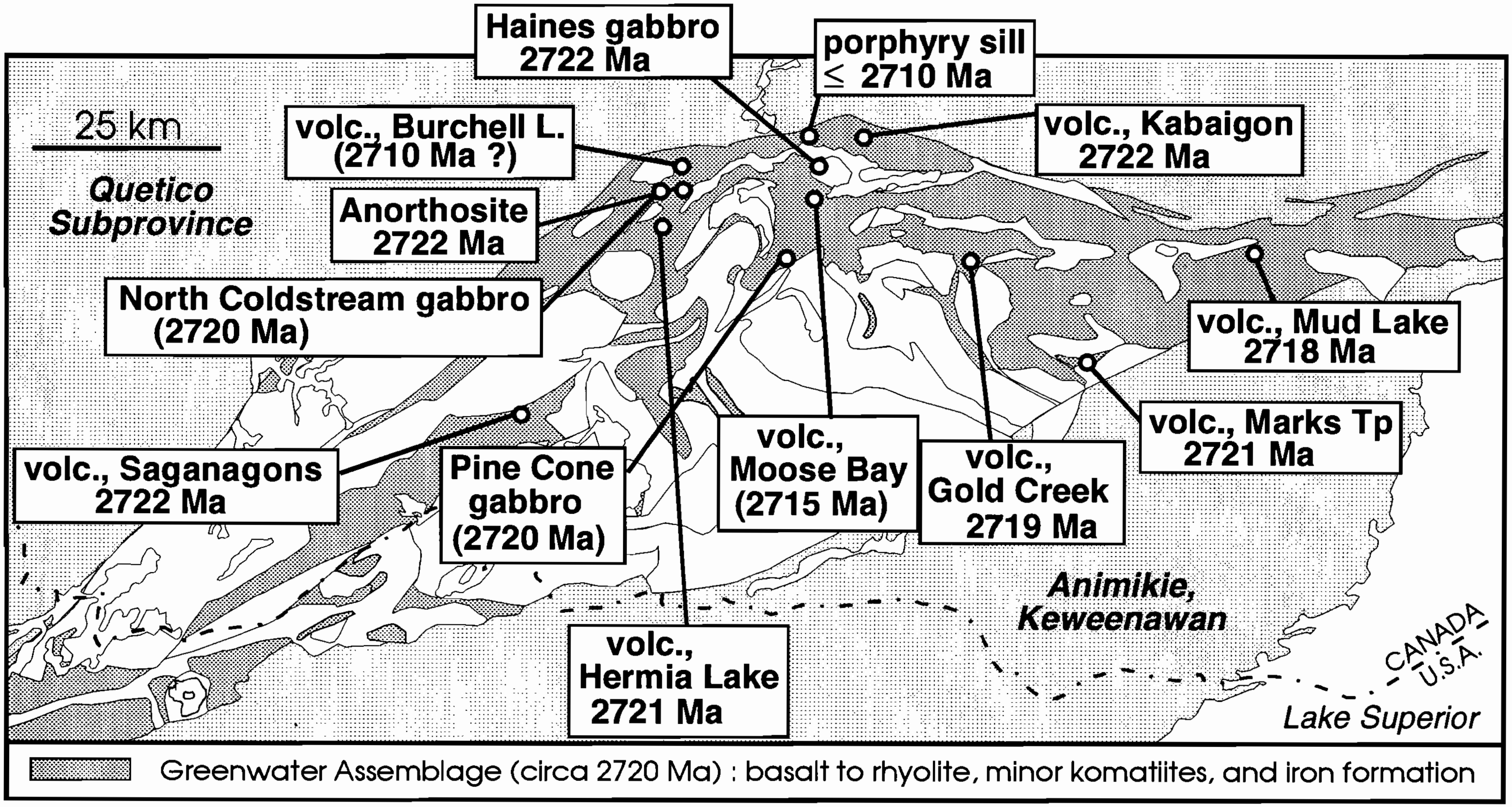


Figure 27. U-Pb zircon ages of felsic volcanics and gabbroic intrusions in the Greenwater assemblage (F. Corfu, Royal Ontario Museum).

Tectonic Structures- *The schistosity of volcanic and sedimentary rocks in the belt is typically parallel to bedding and dips subvertically to steeply northward. The most intensely developed schistosity and shearing is found in the northern half of the belt. Overprinting fabric relationships are typically not well displayed except locally, for example, where the long axis of basaltic pillows (parallel to the regional D_1 lineation) is at a high angle to the penetrative mineral lineation (parallel to the regional D_2 lineation). Detailed studies of local deformation phases have received only limited attention; e.g., Borradaile et al., 1994. The western half of the belt has been subdivided into two regional deformation domains (Stott and Schnieders, 1983; Williams et al., 1991) as described below. At the time of the initial structural study of the west half of the belt by Stott (1986), the belt was viewed as a likely illustration of significant transpressive deformation along the northern half of the belt, developed during Archean lithospheric plate collision between the Wabigoon and Wawa subprovinces (or superterranes).*

D_1 and D_2 Deformation Domains- *The western half of the Shebandowan greenstone belt can be divided into two distinctive structural domains, D_1 and D_2 (Figure 29). These domains are characterized by contrasts to some degree in strain intensity and more uniquely by differences in the plunge azimuth of the stretching lineations. The D_1 domain possesses little evidence of the shearing so typical of the D_2 domain. In contrast, most of the gold mineralization in the greenstone belt is restricted to the D_2 domain where numerous gold-quartz veins are concentrated in shear zones. The D_2 domain can be attributed to northwest-southeast directed oblique convergence of Archean crustal plates. Such collision could have produced zones of transpression such as the northern half of the Shebandowan greenstone belt, where both crustal shortening and simple shearing are evident. The relationship if any of the D_1 domain to the D_2 transpression event is not clear. Domains of transpression have also been observed in the northern part of the Vermilion greenstone belt (Hudleston et al., 1988), an extension of the Shebandowan belt in Minnesota, and along the northern margin of the Wabigoon Subprovince, particularly near Kenora (Sanborn-Barrie, 1991).*

The boundary between the Quetico metasediments and the Shebandowan belt is typically marked by faulting (e.g., Osmani, 1996) and shows subvertical displacement locally. The boundary is an interface of dislocation within a broad zone of transpression represented by the D_2 fabrics in both the northern part of the Shebandowan belt and the upright, shallowly-east plunging folds of the Quetico metasedimentary rocks and gneisses. This transpression zone is typical of the boundaries between metasedimentary and metavolcanic subprovinces. Similar structures are to be seen straddling the Wabigoon-Winnipeg River and Winnipeg River-English River subprovince boundaries.

A summary of the geochronologic sequence of events, determined largely from preliminary age determinations by F. Corfu (work continuing in progress), and from the regional structural patterns of G. Stott (Williams et al., 1991), is shown in Figure 30 and the general tectonic evolution of this region, is schematically represented in Figure 31.

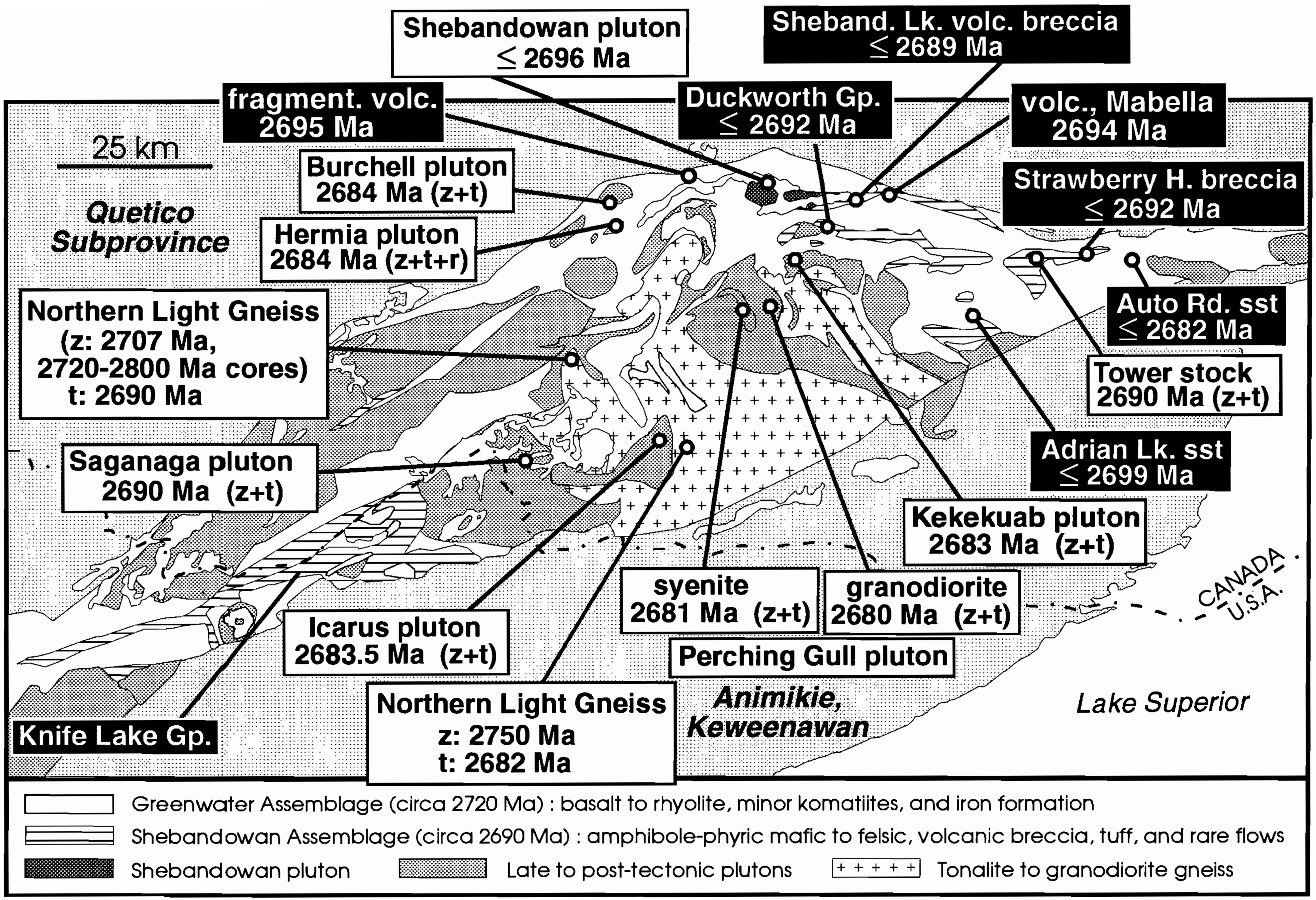


Figure 28. U-Pb zircon ages of volcanic and detrital sedimentary zircons from the Shebandowan assemblage, and zircon and titanite ages from felsic plutons and gneisses (F. Corfu, R.O.M.).

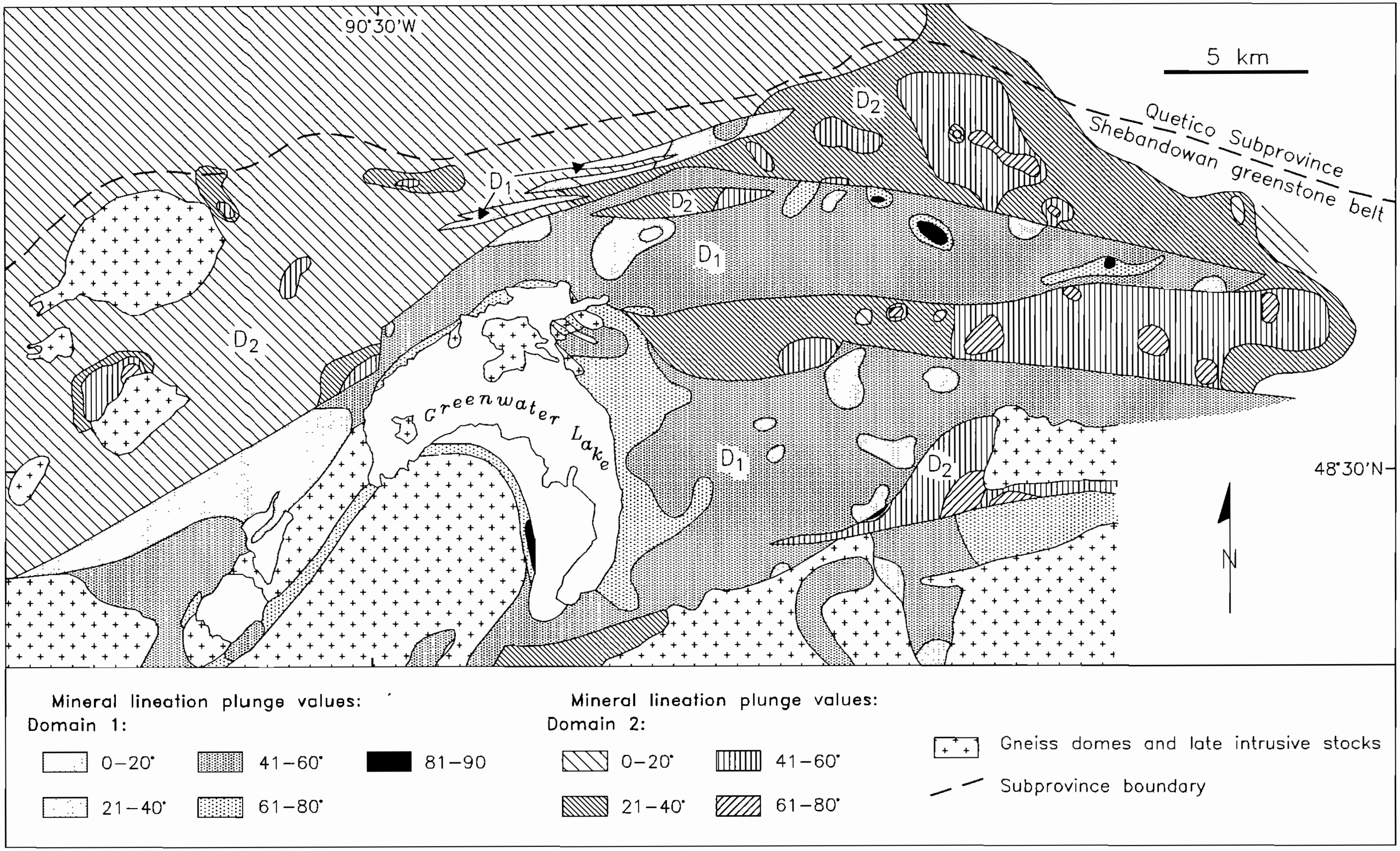


Figure 29. Structural domains D1 and D2 in the western part of the Shebandowan greenstone belt, characterized by contrasts in plunge directions of mineral and shape lineations. Contoured values of plunges of stretching lineations show lineations plunge westward in D1 domains and eastward in D2 domains. The shallowest plunge values occur in the northwestern part of the belt and into the Quetico Subprovince. Lineations steepen within the contact strain aureoles of late plutons (modified from Williams et al. 1991).

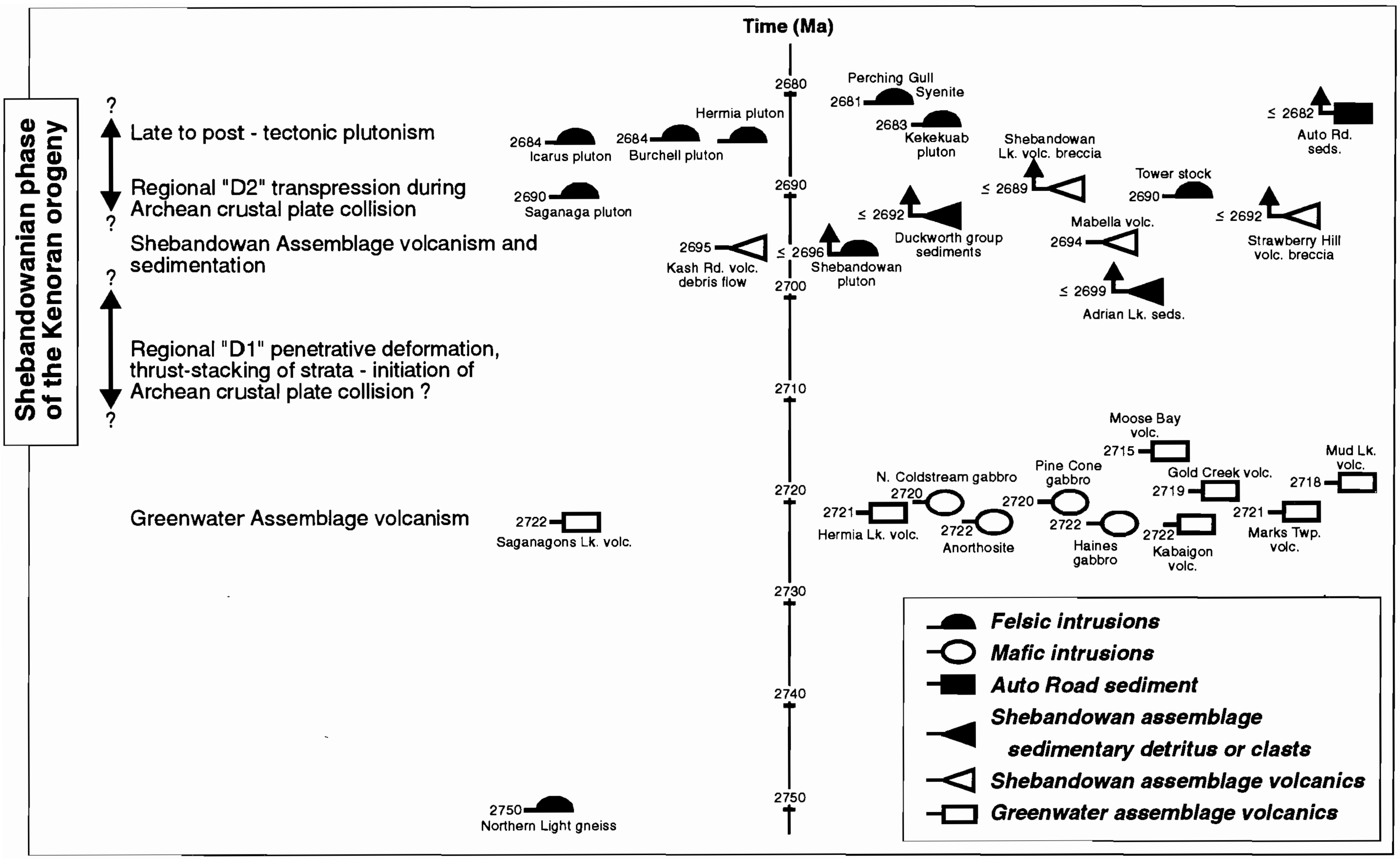


Figure 30. Sequence of regional structural and geochronologic events in the history of the Shebandowan greenstone belt and the Northern Light - Perching Gull plutonic complex to the south, based on Stott (1986), Corfu and Stott (1986, 1995).

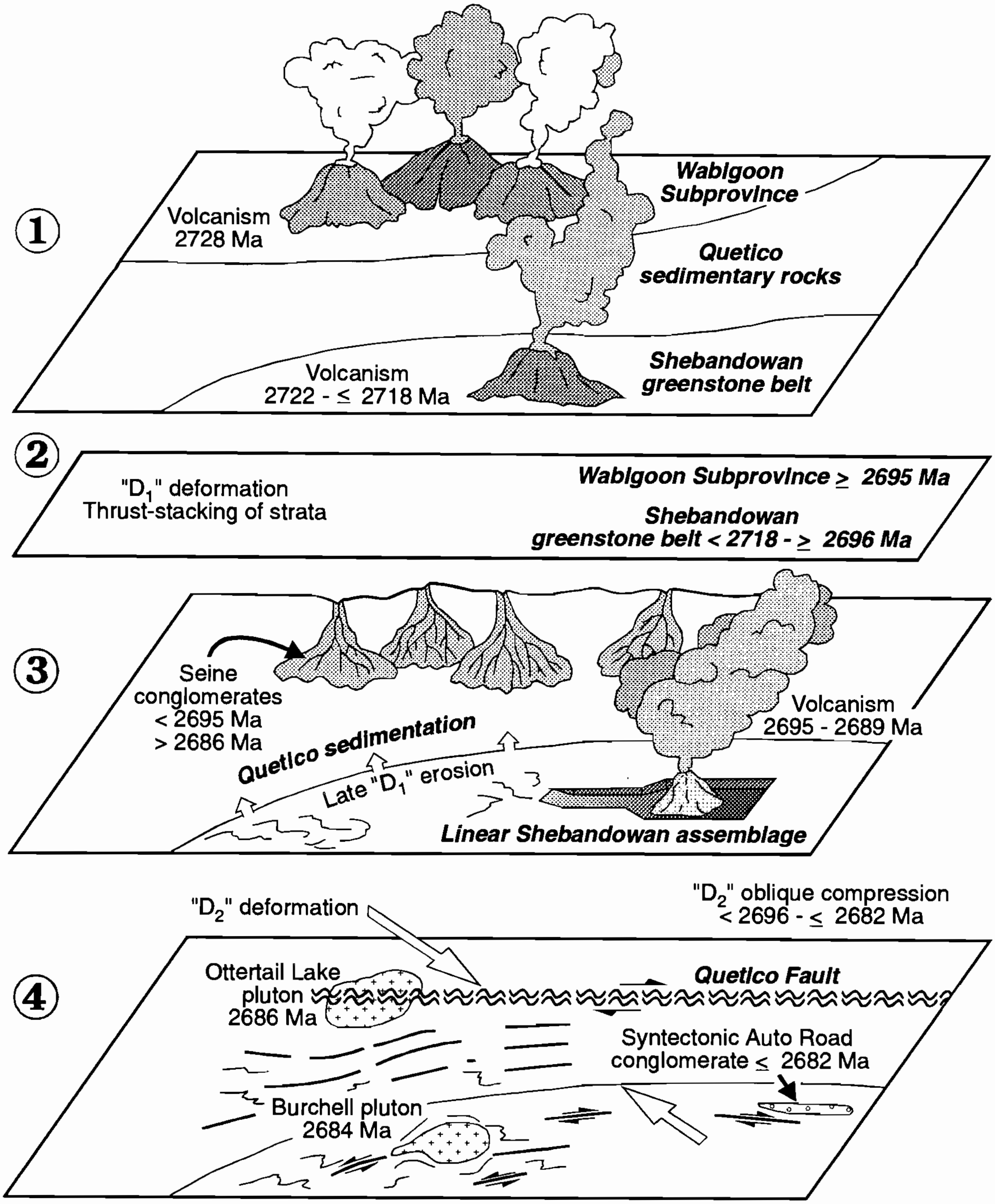


Figure 31. A 4-stage sequence of events across the Wabigoon-Quetico-Wawa subprovince boundaries. Early thrusting and penetrative deformation (Regional D1 of Shebandowan) follow deposition of the Greenwater assemblage volcanics and minor sedimentary units. Plate collision between the Wabigoon and Wawa subprovinces or superterranes is probably the cause of D2 transpressive deformation apparent from the Quetico Fault, upright, shallowly plunging folds across the Quetico subprovince, and comparable structures of the D2 domain of the Shebandowan belt. Prior to and concurrent with regional D2 formation is the deposition of the Shebandowan assemblage units unconformably on the Greenwater assemblage. Auto Road conglomerate sedimentation ≤ 2682 Ma, concurrent with late-stage D2 deformation as well as emplacement of late to post-tectonic plutons, unaffected by D2, in the southern part of the belt. (Modified and updated from Williams et al., 1991).

Gold Metallogeny-..*Gold in the Shebandowan belt is concentrated in three structural environments: 1) within altered shear zones in the D_2 structural domain dominating the northern half of the belt (Stott and Schnieders; 2) within the contact strain aureole of the late to post-tectonic 2683 Ma Kekekuab pluton at the southern margin of the belt, where the Gold Creek occurrences are found; 3) and as gold-pyrite mineralization within late tectonic hornblende lamprophyre-gabbro-diorite-syenite stocks, such as the 2690 Ma Tower stock Figure 28, (Carter, 1992) and dikes (e.g., dikes within the Duckworth group of the Shebandowan assemblage), possibly derived from a metasomatized mantle wedge (Carter 1993).*

Base Metal Metallogeny- *The west-central part of the belt contains the highest concentration of base metal mineralization in the Shebandowan greenstone belt. One producing (Shebandowan Ni-Cu-PGE) and one past producing (North Coldstream Cu) mines are located within this part of the belt. The various styles of base metal mineralization include stratabound and/or VMS Cu (±Zn±Ag±Au) mineralization, intrusion-hosted, late-magmatic Cu (±Ni±PGE±Ag±Au) mineralization, and shear/fault hosted Cu-rich sulphide mineralization related to late, possibly D_3, deformation in the belt. The Shebandowan lakes area is characterized by the repetition of base metal mineralized metavolcanic stratigraphy that includes tholeiitic mafic metavolcanic rocks and locally FIII-like felsic metavolcanic rocks. Calc-alkaline rocks are also present. Stratigraphic repetition is the result of thrust stacking or tight, isoclinal folding concentrated along the Wawa-Quetico subprovince boundary during convergence.*

Numerous gabbroic bodies, the largest being the Haines gabbroic complex, occur throughout the belt. Field relationships and recent U-Pb dating of both the Haines gabbroic complex and host volcanic rocks have shown them to be of similar age (approximately 2720 Ma). Thus, the Haines complex and related gabbros represent synvolcanic intrusions that provided the heat necessary to drive the base metal mineralizing hydrothermal fluids. In addition, the Haines gabbroic complex hosts late magmatic Cu (±Ni±PGE) mineralization related to pegmatoid and chlorite+amphibole+magnetite alteration development. Late shearing has locally been superimposed on these zones. The origin of the North Coldstream Mine base metal sulphide mineralization and associated alteration and replacement is in question, but alteration of the host gabbro is partly represented by chlorite+amphibole+magnetite schist, similar to that observed in the Haines complex. This suggests that late magmatic, hydrothermal activity was important in the development of the North Coldstream orebody and may have been focused at the intersection of long-lived regional structures.

Stop 6-6 Pillowed basalt *(Optional)*

This outcrop is not far from the Quetico-Shebandowan greenstone belt boundary and illustrates the reasonably good preservation of primary volcanic features in the rock. The stratigraphic top to north can be confirmed from the pillow shapes. Bedding orientation is evident from parallel layers of collapsed lava channels. Note that the dominant mineral lineation accentuated by fine-grained amphibole crystals is parallel to the long axis of varioles and plunges eastward. This lineation orientation is parallel to that of the D_2 lineations and fold axes of metasedimentary rocks nearby in the Quetico Subprovince (Sawyer 1983). This orientation of mineral lineations characterizes the northern half of the Shebandowan greenstone belt (Figure 29).

Continue eastward on Highway 11 for 7.5 km and turn right (south) onto a rough paved road (Highway 802 south). Drive 11 km to a road leading to the entrance to the abandoned North Coldstream Mine. A gate is located 200 m up this mine road. Note that this is private property and permission is required.

Stop 6-7 The North Coldstream Mine

The North Coldstream Cu Mine operated intermittently between 1906 and 1967, and produced 102 million pounds of Cu, 440 000 ounces of Ag and 22 000 ounces of Au from 2.7 million tons of ore (Osmani 1993). Accessibility to underground mine workings has not been possible since the mine closed in the late 1960's. However, excellent exposure exists on surface in the vicinity of the headframe and mill buildings.

The Cu-rich mineralization (Figure 32) is hosted by a highly silicified part of the North Coldstream gabbro near the intersection of two major regional structures, the Knife Lake-Burchell Lake fault and the North Coldstream Mine fault. The progression of silicification, from relatively weakly altered gabbro to the development of "silicalite", a quartz dominated rock with between 82 and 97 wt.% SiO_2, is displayed in the surface exposures at the minesite. With increasing alteration intensity, the gabbro progresses into a chlorite + amphibole + magnetite + carbonate schist with increasing quartz content. The quartz occurs as centimetre-long streaks and rounded aggregates, until it composes greater than 80% of the rock volume. The contact between the schist and the silicalite is commonly marked by a highly strained quartz + sericite ± andalusite ± chloritoid schist that was the focus for deformation at the margin of the more competant silicalite due to late movement on the North Coldstream Mine fault. Brown, white, blue, purple and dark grey varieties of silicalite have been identified, with each colour variation corresponding to trace mineral variations. Trace minerals in the silicalite include magnetite, ilmenite, rutile, leucoxene, titanite and green, Cr-, V-, and Fe-bearing mica. Chalcopyrite and pyrite occur as disseminations, stringers, veins, and semi-massive to massive pods.

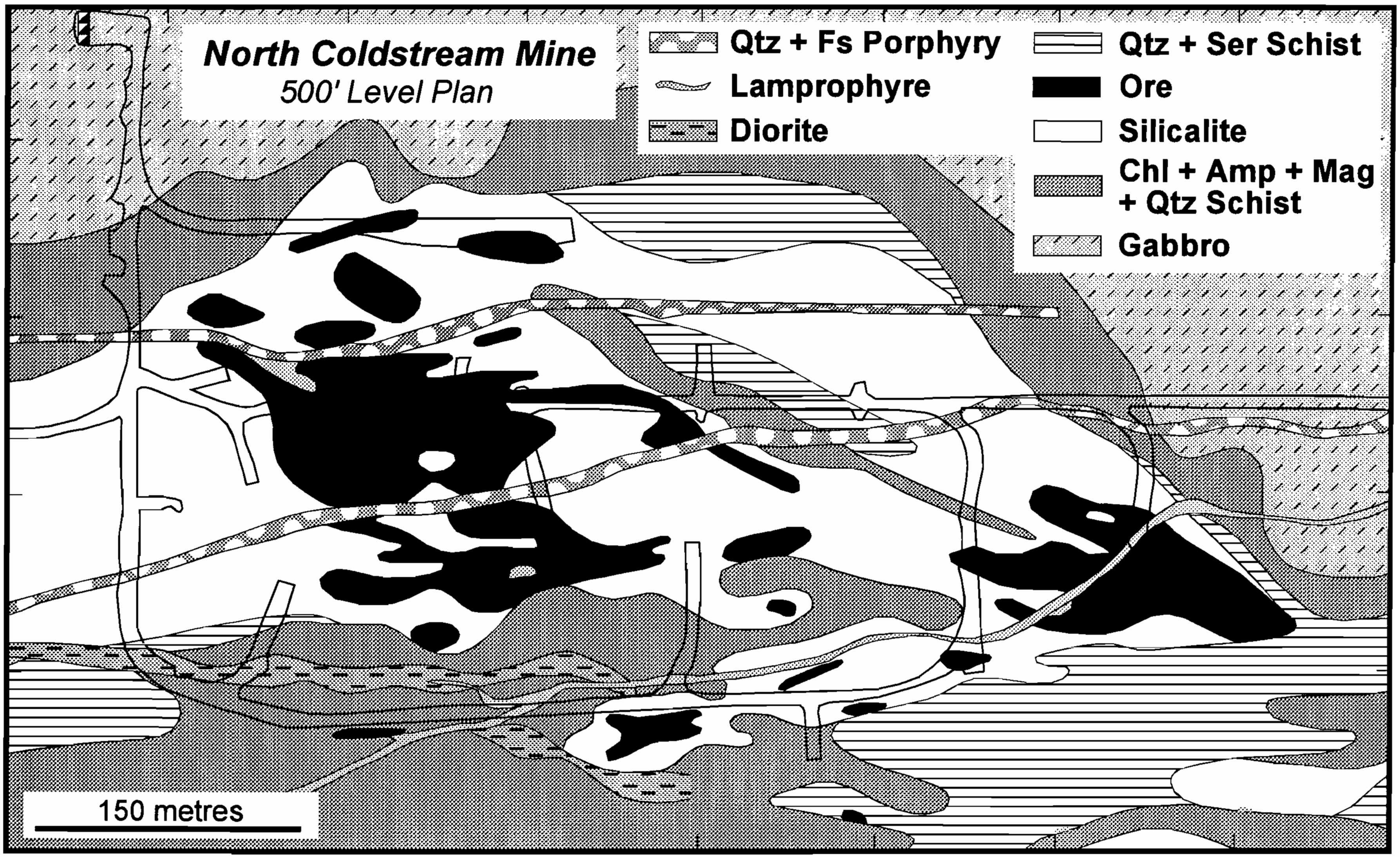


Figure 32. Level plan of the North Coldstream mine, Stop 6-7.

Return eastward along Highway 802 for about 1.6 to 2.4 km to one of several small outcrops on the north side of the road illustrating the intense fissile character of the felsic volcanics in this part of the belt.

Stop 6-8 Highly Fissile Felsic Volcanic Rocks

This outcrop of felsic volcanics illustrates the markedly schistose character of rocks widely observed particularly in the northern half of this greenstone belt. It characterizes the D_2 domain of deformation outlined in Figure 29. The most intensely developed schistosity in the belt can be seen in this area, typically concentrated in broad zones. This location near the Burchell Lake pluton is at the axis of the regional "bend" in the trend of the greenstone belt. Note the relatively shallow plunge of lineations typical of the D_2 deformation domain in this area.

Continue eastward back to Highway 11. Turn right and drive on Highway 11 for 1.8 km. Turn right onto a gravel road (directly across the highway from the Kashabowie Road (Hwy 802 North). Continue for 0.5 km down this road. The stop, which is a large cleared outcrop, is only a short distance east of the road. This is private property and permission is required. Inquire at the Resident Geologist's office in Thunder Bay for current ownership.

Stop 6-9 The Vanguard East Cu-Zn-Au-Ag Prospect.

This massive sulphide occurrence is well exposed and illustrates some of the structural and alteration complexity associated with the 2720 Ma volcanic rocks in the northern half of this greenstone belt. Similar stratigraphic and volcanological relationships are evident 10 km southwest of this stop, at the East Coldstream base metal showing. It is possible that packages of rock with VMS potential may be traced for many kilometres, facilitating regional stratigraphic and structural reconstruction. We speculate that the Vanguard East and West prospects are two of a number of 2720 Ma VMS deposits and occurrences, scattered along the length of the Wawa Subprovince, close to its northern margin; these deposits include the Winston Lake Mine north of Lake Superior and the numerous deposits now mined out at Manitouwadge, 260 km northeast of Thunder Bay.

The Vanguard East and West prospects, 1500 m apart, together comprise 300 000 tons of ore (1.2% Cu, 0.02 ounce Au per ton and appreciable Zn and Ag). A sequence of altered mafic to locally intermediate metavolcanic rocks envelopes the base metal mineralized horizons, accompanied by hyaloclastite, flow-top or debris flow breccias, intermediate tuff, and silica- and siderite-rich chemical sedimentary or replacement horizons (Figure 33). Devitrification processes are evident from the presence of varioles up to 3 cm in diameter. Alteration has transformed the mafic metavolcanic flows into quartz + sericite ± chlorite ± ferroan dolomite ±

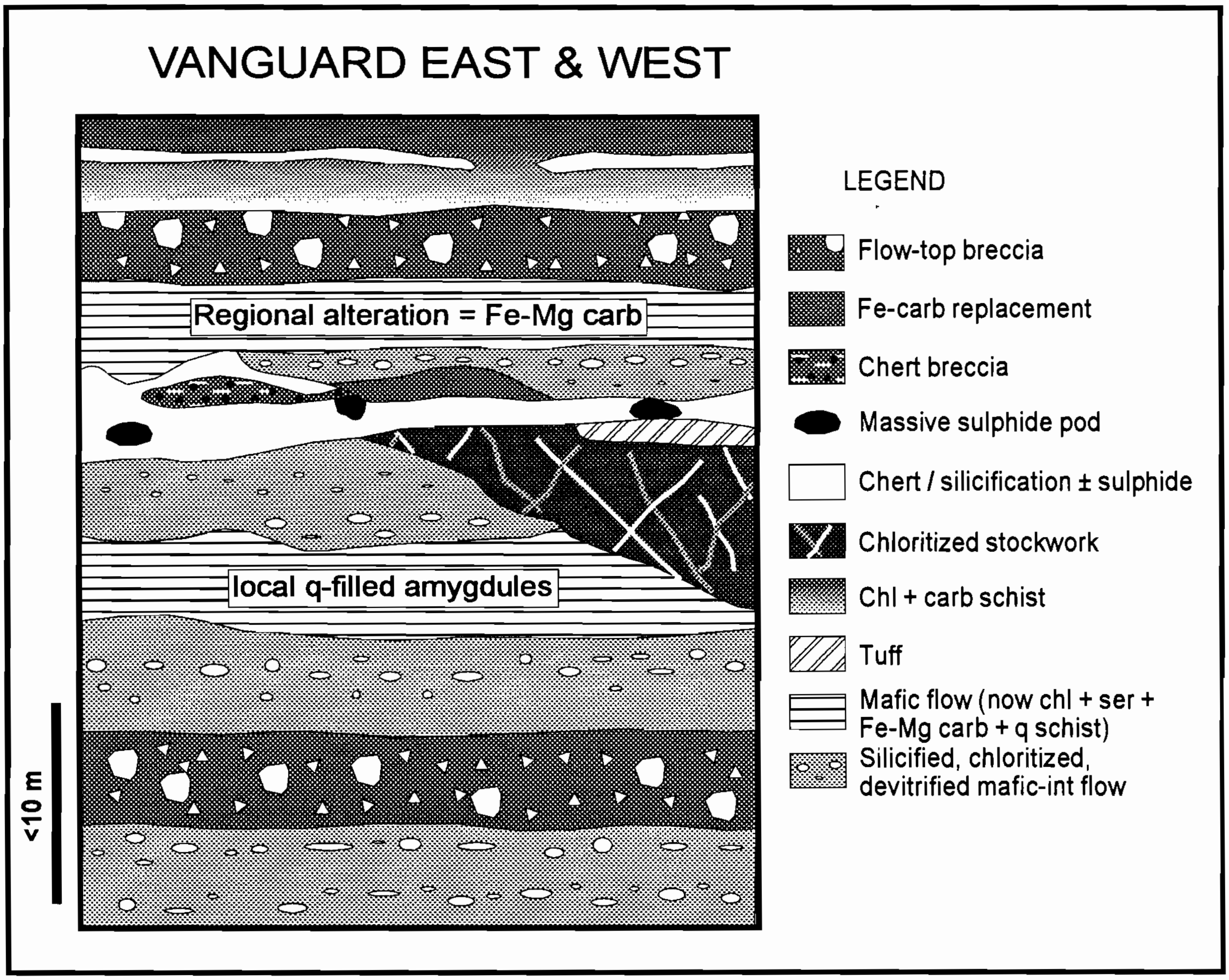


Figure 33. Schematic stratigraphic section of the Vanguard area, Stop 6-9.

chloritoid schists. Several generations of alteration are identified: silicification and chloritization, associated with base metal sulphide mineralization, is followed by syn to post-D_2 formation of sericite, ferroan dolomite and chloritoid. The dominant (D_2) foliation is subparallel to lithologic contacts and dips moderately to steeply northward, as is typical for most of the strata in this greenstone belt. This part of the greenstone belt is tightly folded and stratigraphic top directions change across strike. This outcrop is on one limb of a fold; pillows and graded bedding in the vicinity show tops to the south. The sulphide mineralization is concentrated in chert-silicification horizons, chloritized stringer stockworks, and massive pods up to 50 cm in diameter. D_3 deformation is represented by northwest-southeast and conjugate east-west sets of shears and fractures and kink folds.

Return to Highway 11, turn right and continue eastward for 2.1 km to a low roadcut at the Kash River Road entrance, which leads to the Kashabowie River Resort.

Please be wary of the traffic along the highway.

Stop 6-10 Felsic Debris Flow and Feldspar Porphyry Intrusion

This exposure of felsic debris flow can be observed on both sides of the highway. The unit resembles a lapilli tuff and contains fragments of white and black chert and minor mafic clasts with fragments of the dominant quartz phyric rhyolite. A feldspar porphyry dike obliquely cross cuts the debris flow on the north side of the highway. Although the intrusion is less intensely deformed, both units in this outcrop typify the high degree of flattening and penetrative shear that characterizes much of the region north of the Shebandowan lakes. This deformation produced a stretching lineation that plunges eastward and typifies the "D_2 domain" of Figure 29. A U-Pb zircon age of 2695 Ma was obtained from the quartz phyric fragmental and corresponds in age with the 2696 Ma Shebandowan pluton, centred on Lower Shebandowan Lake.

Continue east along Highway 11 for 26.1 km; as you are driving, you can take note of the fissility of rocks along the highway. The strain intensity reflected in the steeply dipping schistosity varies considerably as we proceed to the next stop. Stronger deformation and evidence of shearing is more widely observed in the western part of the belt, east and south of Burchell Lake. Cross over a railway track and park off the north side of the highway beside a flat outcrop of pillowed basalt. This is a brief optional stop to illustrate the difference in deformation intensity along the length of the D_2 domain in the northern part of the belt.

***Stop 6-11 Pillowed basalt with chlorite-filled vesicles** (Optional)*

This outcrop is an exceptional example for this belt of well-preserved, large mattress-type pillows, some with double cusps that provide good indication of stratigraphic top to the north. The rock contains flattened chlorite-filled, ellipsoidal vesicles that display a pronounced eastward plunge of the long axis of the vesicular ellipse in the foliation plane. The intensity of D_2 deformation of Stott and Schnieders (1983) or D_1 of Borradaile et al. (1994) observed along Highway 11 varies considerably.

Day 7 Lower Shebandowan Lake Area and Highway 102

Drive to the junction of Highway 11 and Shebandowan Road. Turn south onto the road and proceed all the way along this paved road to the Shebandowan mine gate and turn around. We will examine several outcrops along this road before proceeding to Highway 102. This morning we will look at the less intensely deformed basaltic rocks in the D_1 deformation domain, with faintly observable westward plunging lineations, and then cross into the younger Shebandowan assemblage where fabric typical of the D_2 domain is evident.

From the mine gate, proceed back east for 1.5 km to a long roadcut. Stop near the east end. The best features are visible on the north side of the road.

Stop 7-1 Pillowed to Massive Basalt With Hot-Water Discharge Zones (Hydrobreccia)

This outcrop typifies the rocks of the D_1 domain of Figure 29. The D_1 fabric, characterized by westerly plunging stretching lineations is apparent on close inspection of the shapes of small varioles. The lineations are weakly developed and a challenge to identify. These varioles appear to have been formed by local rock-water interaction along synvolcanic hot water discharge zones. By standing on the south side of the road and looking north one can observe the larger features in this cross section. Subvertical dark green chloritic vein-like structures can be seen transecting the lighter green-grey basalt. These vein-like structures are interpreted as the hot water discharge zones. A good illustration of these zones can be seen near the east end of the outcrop on the north side. These chloritic zones display evidence of synvolcanic fracturing of the host rock, intense alteration within the zones and finely concentric rock-water interaction rims of devitrification varioles that are locally fragmented. At the east end of the outcrop, a zone of pillows is flanked on both sides by highly fractured pillow breccias.

Continue east for another 1.5 km to a small basaltic outcrop on the south side of the road.

Stop 7-2 Pillow Breccia and Hyaloclastite Matrix

This small outcrop is a brief stop to illustrate on a flat surface the relatively weak tectonic strain recorded in this part of the D1 domain. Angularity of devitrified glass shards is still evident. The westward plunging lineations are very weak but visible along a weakly developed schistosity in the rock, best seen along the margins of pillow breccia fragments. This outcrop is part of the same basaltic unit seen at the previous stop.

Continue east for 11.1 km. You will be travelling close to and obliquely crossing the contact between the Shebandowan assemblage and the underlying, 2720 Ma volcanics. A marked change in orientation of stretching lineations occurs along this contact, from the D_2, easterly plunging lineations of the D_2 domain observed thus far to the westerly plunging D_1 domain observed in the next two outcrops and characterizing the southern half of the greenstone belt (Figure 29).

***The Shebandowan Assemblage** comprises both a suite of volcanic rocks, typified by subalkalic to alkalic pyroclastic units (Carter, 1993), and a suite of tightly folded, clastic and chemical metasedimentary rock (Figure 34 is a map of a representative portion of this assemblage). The metasedimentary units generally overlie the volcanic rocks and together they unconformably overlie the older 2720 Ma volcanics. The unconformities are generally not visible but are inferred from geological and aeromagnetic map patterns. Banded magnetite-chert iron formation, including jasper iron formation, occurs interbedded with greywacke. Jasper iron formation is a distinctive feature of this greenstone belt and is widely interbedded with older 2720 Ma volcanic rocks as well as units of the 2690 Ma Shebandowan assemblage. The clastic metasedimentary units include pebble and cobble conglomerate derived from late plutonic and volcanic sources as well as eroded sedimentary rocks, including jasper iron formation. The Shebandowan assemblage is restricted largely to two subparallel zones that locally show geophysical evidence that they unconformably overlie subvertically dipping 2720 Ma volcanics. The Shebandowan assemblage has been interpreted (e.g. Shegelski 1980) as a subaerial to shallow subaqueous assemblage restricted to fault-bounded basins infilled with alkalic volcanic rocks and alluvial-fluvial and submarine fan*

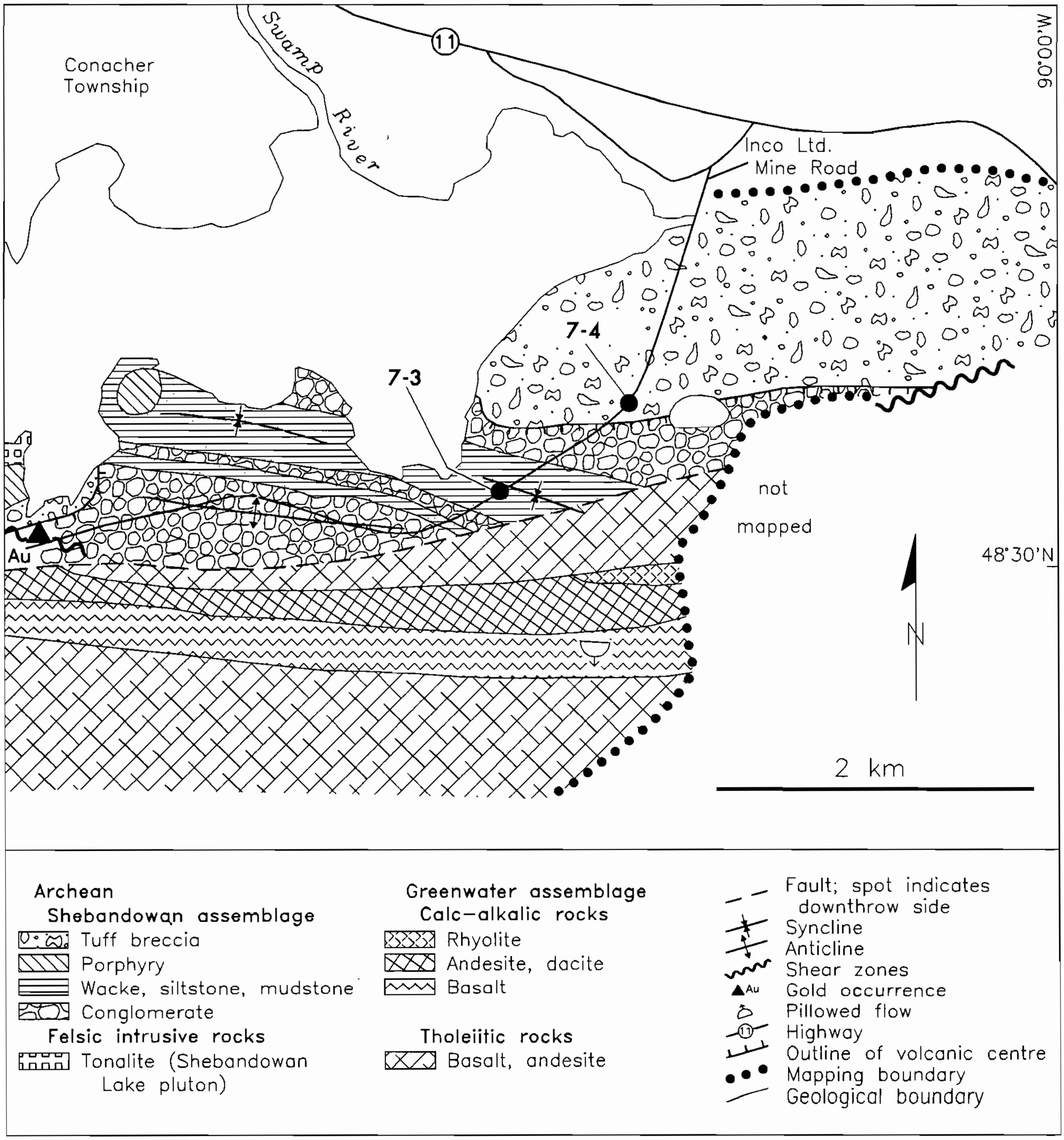


Figure 34. Inferred faulted contact between folded rocks of the Shebandowan assemblage and older 2720 Ma rocks of the Greenwater assemblage (modified from Williams et al. 1991).

deposits. The suite of rocks of this assemblage is similar to rocks of the Timiskaming assemblage concentrated along a major fault zone in the Abitibi greenstone belt.

Stop 7-3 Conglomerate and Slate of the Shebandowan Assemblage

The conglomerate comprises clasts derived from granitoid porphyry, volcanic rocks, chert and hematitic jasper iron formation. It is in contact with a dark grey siltstone unit. Note that the long axes of clasts plunge consistently eastward. A sample of a quartz-phyric trondhjemite clast taken from this outcrop gave a U-Pb zircon age of 2704 +/-2 Ma, slightly older than the nearby 2698 Ma Shebandowan pluton of similar composition and older than the 2689 +3/-2 Ma latitic breccia of Stop 6-6.

CAUTION: THE ROAD BEND PREVENTS CLEAR VISIBILITY. PLEASE WATCH OUT FOR ANY TRAFFIC WHILE YOU WANDER BACK AND FORTH ACROSS THE ROAD.

Stop 7-4 Shoshonitic Autobreccia and Sandstone/Siltstone of the Shebandowan Assemblage

This is a typical exposure of the dark red to olive green, hornblende-phyric volcanic breccia that so characterizes the Shebandowan Assemblage and can be seen further east along Highway 11 and 11/17. It is interbedded with grey, thinly bedded (and locally folded) sandstone and siltstone. Note that the colour variation across the breccia is locally diffuse or sharp across fragment contacts. Colour rims mark some of the fragments and both matrix and fragments contain hornblende phenocrysts, a common distinguishing feature of late volcanic units in some Archean greenstone belts. This autoclastic flow is characterized by a monomictic jigsaw collection of angular to subrounded blocks with little or no evidence of quenched clast margins. Owing to the potential for such subaerial or shallow water units to be eroded and closely associated with their erosional equivalents as sedimentary conglomerates, it can be a challenge to distinguish between them in the field. The 2690 Ma volcanic breccia is latitic in composition and is part of a shoshonitic suite accompanied by monzodioritic to syenitic dikes, sills and stocks in this region. Most alkalic stocks, including syenitic plutons and pegmatite masses in the belt, postdate the deformation that dominates the northern half of the belt and are part of the sanukitoid suite of plutons described by Stern et al. (1989). An age of 2689 Ma was obtained from this outcrop (Corfu and Stott, 1986) and typifies the age of vocanism for the Shebandowan assemblage (Figure 28). The timing relationships will be discussed on the outcrop, based on more recent age determinations of volcanic and epiclastic zircons by Fernando Corfu.

Return to Highway 11 where we turn right and proceed to the junction of Highway 11 and 17. Turn right and continue on 11/17 for 20.8 km to Highway 102. Turn left and continue down Hwy 102 for 3.9 km to the Kaministiquia River, cross the bridge into Ware Township and continue for another 4.6 km to a long, high roadcut. We will pull into a road on the right just past the roadcut. Please do not park on the side of the highway near the roadcut, there is very little clearance.

(If we have time we can first make an optional extra stop on Silver Falls Road, just 0.6 km east of the Kaministiquia River. We can proceed about 1 km up this road to look briefly at the intense shear zone in basalt marking the boundary between the Shebandowan belt and the Quetico metasedimentary subprovince.)

Please beware of the danger of falling rock at this roadcut.

Stop 7-5 Shebandowan Assemblage - "The Strawberry Hill" Autobreccia and Lapilli Tuff

This roadcut is a volcanic breccia typical of the broad aprons of reddish to dark green weathered fragmental rocks that give a maximum U-Pb zircon age of 2692 Ma (Corfu and Stott, 1995). The various autoclastic to pyroclastic units in this assemblage include a spectrum of compositions ranging from calc-alkalic to locally alkalic basalt to rhyolite (Brown 1995). The coarse breccia units and lapilli tuff are most typical; recognizable fine-grained flows and ash-flows are rare. All units of this assemblage contain amphibole phenocrysts. Many units are marked by hematite staining and feature weaker deformation than most units of the older Greenwater assemblage.

Continue eastward along Highway 102 for about 2 km. Park just west of the outcrop. Cross to the north side of the highway. Be careful of the traffic.

Stop 7-6 Mud Lake Calc-Alkalic Felsic Volcanics and Cu-Zn Occurrence

The Mud Lake Cu-Zn occurrence is part of a 50 to 70 cm wide banded chert + sulphide ± magnetite horizon on the northeast side of Highway 102. Top directions from mafic pillowed flows 400 m southeast of the occurrence indicate that the metavolcanic rocks young to the northwest. Assuming this stratigraphic orientation, the footwall succession to the base metal mineralized horizon is an approximately 100 m thick succession of quartz ± feldspar-phyric felsic flows and rare lapilli tuffs. Quartz and feldspar phenocrysts are commonly between 1 and 7 mm long. The stratigraphic hanging wall is composed of felsic to intermediate feldspar-phyric fragmental rocks and minor massive flows. Fragment and matrix compositions are similar. The Mud Lake succession is calc-alkaline and proximal to a nearby felsic centre to the south. This is in contrast to the aphyric FIII-like felsic fragmental rocks in the more economic central part of the greenstone belt. The base metal mineralized horizon at Mud Lake is considered to be of limit extent and has not been identified along strike. A U-Pb zircon age of 2718 Ma has been obtained from the quartz phyric rhyolite of this outcrop.

Continue east on Highway 102 for about 1.8 km to Auto Road. Turn left onto this road and drive just over 2 km to a small outcrop of conglomerate on the left.

Stop 7-7 Auto Road Conglomerate, Ware Township

This small outcrop represents a very late syntectonic epiclastic deposit. Detrital zircons from the matrix of this rock have been dated by F. Corfu at ≤2682 Ma, which implies that some comparatively very young sedimentation occurred at least in this area, close to the boundary with the Quetico subprovince. The conglomerate is dominantly composed of medium-grained granodiorite cobble and pebble clasts in a wacke matrix. Other scarce clasts include massive sulphide and a plagioclase porphyry with a very fine grained reddish matrix resembling porphyries associated with the Shebandowan assemblage. This outcrop represents one of several conglomerate units, which elsewhere typically grade to wacke and siltstone and are structurally interlayered with the steeply foliated, enveloping volcanics.

We will drive to the Thunder Bay airport at the conclusion of this trip, unless participants have made other travel arrangements.

End of Trip.

References

Ayres, L.D. and Corfu, F. 1991. Stacking of disparate volcanic and sedimentary units by thrusting in the Archean Favourable Lake greenstone belt, central Canada; Precambrian Research, v. 50, p. 221-238.

Barrie, C.T., Ludden, J.N. and Green, T.H., 1993. Geochemistry of volcanic rocks associated with Cu-Zn and Ni-Cu deposits in the Abitibi subprovince. Economic Geology, 88, 1341-1358.

Beakhouse, G.P., 1989. The Sioux Lookout Terrane: An imbricate thrust stack related to a 2.71 Ga arc-continent collision. Geological Association of Canada Abstracts, 14, p.A35-36.

Beakhouse, G.P., 1991. Winnipeg River subprovince; in Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 1, p.279-301.

Beakhouse, G.P., 1993. Werner-Bug Lakes area: Base metal mineralization in a high-grade metasedimentary terrane. in Summary of Field Work and Other Activities 1993, Ontario Geological Survey, Miscellaneous Paper 162,

Beakhouse, G.P. and McNutt, R.H., 1986. Geochemistry of granitoid rocks from the western Superior Province- Evidence for 2- and 3-stage crustal evolution models, in Workshop on Early Crustal Genesis: The World's Oldest Rocks, Lunar and Planetary Institute, Technical Report 86-04, p.23-26.

Beakhouse, G.P. and McNutt, R.H., 1991. Contrasting types of Late Archean plutonic rocks in northwestern Ontario: Implications for crustal evolution in the Superior Province. Precambrian Research, 49, 141-165.

Beus, A.A., Berengilova, V.V., Grabovskaya, L.I., Kochemasov, G.G., Leonteva, L.A. and Sitnin, A.A. 1968. Geochemical prospecting for endogeneous ore deposits of rare elements on the example of tantalum; Nedra, Moscow, 429p. (in Russian but translated by the Geological Survey of Canada).

Blackburn, C.E. 1980. Towards a mobilistic tectonic model for part of the Archean of northwestern Ontario; Geoscience Canada, v.7, p.64-72.

Blackburn, 1981. Geology of the Boyer Lake- Meggisi Lake area, District of Kenora. Ontario Geological Survey Report 202, 107 p. Accompanied by Maps 2437 and 2438, scale 1:31680.

Blackburn, C.E. and Young, J.B. 1993. Geology of the Separation Lake greenstone belt; in Summary of field Work and Other Activities 1993, Ontario Geological Survey, Miscellaneous Paper 162, p.68-73.

Blackburn, C.E., Johns, G.W., Ayer, J. and Davis, D.W., 1991. Wabigoon subprovince; in Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 1, p.303-381.

Borradaile, G.J., Kehlenbeck, M.M. and Werner, T.W. 1994. A magnetotectonic study correlating late Archean deformation in northwestern Ontario; Canadian Journal of Earth Sciences, v.31, p.1449-1460.

Breaks, F.W., 1991. English River subprovince; in Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 1, p.239-277.

Breaks, F.W., 1993. Granite-related mineralization in northwestern Ontario: Raleigh Lake and Separation Rapids (English River) rare element pegmatite fields. in Summary of Field Work and Other Activities 1993, Ontario Geological Survey, Miscellaneous Paper 162, 104-110.

Breaks, F.W. and Moore, J.M., 1992. The Ghost Lake batholith, Superior Province of northwestern Ontario: a fertile, S-type, peraluminous granite-rare-element pegmatite system; Canadian Mineralogist, volume 30, 835-875.

Breaks, F.W. and Janes, D.A. 1991. Granite-related mineralization of the Dryden area, Superior Province of northwestern Ontario; Geological Association of Canada and Mineralogical Association of Canada, Field Trip Guidebook B7, 71p.

Brown, G.H. 1995. Precambrian Geology, Oliver and Ware townships; Ontario Geological Survey, Report 294, 48p.

Card, K.D., 1990. A review of the Superior Province of the Canadian shield, a product of Archean accretion. Precambrian Research, 48, 99-156.

Card, K.D. and Ciesielski, A., 1986. Subdivisions of the Superior Province of the Canadian shield. Geoscience Canada, 13, 5-13.

Carter, M.W. 1992. Geology and mineral potential of the Tower Syenite stock, Conmee Township, District of Thunder Bay; *in* Summary of Field Work and Other Activities 1992, Ontario Geological Survey, Miscellaneous Paper 160, p.60-63.

Carter, M.W. 1993. The Geochemical Characteristics of Neoarchean alkalic magmatism in central Superior Province, Ontario; *in* Summary of Field Work and Other Activities 1993, Ontario Geological Survey, Miscellaneous Paper 162, p.13-19.

Cerny, P. and Meintzer, R.E. 1988. Fertile granites in the Archean and Proterozoic fields of rare-element pegmatites: crustal environment, geochemistry and petrogenetic relationships in Recent Advances in the Geology of Granite-Related Mineral Deposits (R.P. Taylor and D.F. Strong, editors); Canadian Institute of Mining and Metallurgy, Special Volume 39, 170-207.

Cerny, P., Trueman, D.L., Ziehlke, D.V., Goad, B.E. and Paul, B.J., 1981. The Cat Lake - Winnipeg River and the Wekusko Lake pegmatite fields, Manitoba. Manitoba Department of Energy and Mines, Economic Geology Report ER80-1, 216 p.

Corfu, F., 1988. Differential response of U-Pb systems in coexisting accessory minerals, Winnipeg River subprovince, Canadian shield: Implications for Archean crustal growth and stabilization. Contributions to Mineralogy and Petrology, 98, 312-325.

Corfu, F. and Andrews, A.J. 1987. Geochronological constraints on the timing of magmatism, deformation, and gold mineralization in the Red Lake greenstone belt, northwestern Ontario; Canadian Journal of Earth Sciences, v. 24, p. 1302-1320.

Corfu, F. and Davis, D.W., 1992. A U-Pb geochronological framework for the western Superior Province, Ontario; in Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 2, p.1335-1346.

Corfu, F. and Stott, G.M. 1986. U-Pb ages for late magmatism and regional deformation in the Shebandowan Belt, Superior Province, Canada; Canadian Journal of Earth Sciences, v.23, p.1075-182.

Corfu, F. and Stott, G.M. 1995. U-Pb geochronology of the Shebandowan greenstone belt, Superior Province, Ontario [abst]; Institute on Lake Superior Geology, 41st Annual Meeting, Marathon, ON, 1995, v.41, part 1, p. 6.

Davies, J.C., 1965. Geology of the High Lake - Rush Bay area, District of Kenora. Ontario Department of Mines, Geological Report 41, accompanied by Maps No. 2068 and 2069, scale, 1:31680.

Davis, D.W. 1990. The Seine-Couchiching problem reconsidered: Up-Pb geochronological data concerning the source and timing of Archean sedimentation in the Western Superior Province; *in* Proceedings, 36th annual Meeting, Institute on Lake Superior Geology, pt.1, Abstracts, p.19-21.

Davis, D.W. and Edwards, G.R., 1985. The petrogenesis and metallogenesis of the Atikwa - Lawrence volcanic - plutonic terrain; in Geoscience Research Grant Program, Summary of Research 1984-85, Ontario Geological Survey, Miscellaneous Paper 127, 101-111.

Davis, D.W. and Edwards, G.R., 1986. Crustal evolution of Archean rocks in the Kakagi Lake area, Wabigoon subprovince, Ontario as interpreted from high-precision U-Pb geochronology. Canadian Journal of Earth Sciences, 23, 182-192.

Davis, D.W. and Jackson, M.C. 1988. Geochronology of the Lumby Lake greenstone belt: a 3 Ga complex within the Wabigoon Subprovince, northwest Ontario; Geological Society of America, Bulletin, v. 100, p. 818-824.

Davis, D.W., Blackburn, C.E. and Krogh, T.E., 1982. Zircon U-Pb ages from the Wabigoon - Manitou Lakes region, Wabigoon subprovince, northwest Ontario. Canadian Journal of Earth Sciences, 19, 254-266.

Davis, D.W., Sutcliffe, R.H. and Trowell, N.F., 1988. Geochronological constraints on the tectonic evolution of a Late Archean greenstone belt, Wabigoon subprovince, northwest Ontario, Canada. Precambrian Research, 39, 171-191.

Davis, D.W. and Smith, P.M., 1991. Archean gold mineralization in the Wabigoon subprovince, A product of crust accretion: Evidence from U-Pb geochronology in the Lake of the Woods area, Superior Province, Canada. Journal of Geology, 99, 337-353.

Devaney, J.R. and Williams, H.R., 1989. Evolution of an Archean subprovince boundary: A sedimentological and structural study of part of the Wabigoon-Quetico boundary in northern Ontario. Canadian Journal of Earth Sciences, 26, 1013-1026.

Fumerton, S.L., 1982. Redefinition of the Quetico fault near Atikokan, Ontario. Canadian Journal of Earth Sciences, 19, 222-224.

Fyon, J.A., Breaks, F.W., Heather, K.B., Jackson, S.L., Muir, T.L., Stott, G.M. and Thurston, P.C., 1992. Metallogeny of Metallic Mineral Deposits in the Superior Province of Ontario in Geology of Ontario, Special Volume No. 4, 1091-1176.

Gower, C.F., 1978. The tectonic and petrogenetic history of Archean rocks from the Kenora area, English River subprovince, northwest Ontario. Unpublished PhD thesis, McMaster University, Hamilton, Ontario, 622 p.

Gower, C.F., Crocket, J.H. and Kabir, A., 1983. Petrogenesis of Archean granitoid plutons from the Kenora area, English River subprovince, northwest Ontario, Canada. Precambrian Research, 22, 245-270.

Hoffman, P.F., 1989. Precambrian geology and tectonic history of North America- An overview. Geological Society of America, The Geology of North America, v. A, p.447-512.

Hudleston, P.J., Schultz-Ela, D. and Southwick, D.L. 1988. Transpression in an Archean greenstone belt, northern Minnesota; Canadian Journal of Earth Sciences, v.25, p.1060-1068.

Jolliffe, A.W. 1955. Geology and iron ores of Steep Rock Lake; Economic Geology, v. 50, p.373-398.

Lesher, C.M., Goodwin, A.M., Campbell, I.H. and Gorton, M.P., 1986. Trace-element geochemistry of ore-associated and barren felsic metavolcanic rocks in the Superior Province, Canada. Canadian Journal of Earth Sciences, 23, 222-237.

Mackasey, W.O., Blackburn, C.E. and Trowell, N.F. 1974. A regional approach to the Wabigoon -Quetico belts and its bearing on exploration in northwestern Ontario; Ontario Department of Mines, Miscellaneous Paper 58, 29p.

Ontario Geological Survey, 1991. Bedrock geology of Ontario, west-central sheet. Ontario Geological Survey Map 2542, scale 1:1000000.

Osmani, I.A. 1993. Geology and mineral potential of Burchell-Greenwater lakes area, District of Thunder Bay. Ontario Geological Survey, Open File Report 5866, 81p.

Osmani, I.A. (submitted). Geology and mineral potential of the Greenwater Lake area, west-central Shebandowan greenstone belt. Ontario Geological Survey, unpublished geological report.

Parker, J.R., Blackburn, C.E. and Davis, D.W. 1989. Constraints on timing and placement of gold mineralization in the Wabigoon Subprovince near Dryden, Ontario: evidence for synvolcanic through late tectonic emplacement; *in* Program with Abstracts, Geological Association of Canada-Mineralogical Association of Canada, Annual Meeting, v.14, p.A92.

Percival, J.A., 1989. A regional perspective of the Quetico metasedimentary belt, Superior Province, Canada. Canadian Journal of Earth Sciences, 26, 677-693.

Percival, J.A. and Card, K.D., 1983. Archean crust is revealed in the Kapuskasing uplift, Superior Province, Canada. Geology, 11, 323-326.

Peterman, Z.E. and Day, W. 1989. Early Proterozoic activity on Archean faults in the western Superior Province - evidence from pseudotachylite; Geology, v.17, p.1089-1092.

Pichette, R.J., 1976. Petrology and geochemistry of the Taylor Lake stock, Superior Province, northwest Ontario. Unpublished BSc thesis, McMaster University, 99 p.

Sanborn-Barrie, M. 1988. Geology of the tectonic boundary between the English River and Winnipeg River subprovinces, northwestern Ontario; *in* Summary of Field work and Other Activities 1988, Ontario Geological Survey, Miscellaneous Paper 141, p.12-17.

Sanborn-Barrie, M., 1991. Structural geology and tectonic history of the eastern Lake of the Woods greenstone belt, Kenora District. Ontario Geological Survey, Open File Report 5773, 137 p.

Sawyer, E.W. 1983. The structural history of a part of the Archaean Quetico metasedimentary belt, Superior Province, Canada; Precambrian Research, v.22, p.271-294.

Schaeffer, S.J. and Morton, P. 1991. Two komatiitic pyroclastic units, Superior Province, northwestern Ontario: their geology, petrology and correlation; Canadian Journal of Earth

Sciences. v.28, p.1455-1470.

Shegelski, R.J. 1980. Archean cratonization, emergence, and red bed development, Lake Shebandowan area, Canada; Precambrian Research, v.12, p.331-347.

Spilde, M.N. and Shearer, C.K. 1992. A comparison of tantalum-niobium oxide assemblages in two mineralogically distinct rare-element granitic pegmatites, Black Hills, South Dakota; Canadian Mineralogist, Volume 30, 719-737.

Stern, R.A., Hanson, G.N. and Shirey, S.B., 1989. Petrogenesis of mantle-derived, LILE-enriched Archean monzodiorites and trachyandesites (sanukitoids) in southwestern Superior Province. Canadian Journal of Earth Sciences, 26, 1688-1712.

Stone, D., Kamineni, D.C. and Jackson, M.C. 1992. Precambrian geology of the Atikokan area, northwestern Ontario; Geological Survey of Canada, Bulletin 405, 106p.

Stott, G.M. 1986. A structural analysis of the central part of the Archean Shebandowan greenstone belt and a crescent-shaped granitoid pluton, northwestern Ontario, unpublished PhD thesis, University of Toronto, Toronto, Ontario, 285p.

Stott, G.M. and Schnieders, B.R. 1983. Gold mineralization in the Shebandowan Belt and its relation to regional deformation patterns; in The Geology of Gold in Ontario, Ontario Geological Survey, Miscellaneous Paper 110, p.181-193.

Sutcliffe, R.H., Smith, A.R., Doherty, W. and Barnett, R.L., 1990. Mantle derivation of Archean amphibole-bearing granitoid and associated mafic rocks: evidence from the southern Superior Province, Canada. Contributions to Mineralogy and Petrology, 105, 255-274.

Teal, P.R. and Walker, R.G. 1977. Stratigraphy and sedimentology of the Archean Manitou group, northwestern Ontario; in Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, p.181-184.

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

Wilks, M.E. and Nisbet, E.G. 1988. Stratigraphy of the Steep Rock Group, northwestern Ontario: a major Archean unconformity and Archean stromatolites; Canadian Journal of Earth Sciences, v. 25, p. 370-391.

Williams, H.R., 1990. Subprovince accretion tectonics in the south-central Superior Province. Canadian Journal of Earth Sciences, 27, 570-581.

Williams, H.R., 1991. Quetico subprovince; in Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 1, p.383-403.

Williams, H.R., Stott, G.M., Heather, K.B., Muir, T.L. and Sage, R.P., 1991. Wawa subprovince; in Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 1, p.485-539.

Williams, H.R., Stott, G.M., Thurston, P.C., Sutcliffe, R.H., Bennett, G., Easton, R.M. and Armstrong, D.K., 1992. Tectonic evolution of Ontario: Summary and synthesis. in Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 2, p.1255-1332.

WINNIPEG'96 GUIDEBOOK ORDER FORM

A1. Evolution of the Thompson Nickel Belt, Manitoba: Setting of Ni-Cu Deposits in the Western Part of the Circum Superior Boundary Zone
W. Bleeker and J. Macek..........*$12.00 x no. of copies:* ______

A2. Late Holocene Environmental Changes in Southern Manitoba
E. Nielsen, K. D. McLeod, E. Pip and J.C. Doering.......... *$8.00 x no. of copies:* ______

A3. Petrology and Mineralization of the Tanco Rare-Element Pegmatite, Southeastern Manitoba.
P. Cerny, S. Ercit and P. Vanstone..........*$12.00 x no. of copies:* ______

A4. Lithostratigraphic Assembly of the Eastern Rice Lake Greenstone Belt and Structural Setting of Gold Mineralization at Bissett, Manitoba.
K.H. Poulson, W. Weber, R. Brommecker and D.Seneshen..........*$13.00 x no. of copies:* ______

A5/B6. Western Superior Transects: Wabigoon-Quetico-Shebandowen and English River-Winnipeg River-Wabigoon
G. Beakhouse, G. Stott, C. Blackburn, F.W. Breaks, J. Ayer, D. Stone, C. Farrow and F. Corfu..........*$13.00 x no. of copies:* ______

B1. Tectonic Assembly of the Paleoproterozoic Flin Flon Belt and Setting of VMS Deposits
E. Syme, A. Bailes and S.Lucas..........*$15.00 x no. of copies:* ______

B2. Geomorphic and Sedimentological History of the Central Lake Agassiz Basin.
J. Teller, H. Thorleifson and G. Matile..........*$14.00 x no. of copies:* ______

B3. Physical Volcanology, Hydrothermal Alteration and Massive Sulphide Deposits of the Sturgeon Lake Caldera
R. Morton, G. Hudak and E. Koopman..........*$10.00 x no. of copies:* ______

B4. Lower to Middle Paleozoic Stratigraphy of Southwestern Manitoba
R. Bezys and H. R. McCabe..........*$12.00 x no. of copies:* ______

B5. Geology of the Lac du Bonnet Batholith, Inside and Out: AECL's Underground Research Laboratory, S. E. Manitoba
R. Everitt, J. McMurry, C. Davison and A. Brown.......... *$10.00 x no. of copies:* ______

B7. Industrial Minerals of S.E.Manitoba
B. Schmidtke and J. Bamburak..........*$3.00 x no. of copies:* ______

The entire set may be purchased for $95.00

Subtotal: ______
+GST (7%) ______
+Postage and Handling ______
TOTAL ______

Orders should be sent to:
Geological Association of Canada, Winnipeg Section
c/o Geological Services Branch, Manitoba Energy and Mines
1395 Ellice Ave., Suite 360
Winnipeg, Manitoba R3G 3P2

Make cheques or money orders payable to "GAC-MAC '96". Add $2.00 per book for postage and handling. For entire set add $10.00.