# The Archean Crust in the Wawa-Chapleau-Timmins Region

A field guidebook prepared for the 1983
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Genesis Field Conference

by

J.A. Percival, K.D. Card,
Geological Survey of Canada,
Ottawa, Ontario K1A 0E4

R.P. Sage, L.S. Jensen,
Ontario Geological Survey,
Toronto, Ontario M5S 1B3

and

L.E. Luhta,
Resident Geologist,
Ontario Ministry of Natural Resources,
Timmins, Ontario P4N 2S7

## Introduction

The purpose of the trip is to examine the characteristics and interrelationships of Archean greenstone-granite and high-grade gneiss terranes of the Superior Province. A 300-km long west to east transect between Wawa and Timmins, Ontario will be used to illustrate regional-scale relationships.

Figure 1 shows the major geological features of the Superior Province and Figure 2 traces the trip route. The first day will be spent examining features of the Michipicoten belt, a dominantly metavolcanic portion of the Wawa subprovince. On day two, the contact relationships between the Michipicoten supracrustal rocks and intrusions of the Wawa domal gneiss terrane will be examined, followed by a look at the boundary between the Wawa terrane and Kapuskasing structural zone. Day three will be spent mainly in the Kapuskasing zone examining the Shawmere anorthosite complex, and high-grade gneisses, as well as the Ivanhoe Lake cataclastic zone separating rocks of the Kapuskasing zone from those of the Abitibi belt. On day four the geology of the Abitibi belt in the Timmins area will be outlined.

## Regional Setting

The Superior Province is an Archean terrane composed of east-west trending belts of alternate volcanic-rich and sediment-rich character, termed subprovinces (Fig. 1). The continuity of the east-west belts is interrupted by a northeast-trending zone of high-grade metamorphic rocks, the Kapuskasing structural zone (Thurston et al., 1977). The Kapuskasing structure is fault-bounded on the southeast but the western contact is complex and gradational over 120 km to low-grade rocks of the Michipicoten belt near Lake Superior (Figs. 1 and 2).

The Kapuskasing "high", a prominent northeasterly gravity and aeromagnetic anomaly, was interpreted by Wilson and Brisbin (1965) to indicate pronounced upwarp of the Conrad discontinuity. Bennett et al. (1967) concluded that the Kapuskasing structure is a complex horst uplifted during the Proterozoic. The association of 1,100-1,000 Ma
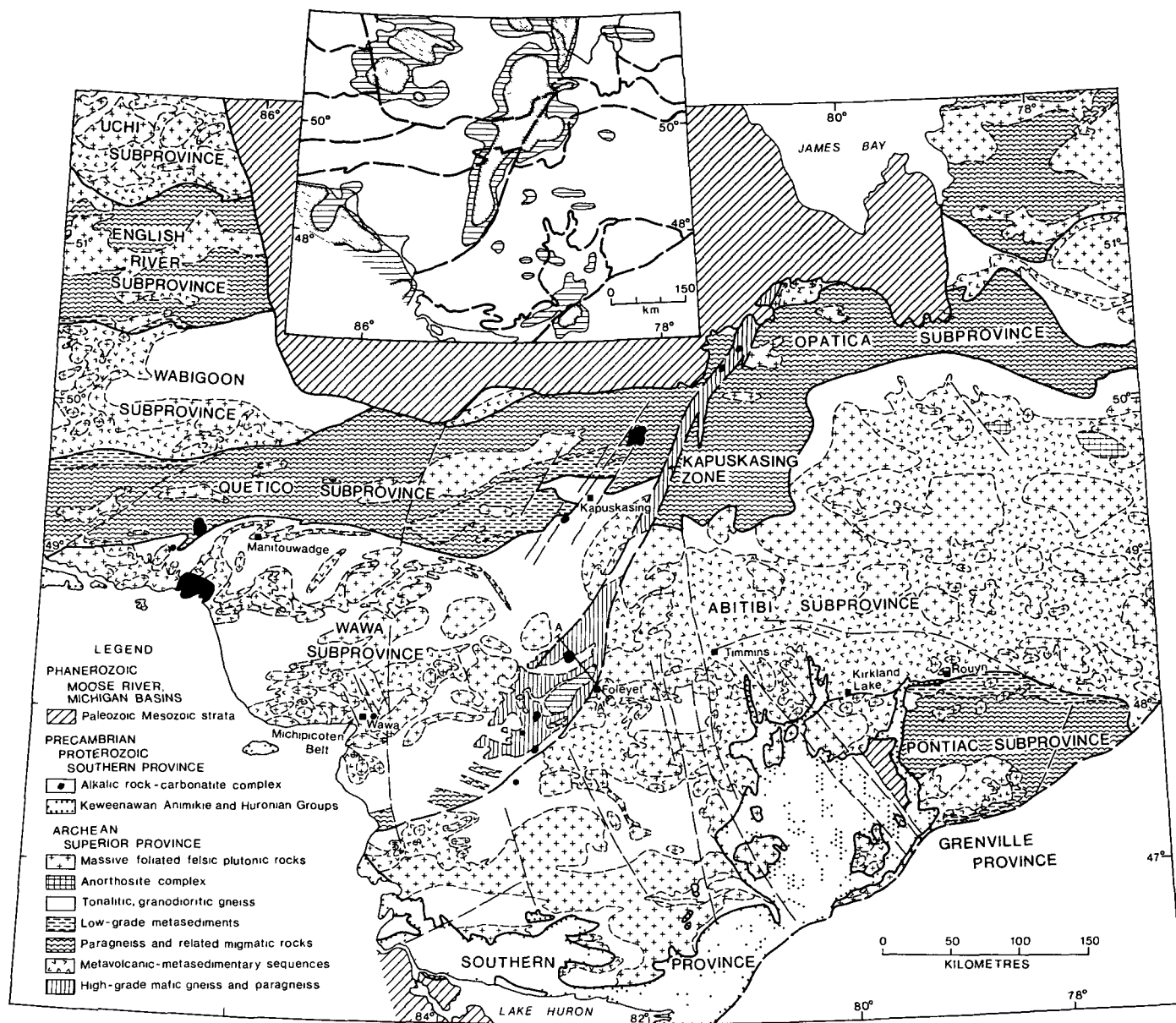
Fig. 1.    Geology of the central Superior Province.  Inset: positive gravity anomalies dark stipple (>-25 mGal); lines (-25 mGal to -35 mGal); blank (<-35 mGal).

alkalic rock-carbonatite complexes led Burke and Dewey (1973) to suggest that the

Kapuskasing structure is a failed arm of the Keweenawan rift structure.  Watson (1980)

postulated that the Kapuskasing zone was uplifted during late Archean or early
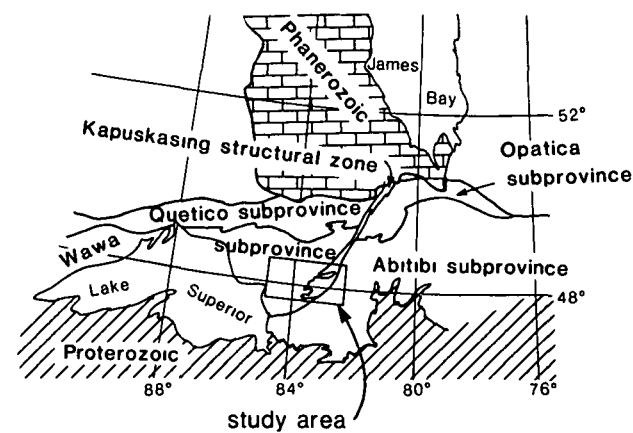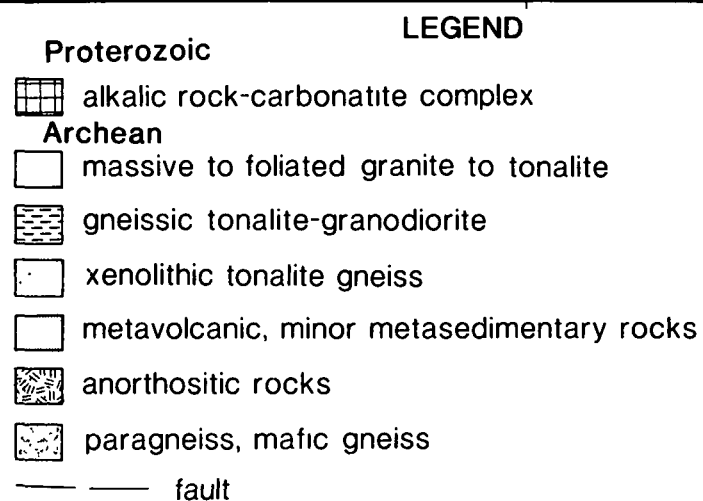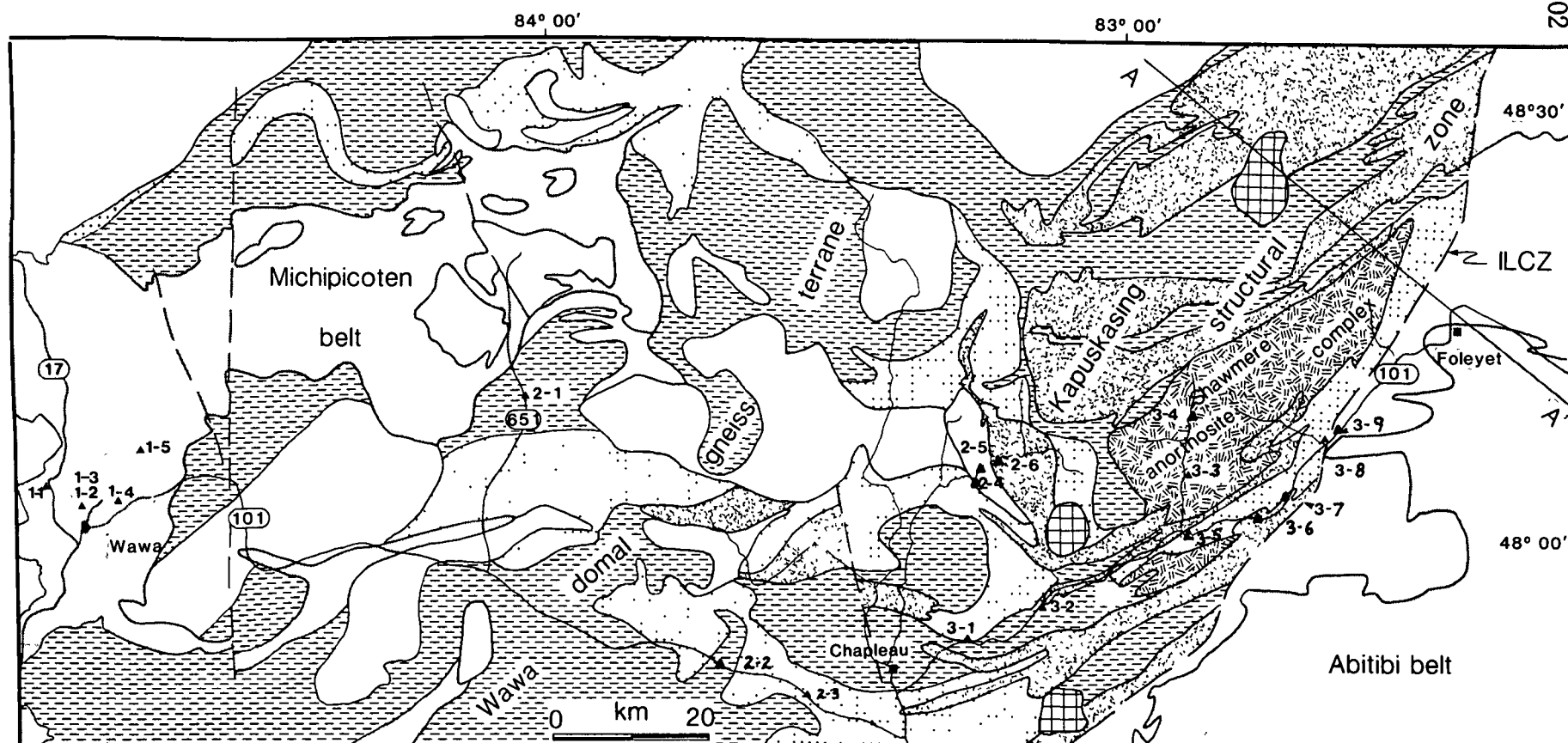
Fig. 2. Geology of the field trip route showing major rock units and tectonic subdivisions.

LEGEND

Proterozoic

▦ alkalic rock-carbonatite complex

Archean

☐ massive to foliated granite to tonalite

▨ gneissic tonalite-granodiorite

☐ xenolithic tonalite gneiss

☐ metavolcanic, minor metasedimentary rocks

▨ anorthositic rocks

☒ paragneiss, mafic gneiss

——— fault

Proterozoic sinistral transcurrent movement. Recent earthquakes in the region indicate that the structure is still active (Forsyth and Morel, 1982).

## Geophysical Characteristics of south-central Superior Province

A map showing apparent crustal thickness in the Lake Superior region, based on seismic refraction studies, was presented by Halls (1982) (Fig. 3). These data show that the apparent thickness of the crust of the Superior Province decreases easterly from values of about 45 km near Wawa to 35 km near Timmins. In addition, there appears to be a step-like decrease in depth to Moho associated with the eastern boundary of the Kapuskasing zone. Values in the 39 km range beneath the Kapuskasing zone drop abruptly to 35 km to the east. Halls viewed the contour map of apparent crustal thickness as preliminary, partly because of insufficient coverage in some areas and partly due to the necessity of averaging crustal and mantle velocities. Accounting for the high-density, presumably high-velocity rocks in the Kapuskasing zone would increase the crustal thickness estimates beneath the structure.

The Bouguer gravity anomaly map for the Wawa-Timmins region is shown in Figure 4. In general, areas underlain by metavolcanic rocks have associated positive gravity anomalies and granitoid-gneissic rocks have negative anomalies. The Kapuskasing structural zone has an associated linear positive gravity anomaly extending from James Bay in the north to some 50 km southwest of Chapleau. In the Wawa-Chapleau-Foleyet area, the gradient is gradual on the west and abrupt on the east, suggesting a west-dipping contact between the Kapuskasing zone and Abitibi subprovince. In this region, the gravity profile (Fig. 11) shows a paired high-low anomaly. The trough of the low is coincident with the fault at the eastern boundary of the Kapuskasing zone.
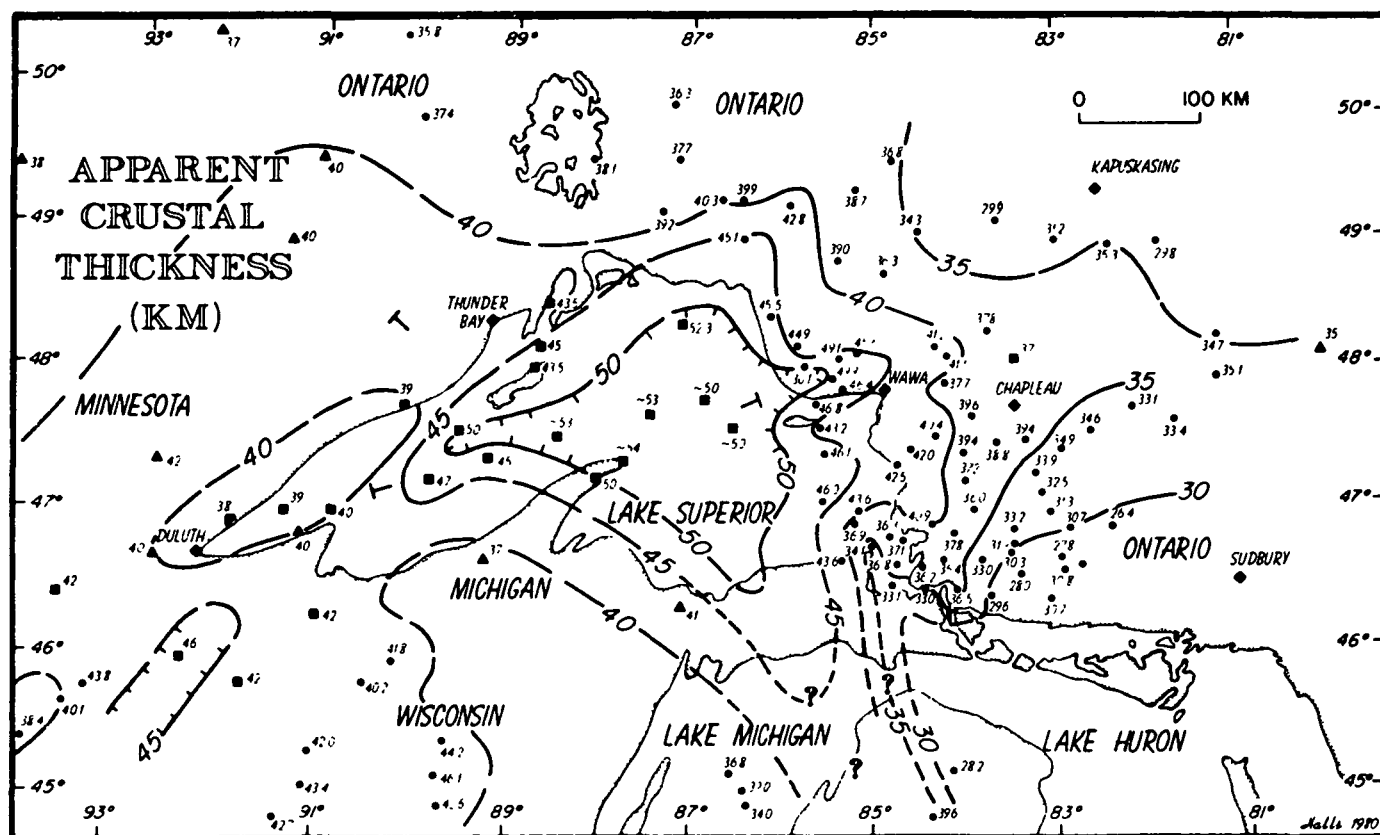
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Fig. 3    Apparent thickness of the crust in the Lake Superior area (after Halls, 1982).
Sources of data are listed by Halls and extreme caution in using map was
advised.

To the north, the positive Kapuskasing anomaly broadens as it coalesces with the

east-west gravity high associated with the Quetico-Opatica metasedimentary

subprovince.

Geology of Wawa and Abitibi subprovinces

The geology of the region including the Michipicoten and Abitibi belts and

Kapuskasing zone is shown in Figure 2.  The Michipicoten belt, part of the volcanic-rich

Wawa subprovince, is mainly composed of metavolcanic rocks of ultramafic, mafic and
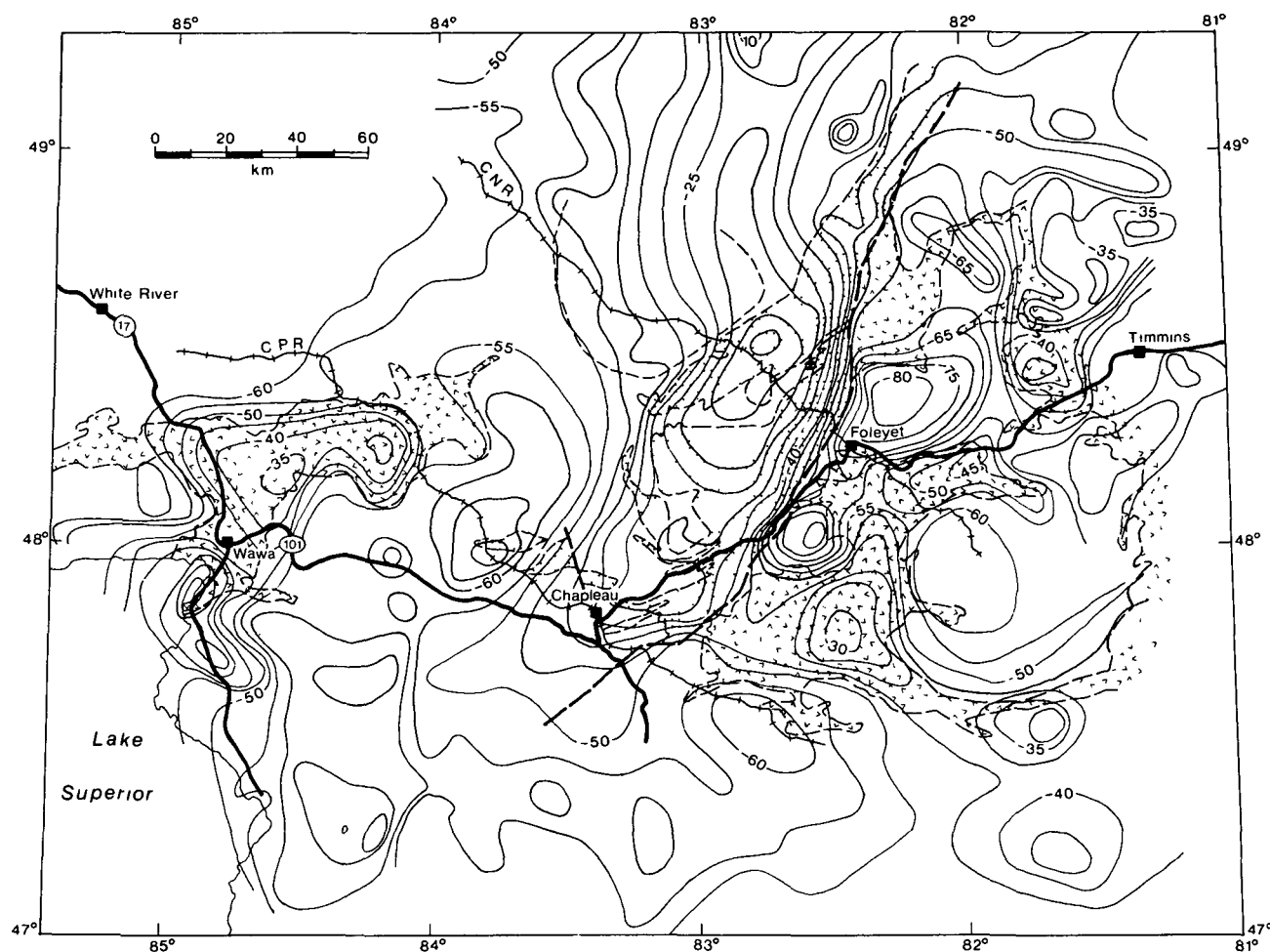
Fig. 4.    Bouguer gravity map superimposed on simplified geology. Gravity from Earth
           Physics Branch map 44078, 44084, 48078 and 48084). Checks - greenstone
           belts; dots - Kapuskasing gneiss.

felsic composition (Goodwin, 1962), with intercalated greywacke, conglomerate, iron

formation and chert. Dome and basin structures (Goodwin, 1962) as well as downward-

facing strata and overturned structures (Attoh, 1980) have been recognized.

Metamorphic grade ranges from sub-greenschist to amphibolite facies

(Fraser et al., 1978). Several suites of intrusive rocks include synvolcanic bodies ranging

from peridotite to granodiorite, younger granodiorite batholiths, and still younger granite

and syenite plutons (Card, 1982).

The Michipicoten belt is intruded to the southeast by tonalitic gneiss and plutons of the Wawa domal gneiss terrane (Fig. 2). The rocks in this region consist of at least four lithologic components: (1) hornblende-plagioclase $\pm$ clinopyroxene mafic and rare paragneiss xenoliths, ranging from centimetres to tens of metres in maximum dimension, making up 5 to 50% of individual outcrops, and enclosed in (2) the volumetrically most abundant phase, hornblende-biotite tonalitic gneiss which is cut by (3) concordant to discordant layers of foliated to gneissic biotite-hornblende granodiorite, which in turn are cut by (4) late discordant quartz monzonite pegmatite. Xenolith-rich tonalitic gneiss units alternate on a 5 to 10 km scale with xenolith-poor units and can be traced for distances of at least 50 km. Layering in mafic xenoliths is locally discordant to layering in enclosing gneiss. Small folds of layering in tonalitic gneiss are commonly truncated by layers of foliated granodiorite.

In the area between the Michipicoten belt and Kapuskasing zone (Fig. 2) the orientation of foliation, gneissosity and axial surfaces of small folds permit definition of several structural domains characterized by domal geometry. The spacing of major domal or antiformal culminations is on the order of 20 to 25 km, although many smaller culminations are also present. The Highbrush Lake and Racine Lake domes have cores of tonalite-granodiorite gneiss whereas the Chaplin Lake dome and Missinaibi Lake arch have granitic cores flanked by foliated to gneissic rocks. A planar fabric in the homogeneous granitic rocks, defined by lenticular quartz and biotite alignment, is generally concordant to gneissosity in mantling gneiss. The Robson Lake dome, adjacent to the Kapuskasing structural zone, has a core of interlayered mafic gneiss, paragneiss and tonalitic gneiss.

In general, asymmetric small folds of gneissic layering do not have a consistent sense of asymmetry with respect to domal culminations and are therefore not congruent with the domes. Near some domal crests, the orientation of gneissic layering, small

folds and lineations are widely variable to chaotic and define a pattern of coalescing domes.

Metasedimentary rocks occur in two locations in the eastern Wawa subprovince. A discontinuous, antiformal to domal belt of paragneiss west of the Racine Lake dome may be continuous to the east with paragneiss of the Kapuskasing zone (Figs. 2 and 5). Stretched-pebble metaconglomerate occurs in association with quartz wacke and amphibolite in the vicinity of Borden Lake. The polymictic (tonalite, granodiorite, meta-andesite, metasediments, amphibolite, vein quartz), clast-supported rock contains cobbles ranging from equant to constricted (1.5 m x 7 x 7 cm) with a prominent shallow northeast plunge. In cross-section the clasts vary from equidimensional to northwest-dipping ellipses.

The Floranna Lake complex is a strongly lineated and foliated complex crescentic pluton of intermediate composition that occurs between the Robson Lake and Racine Lake domes. The margins of the complex are fine- to medium-grained hypersthene-biotite granite, whereas the core contains medium-grained monzonite and diorite with rare gabbro and coarse biotite-clinopyroxene melagabbro layers. The least-deformed interior portions contain relict igneous(?) clinopyroxene and feldspar augen phenocrysts. Migmatitic quartz monzonite layers constitute up to 10% of some outcrops. The complex has similar structural and lithological characteristics to crescentic plutons of the Wabigoon subprovince of northwestern Ontario (Schwerdtner et al., 1979; Sutcliffe and Fawcett, 1979).

The eastern limit of the domal region is a semi-continuous zone of north, northeast and northwest striking, gently easterly-dipping gneissosity and easterly-plunging lineation. This curvilinear feature (Fig. 5) may represent the eastern extremity of a first-order dome of 75-100 km diameter, of which the individual structural domains

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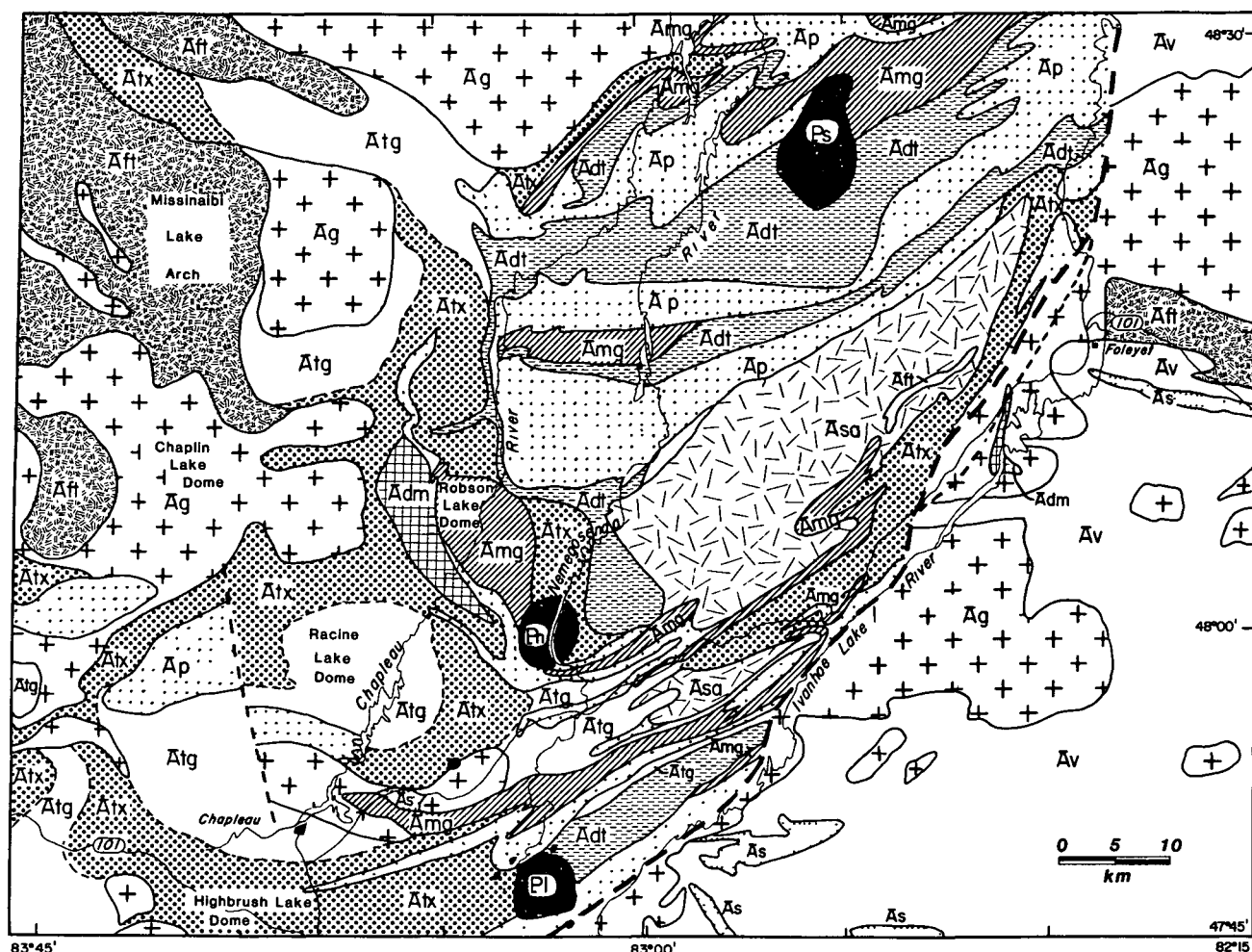


## Proterozoic

**P** Alkalic rock-carbonatite complex: l: Lackner Lake complex;  ⎫
n: Nemegosenda Lake complex; s: Shenango complex          ⎬ 1100 Ma intrusions
                                                           ⎭

## Archean

⊞ **Āg**  massive granite, granodiorite, with minor tonalite

⊞ **Ādm**  diorite-monzonite intrusive complex, minor hornblendite, granite

⊞ **Āft**  foliated to flaser tonalite

⊞ **Ātg,x**  tonalite-granodiorite gneiss; xenolithic

⊞ **Āv**  metavolcanic rocks, mainly metabasalt

⊞ **Ās**  metasedimentary rocks (includes metaconglomerate with tonalite
cobbles with a U-Pb zircon date of 2664±12 Ma)

⊞ **Ādt**  flaser diorite to mafic tonalite – includes minor gabbro,
hornblendite, granodiorite

⊞ **Āsa**  Shawmere anorthosite complex: metamorphosed gabbroic anorthosite,
anorthosite, gabbro, minor tonalite

⊞ **Āmg**  mafic gneiss: high Ca,Al basaltic composition, with tonalitic leucosome

⊞ **Āp**  paragneiss- quartz-rich composition, with up to 15% tonalitic leucosome

〰 fault; Ivanhoe Lake cataclastic zone

2707-2668 Ma sequence

2749-2696 Ma sequence

?

pre-2765 Ma sequence

**Fig. 5. Geology of the Kapuskasing structural zone and vicinity.**

are higher-order domes of similar scale and spacing to those of the Wabigoon subprovince (Schwerdtner and Lumbers, 1980).

Dome development can be related in time to the formation of minor structures in gneiss. The discordant foliations in mafic gneiss predate the gneissic layering in the tonalite-granodiorite host. Small folds of this gneissic layering in turn predate intrusion of granodiorite layers. Crosscutting pegmatite dykes and sills are still younger and are probably the same age as the homogeneous plutonic rocks which locally have a planar fabric defined by lenticular quartz grains, biotite alignment, fracture cleavage, or minor planar zones of granulation. The absence of a consistent sense of asymmetry of small folds with respect to domal culminations and the random orientation of small folds near dome crests argue in favour of re-orientation of pre-existing small folds and gneissic layering during the latest doming. The quartz-lenticle foliation and fracture cleavage in homogeneous plutonic rocks cannot be readily attributed to magmatic flow and therefore suggest that the plutons were emplaced at their present structural level at sub-solidus temperature.

The Abitibi subprovince is dominated by a thick sequence of volcanic and sedimentary rocks of the Abitibi greenstone belt (Jensen, 1981). The supracrustal succession typically comprises sequences of ultramafic, mafic, and felsic volcanics. Intercalated turbiditic sedimentary rocks contain a high proportion of volcanic detritus. In the Abitibi belt, the uppermost group, the Timiskaming, is an unconformity-bounded sequence of alkalic volcanics and fluviatile sediments (Hyde, 1980) localized along major east-west fault zones.

Large areas of the Abitibi greenstone belt are metamorphosed to greenschist facies; subgreenschist, prehnite-pumpellyite facies rocks are common in the Timmins-Rouyn area and narrow aureoles of amphibolite facies rocks occur adjacent to plutonic bodies (Jolly, 1978).

The supracrustal rocks of the Abitibi subprovince display evidence of polyphase deformation in the form of major and minor structures of several ages and orientations. In the Abitibi greenstone belt, older northerly-trending folds are overprinted by east-west trending major and minor folds, forming major dome and basin structures (Pyke, 1982). The major isoclinal folds with east-west striking subvertical axial planes, steeply-plunging minor folds, subvertical axial plane foliation, and steeply plunging stretching lineation were probably formed under subhorizontal, generally north-south major compression. Toward the southern margin of Abitibi belt the major folds are overturned northward, and in the adjacent Pontiac subprovince, folds are recumbent. The Cadillac-Larder Lake fault zone, which constitutes the boundary between the Abitibi and Pontiac subprovinces, dips 45°N to 60°N and probably has both sinistral transcurrent and thrust components of movement.

Several suites of intrusive rocks in the Abitibi subprovince can be distinguished on the basis of composition, structural relationships, setting, and age (Card, 1982). The oldest suite includes synvolcanic sills, dykes and plutons ranging in composition from peridotite to granodiorite; the more felsic intrusions are typically quartz diorite and trondhjemite. Gneissic plutonic rocks of tonalite and granodiorite composition, commonly containing amphibolitic enclaves, occur in the northeastern and southwestern Abitibi subprovince. Massive felsic plutonic rocks intrude both the greenstones and the gneissic rocks in the form of simple and composite plutons and batholiths. They form several suites, including early granodiorites, younger granite batholiths, and still younger syenite-diorite plutons. Contacts between the plutons and the country rocks are commonly concordant and steeply dipping; dominant east-west structural trends are locally deflected about the intrusions.

A time framework for events in the Michipicoten and Abitibi belts can be constructed from U-Pb zircon dates. In the western Abitibi belt, volcanic rocks range in

age from 2,725 to 2,703 Ma (Nunes and Pyke, 1980; Nunes and Jensen, 1980) and in the Michipicoten belt, from 2,749 to 2,696 Ma, with synvolcanic plutons at 2,744 and 2,737 Ma (Turek et al., 1982). A number of late- to post-tectonic plutons from the Abitibi and Michipicoten belts have zircon dates within a few million years of 2,680 (Krogh et al., 1982). In the Wawa domal terrane, tonalite gneiss has a minimum age of 2,707 Ma, partly reset by intrusion of granodiorite at 2,677 Ma (Percival and Krogh, 1983; Fig. 5). Thus the Abitibi and Michipicoten supracrustal sequences and early intrusions developed between 2,750 and 2,700 Ma ago. The dates on volcanics and late plutons bracket the age of deformation and regional metamorphism at between 2,700 and 2,680 Ma ago. Major volcanic, plutonic, and tectonic events of relatively brief duration were essentially synchronous throughout the Abitibi and Wawa subprovinces, a region some 1,200 km long and 200 km wide. The lithologic and age similarities between the Abitibi and Wawa subprovinces strongly suggest original continuity, now interrupted by the Kapuskasing structural zone.

Diabase dyke swarms of late Archean and Proterozoic age are present throughout the region. The oldest dykes, the north-trending Matachewan swarm of the Abitibi subprovince, have a Rb-Sr age of 2633 Ma (Gates and Hurley, 1973). Northwest-striking diabase dykes in Wawa subprovince are petrographically similar to and have been paleomagnetically correlated with the Matachewan swarm (Ernst, 1981). Abitibi and Wawa subprovinces are thus inferred to have been tectonically stable cratons by this time. Northeast-striking tholeiitic dykes are about 2105 Ma old (Gates and Hurley, 1973); northwest olivine diabase dykes (Sudbury swarm) are about 1250 Ma old (Van Schmus, 1975); and east-northeast olivine diabase dykes (Abitibi swarm) are approximately 1100 Ma old (Lowden and Wanless, 1963).

## Kapuskasing Structural Zone

The Kapuskasing structural zone comprises northeast-striking, northwest-dipping belts of paragneiss, mafic gneiss, gneissic and xenolithic tonalite, and rocks of the Shawmere anorthosite complex (Bennett et al., 1967; Thurston et al., 1977) (Figs. 2 and 5).

Migmatitic paragneiss is compositionally layered with garnet, biotite, quartz-rich and rare graphitic varieties. Concordant tonalitic leucosome constitutes up to 20 per cent of many outcrops. Enclaves and layers of mafic gneiss in paragneiss occur on the 10 cm to 1 km scale. Migmatitic mafic gneiss is characterized by garnet-clinopyroxene-hornblende-plagioclase-quartz-ilmenite$\pm$orthopyroxene mineral assemblages and generally contains concordant tonalitic leucosome. Layering, on the 1 to 10 cm scale, is produced by variable proportions of minerals. Table 1 presents two sets of whole-rock analyses from adjacent anhydrous (garnet-clinopyroxene-plagioclase-quartz) and hornblende-bearing layers from mafic gneiss in two different locations. From the analyses it is unclear whether the layering is a preserved compositional heterogeneity or a product of metamorphic differentiation. The bulk composition corresponds to high calcium (10-15 wt% CaO), high alumina (13.4-17.2 wt% $Al_2O_3$) basalt (Table 1). Nickel and chromium abundances of mafic gneiss are in the 95-220 and 12-190 ppm ranges respectively and are not definitive in distinguishing between basaltic igneous and marly sedimentary parentage for the rock type.

In the area of Figure 5, four linear, northeast-striking bodies of flaser-textured to foliated diorite and mafic tonalite occur dominantly within paragneiss terranes. These medium- to coarse-grained, locally migmatitic rocks consist of hornblende, biotite and plagioclase, with up to 10 per cent quartz as well as orthopyroxene, clinopyroxene and rare garnet. Gabbro, hornblendite and rare pyroxenite occur locally as layers 10 cm to 2 m thick, generally within 2 km of paragneiss contacts.

| | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| $SiO_2$ | 47.8 | 46.6 | 52.5 | 43.1 | 47.8 | 49.9 |
| $TiO_2$ | 0.81 | 0.81 | 1.81 | 1.59 | 1.0 | 1.3 |
| $Al_2O_3$ | 15.5 | 15.6 | 17.2 | 13.4 | 16.2 | 17.0 |
| $Fe_2O_3$ | 1.3 | 2.2 | 2.2 | 5.7 | 3.4 | 1.5 |
| FeO | 9.1 | 9.4 | 8.5 | 12.8 | 8.5 | 7.6 |
| MnO | 0.27 | 0.19 | 0.32 | 0.3 | 0.32 | 0.2 |
| MgO | 4.53 | 5.29 | 3.64 | 9.25 | 5.41 | 8.2 |
| CaO | 15.4 | 14.2 | 11.2 | 10.0 | 13.50 | 11.4 |
| $Na_2O$ | 2.0 | 2.4 | 2.8 | 1.6 | 2.3 | 2.8 |
| $K_2O$ | 0.25 | 0.41 | 0.12 | 0.58 | 0.33 | 0.2 |
| $H_2O$ | 0.5 | 1.1 | 0.3 | 1.6 | 0.8 | |
| $CO_2$ | 2.3 | 2.0 | 0.4 | 0.1 | 0.6 | |
| Ni | 0.014 | 0.014 | 0.0095 | 0.0098 | 0.024 | |
| Cr | 0.019 | 0.018 | 0.018 | 0.014 | 0.015 | |
| Total | 100.0 | 100.4 | 100.6 | 100.4 | 100.2 | 100.1 |

### CIPW Norm

| | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| QZ | 1.5 | | 6.6 | | | |
| OR | 1.49 | 2.44 | 0.71 | 3.47 | 1.95 | 1.0 |
| AB | 17.02 | 20.46 | 23.63 | 13.72 | 19.8 | 23.5 |
| AN | 32.77 | 30.81 | 33.92 | 28.03 | 32.97 | 33.4 |
| DI | 11.42 | 11.52 | 6.99 | 10.20 | 13.77 | 18.9 |
| HE | 12.80 | 10.90 | 8.28 | 6.41 | 10.58 | |
| EN | 6.05 | 3.64 | 5.80 | 5.39 | 4.96 | 9.4 |
| FS | 7.83 | 3.95 | 7.89 | 3.89 | 4.43 | |
| FO | | 3.01 | | 9.26 | 1.54 | 9.3 |
| FA | | 3.60 | | 7.36 | 1.21 | |
| MT | 1.9 | 3.21 | 3.18 | 8.37 | 4.95 | 2.2 |
| IL | 1.55 | 1.55 | 2.24 | 3.06 | 1.90 | 2.5 |
| AP | 0.12 | 0.14 | 0.26 | 0.24 | 0.19 | |
| CC | 5.26 | 4.58 | 0.91 | 0.46 | 1.44 | |

Table 1: Whole rock chemical analyses of mafic gneiss from the Kapuskasing zone, with CIPW norms. Analyst: R. Charbonneau, GSC Lab. 1: granulite layer, P79-475 (Gt-Cpx-Pl-Qz, 5% Hb); 2: amphibolite layer, P-475 (Gt-Cpx-Pl-Qz, 25% Hb); 3: granulite layer, P79-371 (Gt-Cpx-Pl, tr Qz); 4: amphibolite layer, P79-371 (Hb 40%, Gt 15%, Cpx 15%, Pl 20%); 5: average of three mafic gneisses from the KSZ (79-84A, 123, 299); 6: high-alumina basalt (Ringwood, 1975).

Discrete belts of xenolithic and gneissic tonalite are present south of the main body of the Shawmere anorthosite complex and small bodies are present to the north. The southern belt is made up of coarse garnet-hornblende-biotite-plagioclase-quartz tonalite containing enclaves of mafic gneiss, paragneiss, hornblendite and garnet-orthopyroxene-hornblende-biotite rocks. Southwest along this belt, garnet decreases in abundance and the composition is granodioritic. Inclusions in this area are amphibolite, hornblendite, and cummingtonite-hornblende-biotite rocks.

The Shawmere anorthosite complex (Thurston et al., 1977) consists of a main northern body, 15 x 50 km and a smaller mass, measuring 5 x 15 km. The bodies taper to the northeast and southwest and thus have concordant contacts. Gneissic textures prevail in the outer portions of the main body, whereas primary igneous minerals and textures are preserved in the interior (Simmons et al., 1980). The main body comprises four distinct lithological-textural units (Riccio, 1981; Fig. 6): (1) a border zone of migmatitic, foliated to gneissic garnetiferous amphibolite, (2) a banded zone consisting of 1 to 30 cm-thick layers of anorthosite, gabbro, garnet-rich, and ultramafic rock, (3) an anorthosite zone containing minor gabbro and (4) a megacrystic gabbroic anorthosite zone with plagioclase phenocrysts to 50 cm and minor anorthosite, anorthositic gabbro, gabbro and melagabbro. A 1 km wide body of foliated garnetiferous tonalite is present within the outcrop area of the anorthosite. Its genetic relationship to the anorthosite complex is not clear although it appears to be temporally related (Simmons et al., 1980). The southern body consists dominantly of coarse gabbroic anorthosite.

The orientation of gneissosity and lithological contacts make up the prominent east-northeast structural grain of the Kapuskasing structural zone. Gneissosity in all rock types is folded or warped about gently-plunging (0-25°) northeast-trending axes. The folds vary from isoclinal with consistent "Z" sense asymmetry when viewed toward the east to northwest-facing monoclinal flexures. Axial surfaces are rarely accompanied
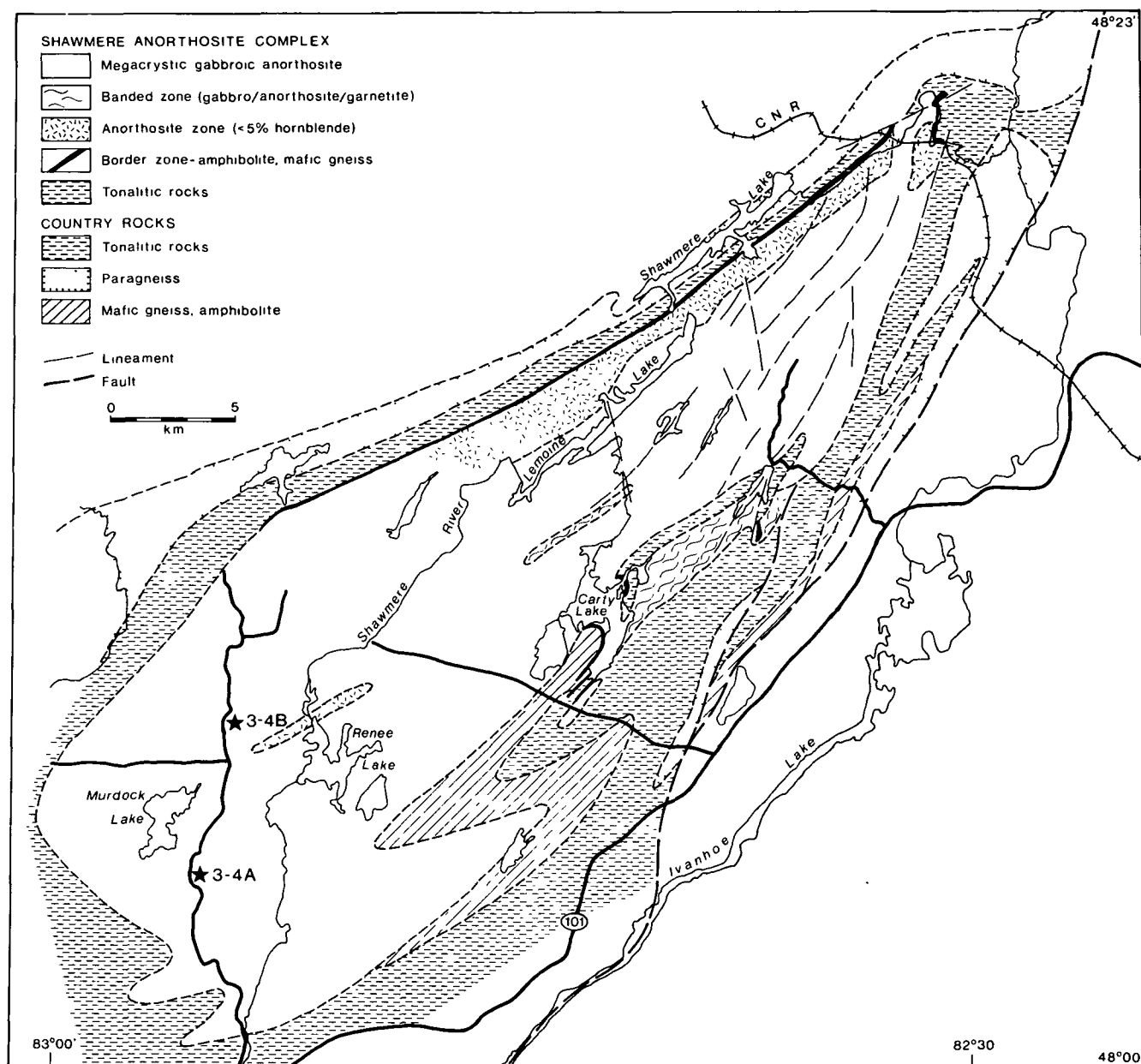
Fig. 6.    Geology of the Shawmere anorthosite complex (after Riccio, 1980 and
Percival, 1981).

by a foliation defined by flattened quartz grains. The trend of lineations and fold axes is northeast-southwest throughout this part of the Kapuskasing zone, but plunge direction varies on a regional scale from dominantly southeasterly in the south to northeasterly in the north. Between these areas, lineations are within 10° of horizontal and abrupt changes in plunge direction occur on the 10 m scale. Both regional and local plunge reversals can be related to a gently southeast-plunging warp axis.

Two high-grade metamorphic zones can be distinguished in this part of the Kapuskasing structural zone. Assemblages characteristic of a lower-grade garnet-clinopyroxene-plagioclase zone are developed in mafic gneiss. Orthopyroxene, present in four areas in most rock types, is diagnostic of a higher-grade orthopyroxene zone (Fig. 7; Percival, 1983).

A continuous reaction resulting in decomposition of hornblende in mafic rocks to produce garnet and clinopyroxene may be written:

$$\text{hornblende} + \text{plagioclase} \rightleftharpoons \text{garnet} + \text{clinopyroxene} + \text{quartz} + H_2O \qquad (1)$$

The coexistence over large areas of this divariant assemblage and tonalitic leucosome veinlets suggests that the reaction was anatectic and also produced a liquid over a range of P-T conditions (Fig. 8):

$$\text{hornblende} + \text{plagioclase} \rightleftharpoons \text{garnet} + \text{clinopyroxene} + \text{tonalite} \qquad (2)$$

A possible reaction leading to the production of orthopyroxene in mafic rocks is:

$$\text{hornblende} + \text{garnet} \rightleftharpoons \text{orthopyroxene} + \text{clinopyroxene} + H_2O \qquad (3)$$
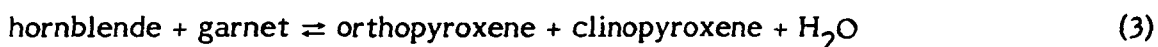
The evolved water would presumably have been taken up by anatectic liquids.

In paragneiss, a reaction producing orthopyroxene in the presence of anatectic melt is:

$$\text{biotite} + \text{quartz} + \text{plagioclase} \rightleftharpoons \text{orthopyroxene} + \text{granodioritic liquid} \qquad (4)$$
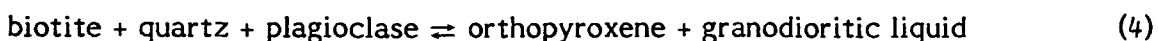
A P-T diagram summarizing continuous reactions in the mafic system and apparent metamorphic conditions based on various mineral geothermometers and
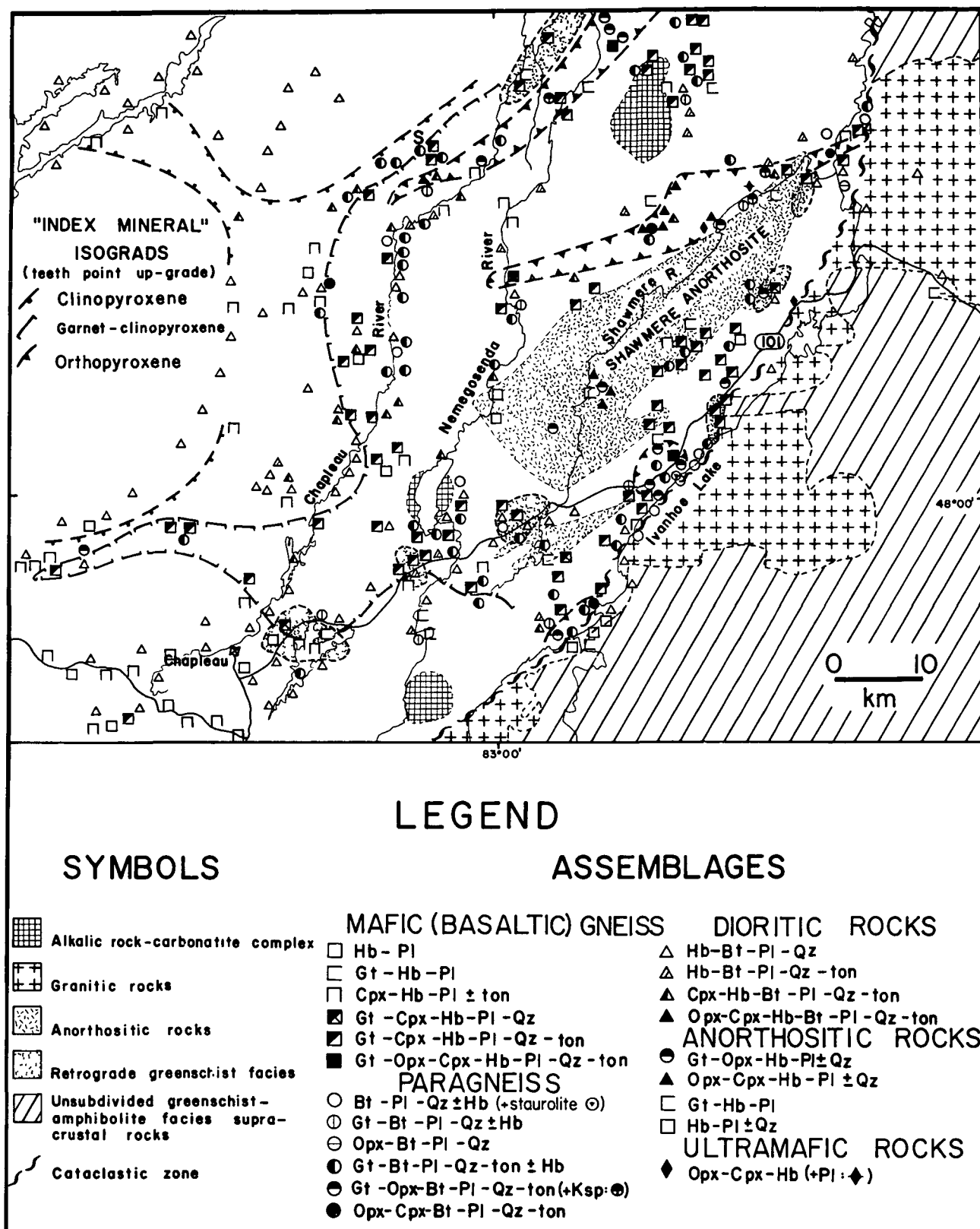
Fig. 7.   Metamorphic mineral assemblages and index mineral isograds for part of the
Chapleau-Foleyet area.  Gt - garnet; Opx - orthopyroxene; Cpx -
clinopyroxene; Hb - hornblende; Bt - biotite; Pl - plagiocase; Ksp - feldspar;
Qz - quartz; ton - tonalitic segregations.

geobarometers, is presented in Figure 8. Apparent pressures, based on Newton and Perkins' (1982) garnet-clinopyroxene-plagioclase quartz barometer, are plotted on a map in Figure 9 and have an average value of 6.3 kbar. Apparent temperatures, based on the Ellis and Green (1979) garnet-clinopyroxene thermometer (Fig. 8) are in the range 700-800°C.

The assemblage almandine garnet-clinopyroxene-plagioclase-quartz is diagnostic of the regional hypersthene zone according to Winkler (1979, p. 260, 267-268). de Waard (1965) and Green and Ringwood (1967) suggeted that this assemblage forms as an alternative to orthopyroxene-plagioclase during high-pressure granulite-facies metamorphism. Turner (1981) attaches a different significance to the assemblage, regarding it as transitional from amphibolite to granulite facies based on Binns' (1964) study. In the present study area, the location of the garnet-clinopyroxene-plagioclase zone between hornblende-plagioclase+clinopyroxene rocks and orthopyroxene-bearing rocks suggests that it characterizes the amphibolite-granulite facies transition. Although the assemblage is the same as that in the Adirondacks (de Waard, 1965) and temperature conditions were similar (cf. Bohlen and Essene, 1977), the path of metamorphism was different. In the Grenville Province, the development of garnet-clinopyroxene assemblages has been attributed to isobaric cooling of orthopyroxene-plagioclase granulites (Martignole and Schrijver, 1971; Whitney, 1978) whereas in the Kapuskasing zone, garnet and clinopyroxene formed during prograde reactions.

Rounded zircons of probable metamorphic origin from Kapuskasing mafic gneiss gave a concordant date of 2,650 Ma and from a leucosome layer in paragneiss of 2,627 Ma (Percival and Krogh, 1983; Fig. 10). A minimum age of emplacement for foliated tonalite from the Shawmere complex is provided by zircons (2,765 Ma) but the U-Pb system has been strongly affected by the high-grade metamorphism (Percival and Krogh, 1983). The rocks intruded by the tonalite are thus older than dated volcanic rocks of the Abitibi and Michipicoten belts.

Fig. 8.    Summary of reactions applicable to mafic rocks and metamorphic pressure-temperature estimates. Temperatures are derived from the garnet-clinopyroxene thermometer (Ellis and Green, 1979) and pressures from garnet-pyroxene-plagioclase-quartz barometers (Newton and Perkins, 1982).
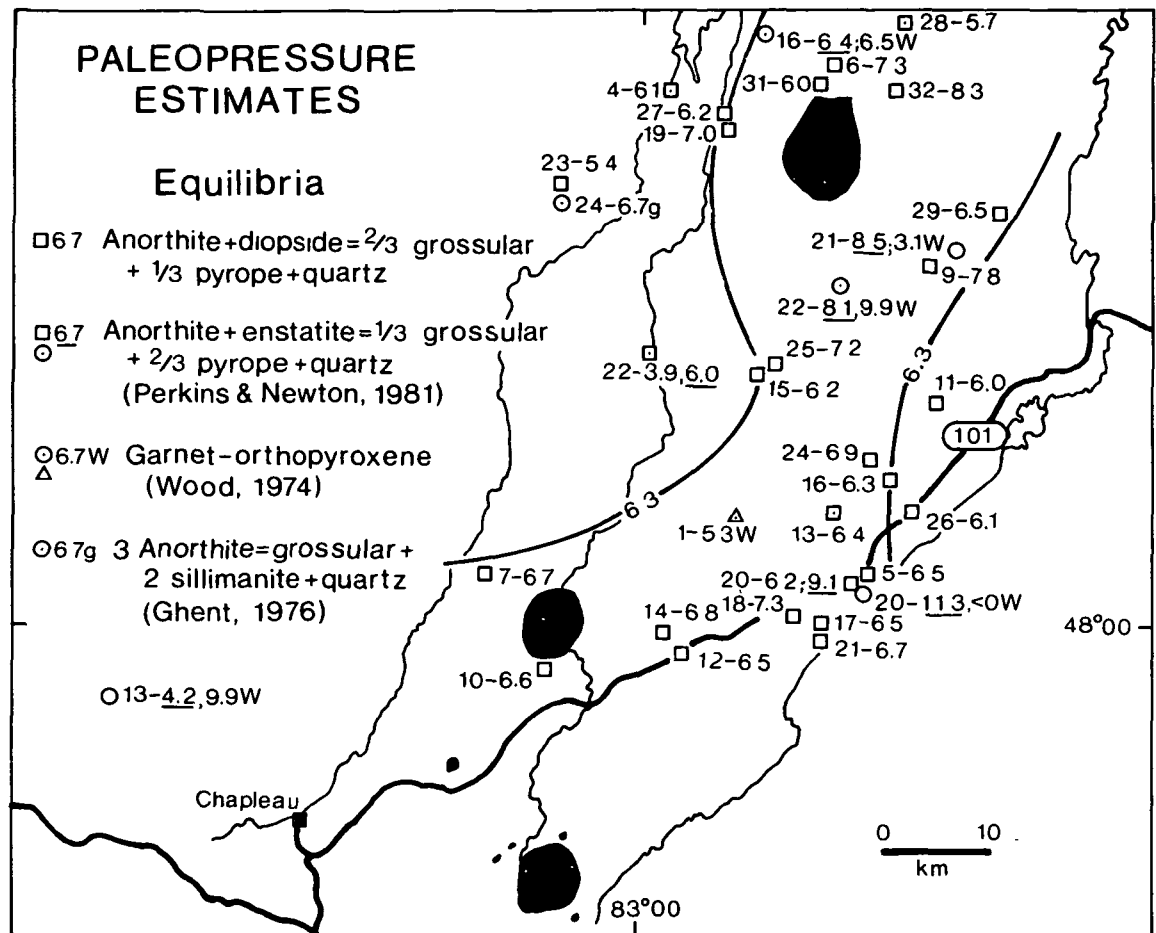
**Fig. 9.** Paleopressure map of the Chapleau-Foleyet area. Symbols represent rock type (circles - paragneiss; squares - mafic gneiss; triangles - orthogneiss). Numbers to the right of the dash are pressure estimaters (kbar) keyed to the equilibrium used to derive the value. The 6.3 kbar reference line is based on garnet-clinopyroxene-plagioclase-quartz equilibrium.

At least two swarms of fresh mafic dykes transect metamorphic rocks of the Kapuskasing zone. East-northeast-striking, southeast-dipping Kapuskasing dykes are 1 to 10 m wide, sparsely plagioclase porphyritic, medium- to fine-grained, ophitic, green-grey gabbro. Northeast-trending olivine-bearing dykes may belong to the Abitibi swarm.

Fig. 10.   Concordia diagram with isotopic ratios of zircon samples. Ab: abraded
(Krogh, 1982); N: non-magnetic (Frantz); Ml: magnetic at 1° side tilt
(Frantz); pr: prismatic; an: anhedral.  Solid lines are empirical lead-loss
trajectories; dashed line: hypothetical lead-loss trajectory (projects to 0 Ma
lower intercept); dash-dot line: empirical mixing line.

Several small alkalic rock-carbonatite complexes are associated with the Kapuskasing zone. The more northerly bodies have K-Ar dates of 1655 to 1720 Ma, whereas those in the south have dates of 1050 to 1100 Ma (Gittins et al., 1967). Thin lamprophyre dykes and a rare diatreme breccia are associated with the complexes; biotite from a lamprophyre dyke in the Chapleau-Foleyet area gave a K-Ar date of 1144 $\pm$ 31 Ma (Stevens et al., 1982).

## Relationship of Kapuskasing Structural Zone to Adjacent Subprovinces

The contact between the Kapuskasing structure and Abitibi subprovince is a zone of faulting and cataclasis, the Ivanhoe Lake cataclastic zone, that separates the two terranes of contrasting lithological, structural, and metamorphic characteristics. The contact is defined in part by a positive, linear north-northeast aeromagnetic anomaly and coincides with the trough of a paired high (Kapuskasing) - low (Abitibi) gravity anomaly (Figs. 4 and 11).

The Ivanhoe Lake cataclastic zone is characterized by narrow veinlets of finely comminuted rock which form discontinuous, randomly-oriented pods and networks. Two types of fault rocks can be distinguished. The first is foliated to massive, semi-opaque mylonite, cataclasite and blastomylonite, partly or totally recrystallized to fine grained epidote, chlorite, carbonate, and actinolite. The second type grades from cataclasite to pseudotachylite with aphanitic, almost opaque matrix and rounded, embayed monomineralic porphyroclasts.

The dip of the Ivanhoe Lake cataclastic zone is not well constrained geologically. Although some fault-rock veinlets are parallel to gneissosity and therefore dip gently northwest, many others have random orientation. The juxtaposition of high-grade against low-grade rocks indicates reverse displacement across the cataclastic zone. The associated paired gravity anomaly is characteristic of many well-documented overthrust terranes (Smithson et al., 1978; Fountain and Salisbury, 1981) and suggests that the
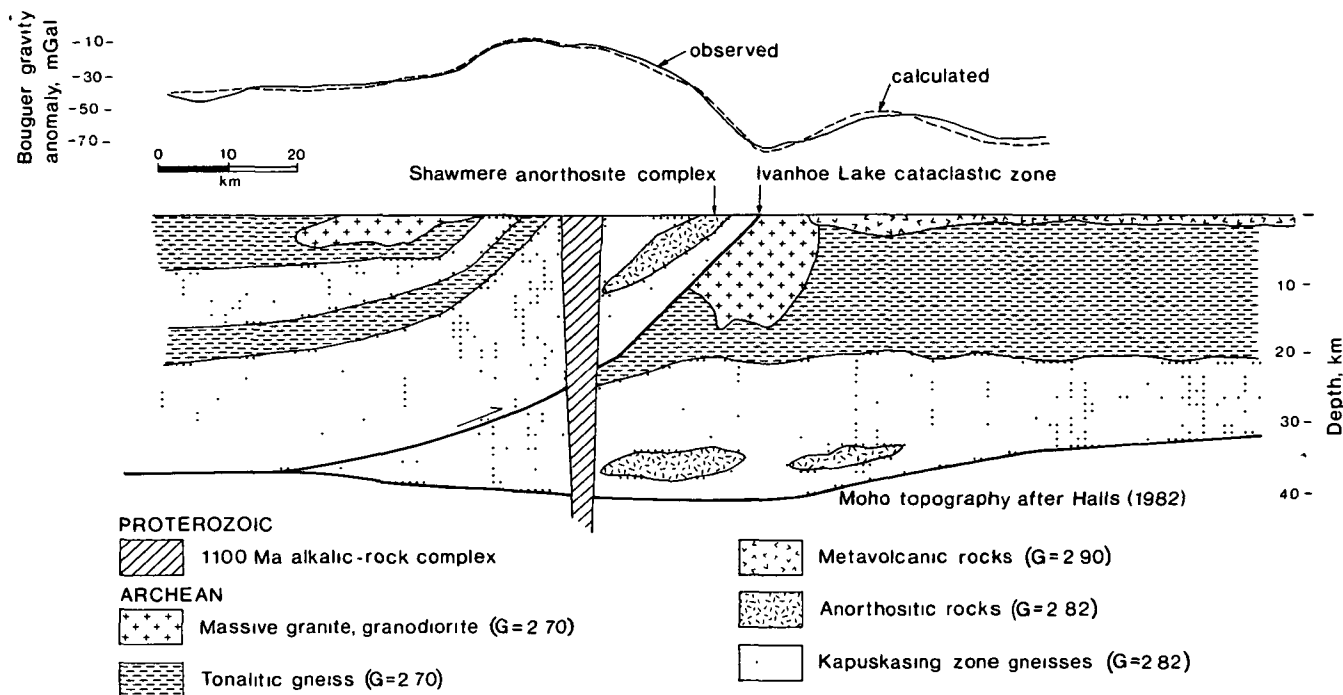
Fig. 11.     Generalized west-east cross-section from the Wawa domal gneiss terrane, through the Kapuskasing structural zone into the Abitibi subprovince, showing gross crustal structure. The gravity model based on the average rock densities: tonalitic gneiss and granite: 2.70; metavolcanics: 2.90; Kapuskasing structural zone and lower crust: 2.82 g/cm 3.

Ivanhoe Lake cataclastic zone is the surface expression of a northwest-dipping thrust fault (Fig. 11).

The Wawa-Kapuskasing boundary is gradational in lithological, structural and metamorphic characteristics. Mafic gneiss with minor paragneiss is typical of the Kapuskasing zone but also occurs in the Robson Lake dome with characteristic structural style of the Wawa subprovince. Garnet-clinopyroxene-hornblende-plagioclase assemblages are common here, suggesting that the metamorphic grade is similar to that in the Kapuskasing structural zone. The discontinuous paragneiss belt that extends for

up to 30 km into the Wawa subprovince may also be a part of the Kapuskasing lithological sequence. Tonalitic gneiss can be traced eastward from the Borden Lake area, where it has the complex structures characteristic of the Wawa subprovince, into strongly foliated and lineated gneiss typical of the Kapuskasing zone.

The change in structural style from domal in the eastern Wawa subprovince to linear ENE belts in the Kapuskasing structural zone can be used to define a transitional boundary zone, but no sharp line can be drawn on this basis. South of Chapleau, the orientation of gneissic layering changes eastward from horizontal near the Highbrush Lake dome, through a zone with a superimposed upright easterly foliation, to strong northeast-striking, northwest-dipping gneissosity. A north-south-trending structural culmination coincides with the eastern domes of the Wawa subprovince. East of the culmination, lineations plunge easterly toward a structural depression into which southwest-trending lineations of the southern Kapuskasing zone also plunge. To the north, lineations plunging northeasterly off the northeastern flank of the Missinaibi Lake arch appear to be continuous with northeast-plunging, reclined folds in the northern Kapuskasing structural zone. Cataclastic veinlets are sporadically present along mafic gneiss-tonalitic gneiss contacts for approximately 20 km southwest of Kapuskasing Lake. A fault zone marks the western limit of garnet-clinopyroxene-hornblende-plagioclase-quartz assemblages and mafic gneiss, but structural trends are continuous across it. The gradational nature of lithological contacts as well as the structural and metamorphic continuity between tonalites and high-grade gneisses suggests that the contacts were etablished prior to metamorphism and doming, and that rock units of the Kapuskasing zone locally occur structurally below the Wawa tonalite-granodiorite gneiss.

Structure of the Kapuskasing Crustal Cross-Section

The transition from the Michipicoten belt to the eastern boundary of the Kapuskasing zone can be interpreted as an oblique crustal cross-section based on the following: 1) metamorphic grade increases eastward from low greenschist facies in the

Michipicoten belt through amphibolite facies in the Wawa domal gneiss terrane to upper

amphibolite and granulite facies in the Kapuskasing zone; 2) the proportion of plutonic to

supracrustal rocks increases eastward in the Wawa subprovince; 3) the oldest rocks

($>2,765$ Ma) are in the Kapuskasing zone at the inferred base of the section; 4) the

gravity anomaly can be best modelled by using a west-dipping crustal slab (Fig. 11).

Construction of a generalized crustal cross-section (Fig. 12) requires several

assumptions: 1) the dip of the crustal slab is constant; 2) pressure is a function of depth

so that estimates of metamorphic pressure can be used to derive the thickness of the

section; 3) the metamorphic assemblages are the product of a single metamorphic event;

and 4) post-metamorphic vertical displacement on faults within the section is negligible.

The highest-grade assemblage from the Wawa area is garnet-andalusite in

metagreywacke (Ayres, 1969), indicating a maximum pressure of 3.3 kb and a depth of

about 11 km (Carmichael, 1978). The range of pressures estimated from the Kapuskasing

zone, based on Newton and Perkins' (1982) garnet-clinopyroxene-plagioclase-quartz

barometer, is 5.4 to 8.4 kb (average of 6.3 kb, Percival, 1983) but the lower values may

result from re-equilibration during cooling. These values correspond to depths of 18 to

28 km (average 21 km). The minimum erosion-level difference is therefore 7 km, but the

difference is probably closer to 15 km. The minimum and maximum dip estimates over a

constantly-dipping slab 120 km long are approximately 5° and 10°.

The dips of post-metamorphic dykes in the Kapuskasing zone and eastern Wawa

subprovince may provide an independent estimate of the tilt of the slab in this area.

Matachewan dykes dip NE at 75° to 85° and ENE Kapuskasing dykes dip SE at 70° to 85°

based on measurements of dykes with vertical exposure in roadcuts. Post-metamorphic

mafic dykes in the Shield generally have near-vertical orientations, as do Matachewan

dykes in the Abitibi subprovince (Thurston et al., 1977; Milne, 1972). The consistent non-

vertical dip may thus have resulted from large-scale crustal rotation. To restore the

dykes of both swarms to vertical, a 14° counter-clockwise rotation about an axis
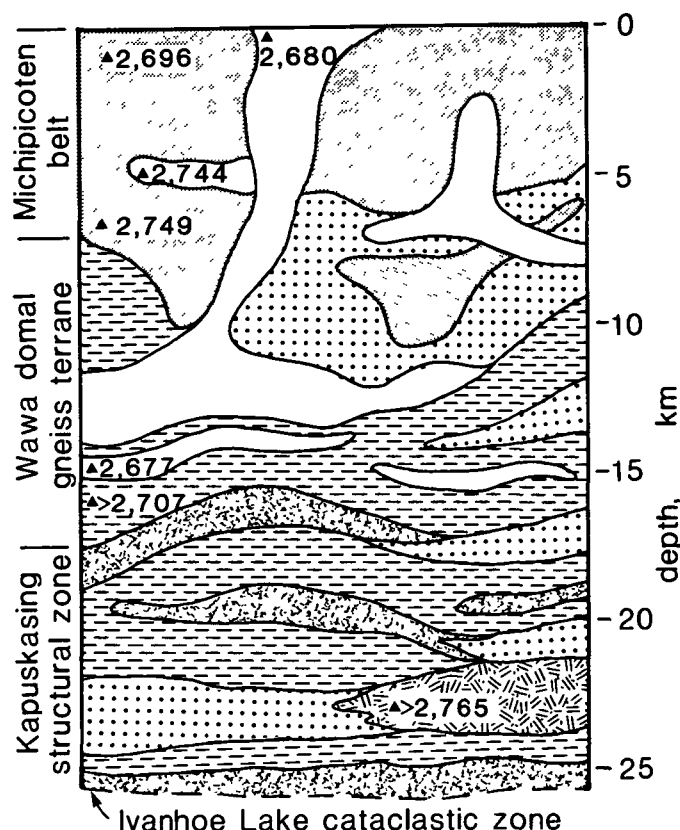
Fig. 12. Restored vertical section through the Michipicoten belt, domal gneiss terrane and Kapuskasing zone. Numbers are zircon dates ($\pm$4 Ma) on igneous and meta-igneous rocks.

trending 038° is necessary. Thus a 14° northwesterly dip is indicated in this eastern area. The difference in dip estimate provided by these two methods may be due to uncertainties in the data used in the calculations, faulty assumptions, or real differences in dip from east to west. The overall dip must flatten to the northwest and may in fact be reversed northwest of the Michipicoten belt. Ernst (1981, p. 87) reported consistent 85° SW dips of Matachewan dykes northwest of Wawa. Therefore, an intermediate dip value of 10° perpendicular to the fault was chosen for construction of Figure 12. If dips

flatten toward the northwest, this will result in over-estimation of the true thickness of the section.

The generalized section is a valid representation provided that (1) a single regional metamorphic event affected all of these rocks, and (2) late vertical displacement along faults is negligible between the Kapuskasing zone and western Wawa subprovince. In view of the complex relationships described and uncertainties involved, these simplications may be unwarranted; however, the information which can be derived from an exposed cross-section through part of the crust is potentially valuable enough to permit some speculation.

The generalized crustal cross-section, constructed using a dip of 10° (Fig. 12), has at its base a sequence of upper amphibolite to granulite facies gneiss and anorthosite, the full thickness of which is unknown, and of which some 5 to 10 km is exposed in the Kapuskasing zone. Structurally above is 10 to 15 km of tabular batholiths of gneissic and xenolithic tonalite. Massive granitic rocks occur as sheets and deep-rooted plugs at this structural level. In the upper 5-10 km, both granitic rocks and gneissic migmatitic haloes surround the low-grade Michipicoten belt. The interfaces between the adjacent, generally horizontal megalayers are undulating surfaces with several kilometres of relief, manifest as gneiss domes at intermediate structural levels and as intrusive bodies at higher levels.

In the western Superior Province, two seismic discontinuities at 16-19 and 21-22 km, define upper, middle and lower crust (Hall and Brisbin, 1982). Using the Kapuskasing model, the upper discontinuity corresponds to the boundary between a structurally higher granitoid gneissic layer and a subjacent heterogeneous high-grade gneiss complex, whereas the lower discontinuity, corresponding to the middle-lower crustal boundary, is probably a metamorphic isograd (orthopyroxene isograd?) within the heterogeneous gneiss.

Similar models of mega-layered continental crust are based on seismic and gravity data (Smithson and Brown, 1977; Berry and Mair, 1980). Other inferred cross-sections through the crust (Ivrea zone, Pikwitonei region, Musgrave, Fraser ranges; Fountain and Salisbury, 1981) have in common a downward increasing metamorphic grade and a thick, intermediate-depth amphibolite-facies section of quartzofeldspathic gneiss, corresponding to the domal gneiss terrane of the Wawa subprovince. In the central Superior Province section, these gneisses intrude and assimilate both the overlying supracrustal succession and parts of the underlying complex. The entire section down to ~ 20 km was added to the crust in the interval between 2750 and 2680 Ma. The pre-existing crust may have, but need not have been as thick as present continental crust prior to the major thickening event. The high metamorphic grade in this older crust can be accounted for by burial, first by a volcanic pile and somewhat later by intrusion of tonalite sheets.

## Archean Evolution of the Kapuskasing Crustal Structure

The oldest rocks so far recognized, paragneiss and mafic gneiss of the Kapuskasing zone, are considered part of a sedimentary-volcanic succession deposited prior to 2765 Ma ago. The Shawmere anorthosite was emplaced into this succession, probably also prior to 2765 Ma ago and probably as a stratiform body at depths of less than 20 km, as inferred from the presence of relict olivine (Thurston et al., 1977; Kushiro and Yoder, 1966). As suggested by Simmons et al. (1980), the intrusion may represent the differentiation product of tholeiitic basalt magmas which also erupted at surface.

Major eruption of volcanic rocks and deposition of sediments occurred between 2749 and 2696 Ma ago in the Michipicoten belt (Turek et al., 1982) and between 2725 and 2703 Ma ago in the western Abitibi belt (Nunes and Pyke, 1980). The lowermost volcanics are generally mafic and so have not been dated by the U-Pb zircon method.

Synvolcanic intrusions, including ultramafic, mafic, and trondhjemitic to granodioritic bodies, were intruded into the Michipicoten and Abitibi piles 2750 to

2700 Ma ago. Large volumes of tonalite intruded beneath and adjacent to the greenstone belts at this time. The minimum age of 2707 Ma for Wawa tonalite (Fig. 10) is given by a nearly concordant point and is therefore probably close to the true age. The tonalites could be the subsurface expression of magmas that produced dacites in the upper parts of the volcanic piles. Tonalite intrusions, now gneissic, engulfed and detached fragments of the lower parts of the greenstone succession (now represented as mafic xenolith trains), possible older, tonalite basement enclaves (e.g. Hillary and Ayres, 1980), and the western parts of the Kapuskasing zone which extend into the tonalite gneiss terrane. The tonalitic magmas may represent juvenile magmas derived from the mantle, or may be the products of partial melting of a heterogeneous lower crust similar to that exposed in the Kapuskasing zone. Heat from the tonalitic intrusions was probably sufficient to cause the metamorphism of the volcanics. Tonalitic magmatism thus may have coincided with regional metamorphism and acted as the main agent of heat transfer into the upper crust. Isoclinally folded gneissosity in the tonalite demonstrates that major deformation post-dates these intrusions.

The age of major deformation in the Abitibi and Wawa subprovinces is closely bracketed between 2696 Ma, the approximate age of the youngest volcanics, and 2680 Ma, the approximate age of late- to post-tectonic plutons (Krogh et al., 1982). In supracrustal rocks at high crustal levels, this deformation produced upright to vertically-plunging structural features as well as thrusts and nappe-like structures (Poulsen et al., 1981; Gorman et al., 1978; Thurston and Breaks, 1978). At deeper structural levels, the deformation resulted in gneissosity and subsequent folds in plutonic rock and paragneiss, followed by later doming. Forceful emplacement of massive plutons also deflected structural trends in country rock into concordance with the margins of these bodies. Following intrusion of the massive plutons at 2680 Ma, there was tectonic quiesence in Abitibi and Wawa subprovinces. There is evidence, however, of continued activity in the Kapuskasing zone.

High-grade metamorphic rocks of the Kapuskasing zone yield concordant U-Pb zircon dates of 2650 to 2627 Ma. U-Pb zircon dates are generally considered to record the age of crystallization of the zircons, which in this case are of metamorphic origin. This interpretation would imply that metamorphism in the Kapuskasing zone occurred 2650 to 2627 Ma ago, 25 to 50 Ma after tectonic stabilization of much of the rest of Superior province. A discrete burial and metamorphism event, restricted to the Kapuskasing zone, could explain the deformed metamorphosed conglomerate cobbles from Borden Lake which have a zircon date of 2664 Ma (Percival et al., 1981). However, tectonic mechanisms which could lead to deep burial of the 500 km long x <50 km wide Kapuskasing "sliver" are unknown and seem to be unlikely after termination of the major tectonism in the Abitibi and Wawa subprovinces. This interpretation, that a second metamorphic event affected the area, is based entirely on geochronological results. Without these dates, a single metamorphic event would be adequate to explain the observed relationships.

One must therefore examine the assumption that zircons are closed to lead loss immediately following crystallization, regardless of the cooling history. Slowly decreasing metamorphic temperatures from peak levels of >800°C could result in lead diffusion out of zircon for several million years after crystallization, provided that there is some finite "blocking temperature" for zircon. A value of 700 ± 50°C was estimated for zircon blocking by Mattinson (1978). Invoking this hypothesis to explain the young "metamorphic" dates would allow a simpler geological history involving only the metamorphism at 2700 to 2680 Ma with slightly later ductile shear at depth.

The prominent east-northeast structural trends in the Kapuskasing zone are the result of relatively late tectonism. The structural grain is defined by the orientation of migmatitic and gneissic layering folded about shallow ENE axes. This folding event therefore post-dates crystallization of tonalitic melts, thought to coincide with the metamorphic peak. Similarly, structurally complex tonalitic gneiss units that can be

traced from Wawa subprovince into the Kapuskasing zone have a strong, superimposed ENE foliation and lineation in the Kapuskasing zone (Percival and Coe, 1981). If the correlation is valid between massive granodiorite dated at 2680 ± 3 Ma and granodiorite gneiss adjacent to the Kapuskasing zone in the Abitibi subprovince, then the ductile strain occurred after 2680 Ma but pre-dated post-metamorphic cooling. This timing is consistent with the suggestion of Watson (1980) that sinistral transcurrent movement occurred along the Kapuskasing zone during emplacement of the Matachewan dyke swarm at 2633 Ma (Gates and Hurley, 1973). Late Archean deformation could have resulted in the resetting of 2700-2680 Ma zircons to ages in the range 2650-2627 Ma.

## Uplift of the Kapuskasing Structure

The age of uplift of the Kapuskasing zone is not well constrained. Evidence of late Archean transcurrent movement was cited by Watson (1980) and Percival and Coe (1980), however its magnitude was probably small, judging by the minor apparent offset of the Abitibi-Opatica contact (Fig. 1). Major thrusting could also have occurred at that time, setting U-Pb and K-Ar isotopic systems in the high-grade rocks at 2,650-2,445 Ma.

Geochronological evidence indicates activity at 1,655-1,850 Ma. Three alkalic rock-carbonatite complexes near Kapuskasing have K-Ar dates of 1,655-1,720 Ma (Gittins et al., 1967). A biotite-whole-rock Rb-Sr isochron from tonalite of the Shawmere anorthosite complex is 1,850 Ma (Simmons and others, 1980). A whole-rock $^{40}Ar/^{39}Ar$ analysis of blastomylonite from the Ivanhoe Lake cataclastic zone gave a date of 1,720 Ma (Percival, 1981; Fig. 13).

Three alkalic rock-carbonatite complexes in the southern Kapuskasing zone have K-Ar dates of 1,050-1,100 Ma (Gittins et al., 1967). Plagioclase from amphibolite in the footwall of the Ivanhoe Lake cataclastic zone yields a $^{40}Ar/^{39}Ar$ plateau at 1107 Ma (Fig. 13), suggesting mild resetting, possibly due to faulting. Lower concordia intercepts of zircon discordia in the range 827-1,108 Ma (Percival and Krogh, 1983) may relate to uplift 1,100-1,000 Ma ago.
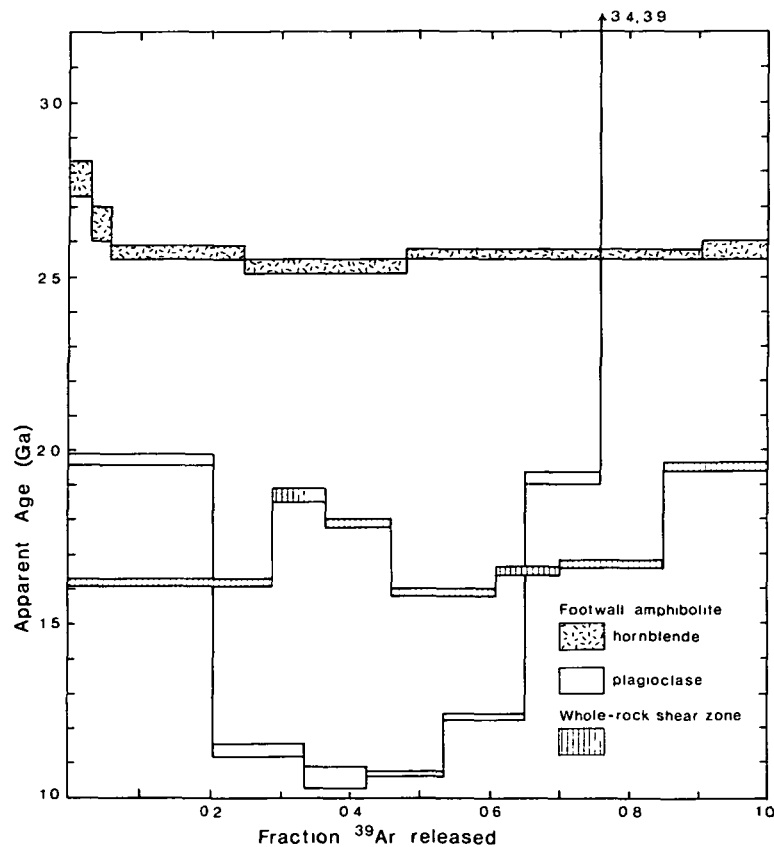
Fig. 13.    $^{40}Ar/^{39}Ar$ age spectra for hornblende, plagioclase and whole-rock separates from near the Ivanhoe Lake cataclastic zone. Increasingly high release temperature from left to right. Width of bars are $2\sigma$ uncertainty estimates.

The coincidence of Proterozoic events along the Kapuskasing structure with major orogenic activity elsewhere in the Shield suggests that the structure is an intracratonic basement uplift related to a distant cause.

## DAY 1 - GEOLOGY OF THE MICHIPICOTEN GREENSTONE BELT, WAWA, ONTARIO.

The Michipicoten greenstone belt trends approximately east-west and consists of mafic and felsic flows and pyroclastics, clastic sediments, and iron formation, notably carbonate(siderite)-rich. U-Pb zircon ages on felsic volcanic units in the southern part of the belt indicate the sequence is about 2696 to 2749 Ma old (Turek et al., 1982). The supracrustal rocks are intruded and surrounded by felsic plutonic and gneissic rocks, including layered mafic-felsic tonalitic and granodioritic gneiss and granitic plutons of quartz diorite to granite and syenite composition. Several of the older, synvolcanic plutons yield U-Pb zircon ages of 2744 and 2737 m.y. A post-volcanic granodiorite is 2685 Ma old, and recently Turek et al. (1983) obtained an age of 2888 $\pm$ 2 Ma on a granite southeast of the Michipicoten belt.

The rocks of the Michipicoten belt have been metamorphosed under low greenschist to amphibolite facies conditions, and have undergone extensive faulting and folding. There are major overturned (recumbent) fold structures with axial surfaces dipping 20° to 30° northeast (Attoh, 1980).

The supracrustal rocks of the southern part of the belt can be divided into three major cycles (Goodwin, 1962; Sage, 1980). A lower cycle, consisting of mafic and felsic volcanics, is capped by Michipicoten-type iron formation, mainly siderite, but with lesser pyrite-, chert- and graphite-rich rocks. The middle cycle comprises mafic volcanics overlain by clastic metasediments and felsic tuffs and breccias. The clastic sediments, including the Doré conglomerate, wacke, siltstone, and crossbedded arkose, are the facies equivalents of the felsic pyroclastics and are formed mainly of detritus eroded from the felsic centres. The upper cycle comprises intermediate to felsic (andesite - dacite) tuffs and quartz-feldspar porphyry. The associated Jubilee Stock, a high-level subvolcanic intrusion, was emplaced within a caldera structure (Sage, 1980).

## Michipicoten Belt Stop Descriptions

### STOP 1-1 - Doré Conglomerate

Exposures of Doré conglomerate will be examined on Highway 17, 9.5 km north of the junction of Highway 17 and 101 at Wawa.

The Doré is a thick, coarse polymictic conglomerate unit that is overlain and underlain by metavolcanic rocks. Eastward there is a facies transition from Doré conglomerate into a sequence of wacke, siltstone, cross-bedded arkose and conglomerate called the "Eleanor Slate". At this locality, coarse felsic tuff-breccias to the north are succeeded southward by a sequence of wacke (reworked tuff?) and Doré conglomerate. Bedding and foliation dip approximately $45°$ southeastward and may face downward (west).

The Doré consists of pebbles and boulders of mafic and felsic volcanics, quartz porphyry, iron formation and trondhjemite in a schistose, chloritic matrix. The pebbles are flattened in the foliation plane and elongated in the east-plunging rodding lineation. Variations in pebble packing and size define crude stratification units.

### STOP 1-2 - Helen Iron Range Section at McLeod Mine, Algoma Ore Properties Ltd.

The McLeod Mine extracts siderite and pyrite from the base of the Helen iron range. This iron range lies at the top of a 1800 m thick unit of intermediate to felsic metavolcanics consisting of oligomictic and polymictic breccia, thin bedded to massive tuffs, lapilli tuffs, spherulitic flows, flow banded flows, and crystal tuffs. These intermediate to felsic metavolcanics are intruded by gabbro to quartz diorite dykes and sills that reach 290 m in thickness.

The intermediate to felsic metavolcanics overlie a dominantly massive and pillowed sequence of intermediate to mafic metavolcanics.

The mafic to felsic metavolcanics are bimodal in composition and represent the oldest cycle of volcanism (Fig. 14).
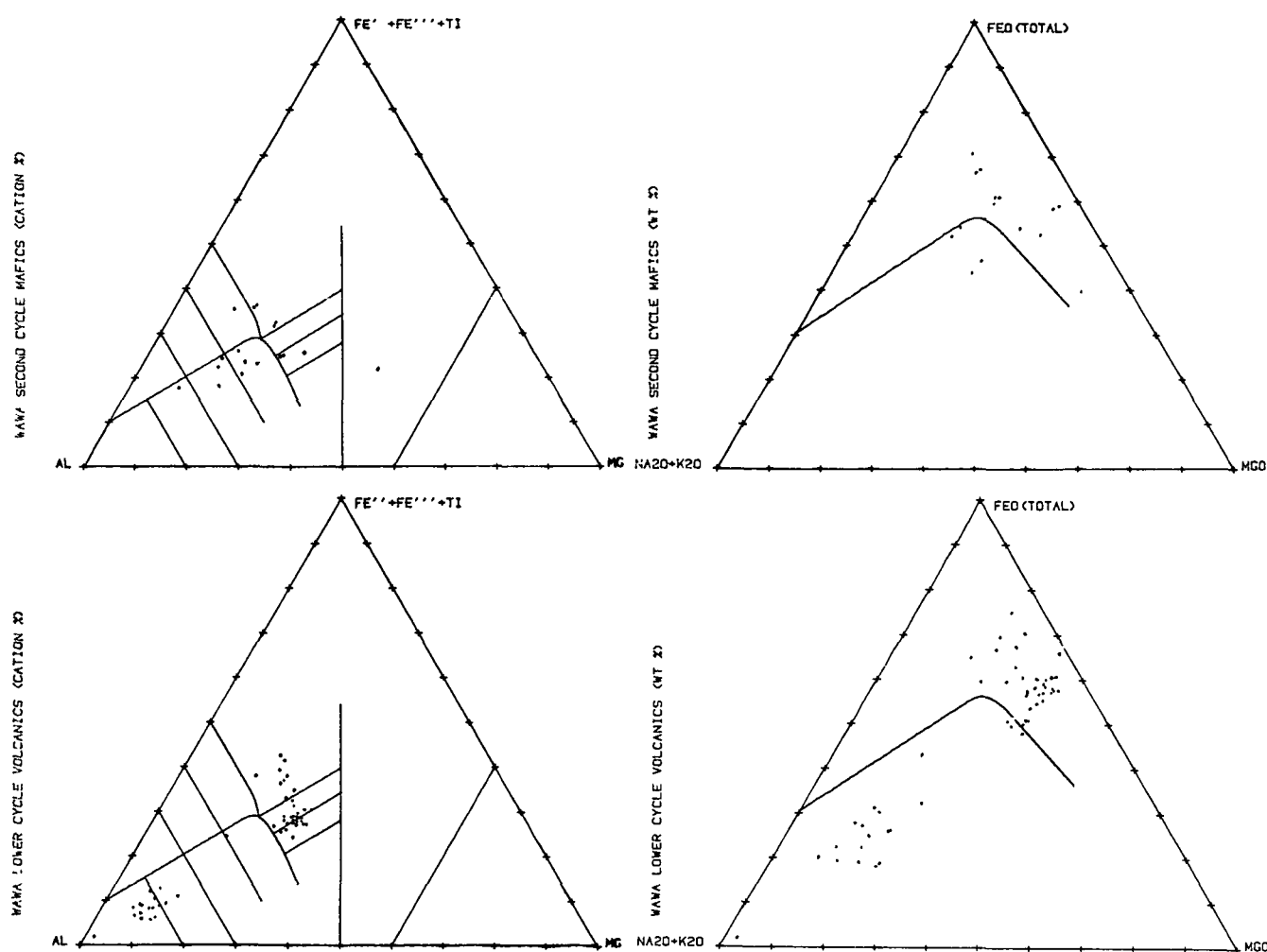
Fig. 14.    AFM and Jensen plots for metavolcanic rocks of the Michipicoten belt.

The iron formation caps the first cycle of volcanism in the Wawa greenstone belt and from the stratigraphic bottom to top consists of five lithologic rock types that are gradational into each other.  Upper and lower contracts are sharp.  From stratigraphic bottom to top the iron formation consists of siderite, pyrite, banded chert, thin bedded chert-wacke-magnetite, and graphite-pyrite.  A U-Pb isotopic age of approximately 2749 Ma has been obtained from the crystal tuffs lying immediately below the iron formation (Turek et al., 1982).

Overlying the iron formation are massive and pillowed intermediate to mafic metavolcanics of cycle two.

The area of the mine displays numerous major and minor faults, and contains several Proterozoic diabase dykes that strike northeast and northwest.

Beneath the iron formation an area of approximately 1800 m by 750 m has been outlined that contains visibly identifiable chloritoid. These crystals are generally up to 2 mm in diameter and are randomly oriented with respect to the schistosity. Chloritoid alteration occurs above the iron formation as well, but its presence is not visually as obvious.

The volcanics above and below the iron range are pervasively soaked with carbonate. Carbonate alteration is of regional extent and occurs in all rock types, volcanic, sedimentary, and intrusive.

STOP 1-3

This stop contains the best exposure of a spherulitic (hollow) flow banded felsic flow within the McLeod Mine area. The spherulitic unit is overlain by a flow breccia containing good fiamme and this unit is in turn overlain by a massive tuff with scattered lapilli size clasts. This exposure lies in the lower part of the felsic part of the oldest cycle of volcanism. Stratigraphic tops are north and the section is overturned, dipping south. Interpreted faulting along Wawa Lake and beach deposits at Wawa prevent any estimate as to how far above the contact with the intermediate to mafic pillowed and massive metavolcanics this felsic section occurs.

STOP 1-4a - Sir James Mine, Eleanor iron range

The Eleanor iron range is the stratigraphic equivalent of the Helen iron range and consequently the stratigraphic sequences of the two are essentially identical. The two ranges are separated by northwest-striking faults. In the Eleanor range, the intermediate to felsic metavolcanics of the stratigraphic footwall have been more

intensely sheared than the footwall rocks of the Helen iron range. As a result of this shearing, the footwall rocks of the Eleanor iron range are fissile and primary textures are not as well preserved. The exact thickness of the intermediate to felsic section is probably on the order of 510 m and has been intruded by dykes and sills of gabbro to quartz diorite.

The Eleanor iron formation is estimated to be approximately 75 m thick.

STOP 1-4b

The upper contact of the Eleanor iron formation consists of a unit of approximately 10 m thickness of graphite schist with pyrite nodules. This is the best exposure of the graphitic upper unit of the Michipicoten iron formation.

Lying in sharp contact with the stratigraphic top of the iron formation is the massive lower part of an overlying intermediate to mafic flow. The graphite unit is cut by an irregular quartz porphyry dyke.

Immediately west of the exposure of the upper contact of the iron formation excellent exposure of pillowed intermediate to mafic metavolcanics. These intermediate to mafic metavolcanics are the base of cycle two volcanics. Note that from pillow shape the sequence faces north.

STOP 1-5 (optional) - Lucy Iron Range

The Lucy iron range is the faulted eastern extension of the Eleanor iron range and is thus stratigraphically equivalent. Left lateral movement along the Midland Lake fault, which separates the two iron ranges, is approximately 3.1 km.

The stratigraphy of the Lucy range is identical to that of the Helen and Eleanor iron ranges and the intermediate to felsic part of the section is of a thickness similar to that of the Eleanor iron range.

The thickness of the underlying massive and pillowed intermediate to mafic metavolcanics of the first cycle is unknown due to faulting. At the Lucy range the mafic

and felsic parts are separated by a polymictic mafic breccia of approximately 50 m thickness. The overlying intermediate felsic metavolcanics consist of lapilli tuffs, bedded tuffs, and a rare fiamme-bearing flow. The intermediate to felsic section is approximately 250 m thick and has been intruded by dykes and sills of possibly original gabbro composition.

The felsic unit is overlain by iron formation of an estimated 75 m width. The stratigraphy of the iron formation is the same as that of the previous iron ranges. The upper section of the Lucy range contains good thin bedded chert-wacke. The wacke beds display grain gradation, and numerous interformational units of brecciated chert-wacke occur. The brecciated units alternate with unbrecciated units; the breccias are primary in origin.

Above the iron formation lie approximately 1000 m of monotonous intermediate to mafic massive and pillowed mafic volcanics of cycle two.

Cycle two mafic volcanics are overlain by metasediments that are dominantly wackes. The volcanic-sedimentary contact is conformable and defined by graphite-pyrite, and pebbly conglomerate with pyrite. Immediately overlying the graphite-pyrite and pebble conglomerate are thin bedded wackes that are presently chlorite schists. Above the chlorite-schist sediments lie wackes that approach arkose in composition. These cleaner wackes display excellent cross-bedding and rare ripple marks, and dessication cracks have been found. The thickness of the overlying metasediments is unknown due to faulting, but at least 450 m is present. The cleaner wackes may be the sedimentary detritus from the felsic centre of cycle two volcanism located 4 km northwest of the McLeod Mine. This centre has been dated by U-Pb methods at approximately 2696 Ma (Turek et al., 1982).

DAY 2 - <u>GEOLOGY OF THE WAWA DOMAL GNEISS TERRANE IN THE WAWA-CHAPLEAU AREA</u>

In this section we will examine the relationships between supracrustal rocks of the Michipicoten belt and plutonic rocks of the Wawa terrane, and the transition from Wawa domal gneiss to high-grade rocks of the Kapuskasing zone.

<u>km</u>

00.0　　Junction of Highway 101E and Mission Road, Wawa. Proceed east on Highway 101.

63.1　　Junction Highway 101 and Highway 651. Proceed north on Highway 651.

80.2　　STOP 2-1 - <u>Mafic gneiss</u>

A large enclave of mafic gneiss is enclosed in and intruded by tonalitic gneiss in a migmatitic zone marginal to the Michipicoten greenstone belt. The mafic gneiss is considered to represent deformed, metamorphosed (amphibolite facies) Michipicoten volcanics. The clinopyroxene-hornblende-plagioclase mafic gneiss is compositionally layered (1 to 40 cm) and contains thin (1 to 5 mm) concordant tonalite layers. It is cut by early tonalitic intrusions, late aplitic and pegmatitic dykes, and still later mafic and lamprophyric dykes. The gneiss displays subvertical foliation, mineral lineation, and tight steeply-plunging isoclinal minor folds.

Return to Highway 101 and proceed east.

127.1　　STOP 2 - <u>Tonalite gneiss and Mafic dykes</u>

Tonalite gneiss is cut by northwest- and northeast-trending mafic dykes with good chilled margins. The older northwest-striking Hearst dykes (Ernst and Halls, 1980) occur west of the Kapuskasing zone. The dykes have a similar trend and similar characteristics, including plagioclase phenocrysts and tholeiitic composition, to Matachewan dykes east of the Kapuskasing zone. The Matachewan dykes have an Rb-Sr whole-rock age of $2633 \pm 75$ Ma (Gates and Hurley, 1973). Ernst and Halls (1980) also reported similar paleomagnetic poles for the two swarms. In a zone 50 km wide west of the Kapuskasing zone, the Hearst dykes average 4 m in width and have a consistent easterly dip of 80° (Ernst, 1982; Percival, 1981). The tonalitic gneiss is thinly layered

and has sparse mafic xenoliths. The structure, although appearing chaotic, is generally subhorizontal. There is evidence for at least two foliations, and older gneissosity that is reoriented by a younger foliation to give complex sigmoidal patterns.

km

153.6   STOP 2-3 - Highbrush Lake Dome

Small domes exposed in roadcuts here are small-scale examples of the regional-scale structural pattern. The main rock type is fine grained, thinly layered biotite and hornblende-biotite gneiss with local augen of potash feldspar.

The outcrop consists of several domes or canoe-shaped folds. Small intrafolial folds are present in some parts of the outcrop, as are amphibolitic xenoliths, locally with clinopyroxene. Pink granitic pegmatite dykes and sills cut gneissic layering in some areas.

161.7   Junction of Highway 101 and Highway 129.

Proceed north on Highways 101-129 through Chapleau. Follow the Missinaibi Provincial Park signs north of town on gravel road. Proceed north and east on park road and logging roads leading to the Chapleau River.

STOPS 2-4 and 2-5 - Floranna Lake Complex

The Floranna Lake Complex is a complex crescentic pluton of intermediate composition. The western margin consists of lineated, fine grained granite with orthopyroxene and biotite. Inside the marginal unit is lineated diorite to monzonite containing hornblende-rimmed augen clinopyroxene phenocrysts. The central part of the complex is foliated and lineated hornblende-clinopyroxene-biotite diorite, gabbro and melagabbro. The well-exposed eastern contact of the complex shows extremely attenuated and contorted layering in granite of the complex adjacent to rocks of the Robson Lake dome to the east.

208.4   STOP 2-4 - Clinopyroxene augen diorite-monzonite

These rocks are strongly lineated, medium- to coarse-grained monzonite. Rod-shaped clinopyroxene phenocrysts, rimmed by hornblende, make up some 20% of the

rock. Original outlines of feldspar grains are visible but these are now polycrystalline aggregates.

<u>km</u>

209.4   <u>STOP 2-5</u>

Foliated and lineated diorite with gabbro and melagabbro layers. Igneous clinopyroxene with hornblende overgrowths and granitic leucosome layers are present.

213.3   <u>STOP 2-6 - Granulite gneiss, Robson Lake dome</u>

The easternmost exposure of the igneous complex is strongly foliated, hornblende porphyritic granite augen gneiss with leucogranite veinlets parallel to foliation on the 1 mm-2 cm scale. The layering is folded about a gently northwesterly-plunging axis and the eastern limb is fine-grained and mylonitic with highly attenuated layering.

To the west, the Robson Lake dome consists of interlayered mafic and tonalitic gneiss. Near the contact with the Floranna Lake Complex, the layering in rocks in the dome is concordant to the contact and dips steeply west, however, the attitude is horizontal farther east in the core of the dome. At this outcrop the mafic rocks consist of garnet-clinopyroxene-hornblende-plagioclase-quartz assemblages, with layering defined by different proportions of minerals, including garnet-rich and hornblende-rich varieties. Concordant tonalitic layers locally have large clinopyroxene crystals rimmed by hornblende, in clots up to 4 cm. The layering is folded about upright isoclinal folds locally.

Small dykes of granite in mafic gneiss and inclusions of mafic gneiss in granite indicate an intrusive contact between the Floranna Lake Complex and Robson Lake dome.

The interpretation of the Robson Lake dome, which has the structural attributes of the Wawa domal gneiss terrane and the lithological characteristics of the Kapuskasing zone, is that rocks like those exposed in the Kapuskasing zone underlie the Wawa domal terrane and have been exposed here in a structural culmination.

## DAY 3 - KAPUSKASING STRUCTURAL ZONE

In this section we will examine the rocks and structures of the Wawa domal gneiss - Kapuskasing zone boundary and of the Kapuskasing zone itself along Highway 101.

km

0.0    Junction of Highway 101E and Highway 129 south of Chapleau. Proceed east on Highway 101E.

13.0   STOP 3-1 - Borden Lake conglomerate

This outcrop consists of stretched-pebble metaconglomerate with a strong rodding lineation and weak, gently north-dipping foliation. The rock is a clast-supported conglomerate containing ~10% matrix of garnet-hornblende-biotite-quartz. The cobbles, which range up to 1 m in length, are felsic metavolcanics, metasediments, granodiorite-tonalite, plagioclase-porphyritic meta-andesite and amphibolite, with rare hornblendite and vein quartz. The metaconglomerate is spatially associated with amphibolite and paragneiss to the south on Borden Lake, and is cut by granite, however, the stratigraphic relations of the supracrustal rocks are unknown.

Tonalitic cobbles extracted from the metaconglomerate have yielded zircons dated at 2664 ± 12 Ma (Percival et al., 1981). The zircons have a corroded appearance and produced discordant data points and hence the interpretation of the data is open. The zircons could preserve the original crystallization age of the source pluton for the cobbles or they could record a later deformation-metamorphic event.

25.9   STOP 3-2 - Mafic gneiss xenoliths with amphibolitic retrograded margins

This exposure demonstrates an important aspect of the boundary between the Kapuskasing zone and Wawa gneiss terrane. Aside from the large-scale structural contrast between the domal Wawa terrane and the linear belts in the Kapuskasing zone, intrusive relations are also instructive. The outcrop consists of two main components: (1) coarse-grained hornblende-biotite tonalite, dominant to the west, and (2) medium

grained mafic gneiss consisting of garnet-clinopyroxene-hornblende-plagioclase-quartz assemblages. Small xenoliths of mafic gneiss in tonalite have margins, up to several cm thick, consisting of hornblende-plagioclase. Dykes of tonalite cutting mafic gneiss are bordered by mafic rock with hornblende-plagioclase assemblages. The interpretation of age relationships is that the high-grade metamorphism that produced the garnet-clinopyroxene assemblages in mafic gneiss preceded the intrusion of tonalite. Water in the tonalite margin was presumably released upon crystallization and hydrated the adjacent less-hydrous mafic rock. Although the tonalite at this outcrop has not been dated, it probably belongs either to the >2707 Ma tonalite gneiss suite or to the 2680 Ma group of plutons. The high-grade metamorphism is therefore older than 2680 Ma. This is in conflict with the direct dates of metamorphic zircons from the Kapuskasing zone that yield ages of $2650 \pm 2$ and $2627 \pm 3$ Ma. The conflict has led to the suggestion (Percival and Krogh, 1983) that the Kapuskasing gneisses, although metamorphosed prior to 2680 Ma ago, remained at high temperatures where radiogenic lead diffused readily out of zircon until <2627 Ma ago.

35.6    Paul Township road - turn left and proceed north.

STOPS 3-3 and 3-4 - Shawmere anorthosite complex

The Shawmere anorthosite complex is a layered calcic anorthosite body. In this traverse we will examine the central portion comprising mainly megacrystic gabbroic anorthosite.

46.8    STOP 3-3

This outcrop consists of coarse grained gabbroic anorthosite with patches of anorthosite and gabbro on the 1 m scale. Original outlines of plagioclase ($An_{50}$) phenocrysts, now polycrystalline aggregates, are up to 10 cm. Local mafic patches contain the assemblage plagioclase ($An_{92}$)-garnet-orthopyroxene-hornblende-gedrite-spinel-sapphirine (Table 2). Garnet is present both as discreet grains and in coronal structures between hornblende and plagioclase.

| | OG-1<br>1 | OG-1<br>2 | OG-1<br>3 | OG-1<br>4 | OG-1<br>5 | OG-1<br>6 |
|---|---|---|---|---|---|---|
| $SiO_2$ | 41.18 | 53.67 | 12.80 | 0.26 | 44.04 | 44.29 |
| $TiO_2$ | 0.03 | 0.00 | 0.08 | 0.03 | 0.10 | 0.05 |
| $Al_2O_3$ | 23.01 | 4.66 | 62.27 | 64.11 | 19.04 | 16.02 |
| $Cr_2O_3$ | 0.09 | 0.05 | 0.06 | 0.06 | 0.08 | 0.08 |
| FeO | 19.70 | 13.48 | 6.33 | 19.09 | 11.01 | 7.80 |
| MnO | 0.55 | 0.12 | 0.06 | 0.06 | 0.18 | 0.08 |
| MgO | 14.15 | 27.74 | 16.75 | 15.43 | 21.11 | 15.50 |
| CaO | 3.38 | 0.07 | 0.04 | 0.00 | 0.62 | 11.00 |
| $Na_2O$ | 0.44 | 0.65 | 0.00 | 0.07 | 2.03 | 1.74 |
| $K_2O$ | 0.01 | 0.08 | 0.00 | 0.01 | 0.00 | 0.21 |
| Total | 102.60 | 100.56 | 98.41 | 99.21 | 98.22 | 96.82 |
| Si | 2.981 | 1.894 | 0.769 | 0.007 | 6.115 | 6.273 |
| Ti | 0.002 | 0.00 | 0.004 | 0.001 | 0.010 | 0.005 |
| Al(iv) | 0.00 | 0.108 | 0.231 | 0.00 | 1.885 | 1.728 |
| Al(vi) | 1.962 | 0.088 | 4.176 | 1.974 | 1.230 | 0.946 |
| $Cr_3$ | 0.005 | 0.001 | 0.003 | 0.002 | 0.008 | 0.011 |
| $Fe_2$ | 0.00 | 0.065 | 0.034 | 0.002 | 0.065 | 0.474 |
| Fe | 1.192 | 0.333 | 0.284 | 0.415 | 1.214 | 0.450 |
| Mn | 0.034 | 0.004 | 0.003 | 0.001 | 0.021 | 0.010 |
| Mg | 1.526 | 1.459 | 1.499 | 0.601 | 4.368 | 3.271 |
| Ca | 0.262 | 0.003 | 0.003 | 0.00 | 0.092 | 1.699 |
| Na | 0.062 | 0.044 | 0.00 | 0.004 | 0.546 | 0.478 |
| K | 0.001 | 0.004 | 0.00 | 0.00 | 0.00 | 0.038 |

Table 2    Microprobe analyses of minerals in magnesian melagabbro from Shawmere
anorthosite complex.  1: garnet; 2: orthopyroxene; 3: sapphirine;
4: spinel; 5: orthoamphibole; 6: clinoamphibole.  Also present is
anorthite ($An_{94}$).

<u>km</u>

54.7   <u>STOP 3-4</u>

At this exposure, large crystals of orthopyroxene, partly rimmed by hornblende and garnet, are preserved in anorthositic gabbro. The more mafic layers alternate with plagioclase-rich layers on a 20 cm scale. The effect of deformation on anorthosite can be locally observed where discrete zones of fine- to medium-grained gabbroic anorthosite grade from well-preserved igneous textures and minerals.

Return to Highway 101 and proceed east.

75.2   <u>STOP 3-5 - Thinly-layered tonalitic gneiss and diatreme breccia</u>

Fine grained tonalitic gneiss at this exposure is strongly foliated and layered on a 1-5 mm scale with garnet, hornblende and biotite-rich layers. Extremely attenuated intrafolial folds are present locally. Units characterized by extremely planar foliation such as this are relatively rare in the Kapuskasing zone. Although the orientation of foliation in this exposure is typical for the Kapuskasing zone, most Kapuskasing gneisses are medium- to coarse-grained and layered with distinctive leucocratic portions. In addition, the layering in the typical gneisses is warped about gently northeast or southwest-plunging axes. The fine grain size and thin planar layering in this outcrop suggest a relatively late, high-strain flattening or shearing event.

A thin diatreme dyke occurs in this same exposure. It has not been dated but presumably is part of a set of lamprophyre dykes of $\sim$1100 Ma age that occur in the Kapuskasing zone and are particularly common in the area between the Lackner and Nemegosenda Lake complexes. Both the matrix and fragments in the dyke are altered, but some fragments can be identified as massive pink granite. As massive granite does not occur in the Kapuskasing zone, the granite fragments are relatively exotic. Their source was probably below the Kapuskasing zone, possibly in granite of the Abitibi belt, which according to the gravity modelling, lies vertically below at a depth of $\sim$15 km.

91.0    STOP 3-6 - <u>Kapuskasing Gneiss</u>

Layered mafic gneiss with tonalitic intrusions and sweats. Layering on the 5 to 10 cm scale is given by alternating hornblende-rich and pyroxene-rich layers. There are numerous minor folds with sheared limbs.

98.0    STOP 3-7 - <u>Kapuskasing gneisses</u>

There are several features of interest at this outcrop (Fig. 15):

A.    Mafic gneiss is present on the northwest side of the road. It is a coarse grained rock consisting of three types of layers on the 5-100 mm scale: i) relatively anhydrous mafic rock made up of garnet, clinopyroxene, plagioclase and quartz, with some hornblende (analogous to analyses 1 & 3, Table 1); ii) more hydrous layers containing less garnet and clinopyroxene and more hornblende (analogous to analyses 2 & 4, Table 1); and iii) tonalitic leucosome layers, both concordant to layering and transverse in the amphibole-rich mafic rocks. Note that the tonalite has no retrogressive effect on adjacent anhydrous mafic gneiss. The tonalitic leucosome veinlets are considered to be in situ anatectic melt segregations developed during prograde metamorphic reactions (see reaction 2). In the western end of the outcrop, submicroscopic symplectites of orthopyroxene-plagioclase form barely-visible coronas around garnet, clnopyroxene and hornblende. Analyses of the symplectite minerals, at the lower size limit of microprobe resolution, are reported along with those from the other minerals in the rock, in Table 3. The rock contains three plagioclase compositions. $An_{89}$ is present in coronas whereas worm-like intergrowths of $An_{35}$ and $An_{50}$ make up the matrix plagioclase.

The mineral compositions yield estimates of 735°C using the Ellis and Green (1979) garnet-clinopyroxene thermometer and 6.2 kbar using the garnet-clinopyroxene-plagioclase-quartz barometer (Newton and Perkins, 1982). At the same temperature the coronal minerals and matrix garnet yield 9.1 kbar with the garnet-orthopyroxene-plagioclase-quartz Newton and Perkins barometer.
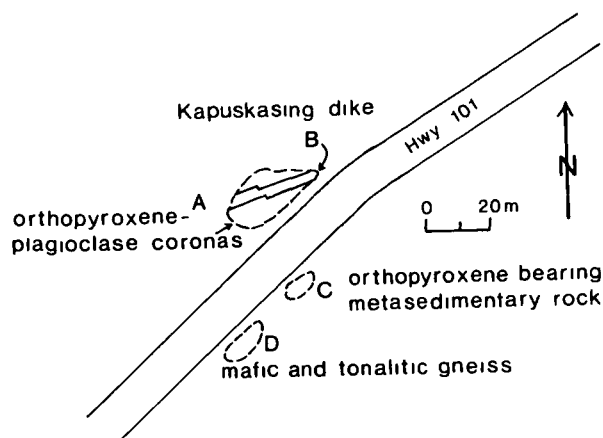
Fig. 15.     Location of outcrops at Stop 3-7.

B.     A Kapuskasing mafic dyke cuts the eastern end of the outcrop. The overall attitude of the dyke is 070/75 SE although the margin is offset by numerous small sinistral(?) faults. The outer 2 cm of the margin is chilled. Sparse plagioclase phenocrysts are present in the dominantly medium grained ophitic olivine-bearing gabbro. Several dykes of this swarm have been dated by the whole-rock K-Ar method and yield "ages" between 2367 and 3649 Ma, indicating the presence of excess argon (Stevens et al., 1981).

C.     Homogeneous metasedimentary rock

South of the road is a flat outcrop of medium grained rock with the assemblage garnet-orthopyroxene-biotite-plagioclase-quartz. Plagioclase occurs as porphyroblasts to 2 cm and orthopyroxene is up to 5 mm. The rock has the same mineral assemblage as high-grade paragneiss in the Kapuskasing zone but lacks the migmatitic layering typical of paragneiss.

|  | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| $SiO_2$ | 38.01 | 51.57 | 49.06 | 42.49 |
| $TiO_2$ | 0.00 | 0.34 | 0.03 | 2.03 |
| $Al_2O_3$ | 20.99 | 2.92 | 1.75 | 12.98 |
| $Cr_2O_3$ | 0.22 | 0.21 | 0.34 | 0.08 |
| FeO* | 28.06 | 11.81 | 31.20 | 18.43 |
| MnO | 0.70 | 0.00 | 0.81 | 0.17 |
| MgO | 4.11 | 11.34 | 13.35 | 9.28 |
| CaO | 8.32 | 22.65 | 1.39 | 11.41 |
| $Na_2O$ | 0.27 | 0.74 | 0.52 | 1.95 |
| $K_2O$ | 0.00 | 0.08 | 0.00 | 0.69 |
| Total | 100.67 | 101.79 | 101.45 | 99.31 |
| Si | 2.973 | 1.908 | 1.885 | 6.252 |
| Al(iv) | 0.00 | 0.092 | 0.115 | 1.748 |
| Al(vi) | 1.935 | 0.035 | 0.100 | 0.513 |
| Ti | 0.00 | 0.009 | 0.001 | 0.226 |
| Cr | 0.014 | 0.008 | 0.010 | 0.009 |
| $Fe^{3+}$ | 0.0 | 0.087 | 0.042 | 0.288 |
| $Fe^{2+}$ | 1.835 | 0.278 | 0.960 | 1.990 |
| Mn | 0.040 | 0.003 | 0.026 | 0.021 |
| Mg | 0.479 | 0.625 | 0.764 | 2.044 |
| Ca | 0.697 | 0.898 | 0.057 | 1.807 |
| Na | 0.041 | 0.053 | 0.039 | 0.559 |
| K | 0.00 | 0.004 | 0.00 | 0.130 |
| (O) | (12) | (6) | (6) | (23) |

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1: garnet; 2: clinopyroxene, 3: orthopyroxene; 4: hornblende

* Total iron as FeO; $Fe^{3+}$ by stoichiometry

Specimen also contains quartz and plagioclase ($An_{33,57}$) in matrix, $An_{88}$ in symplectite)

Table 3. Microprobe analyses of minerals in coronitic mafic gneiss, stop 3-5.

D.      Interlayered mafic and tonalitic gneiss

This outcrop demonstrates complex relations between mafic and tonalitic gneiss. Isoclinal folds of layering are truncated by tonalite pods and dykes, suggesting multiple generations of tonalite.

<u>km</u>

107.2    STOP 3-8 - <u>Ivanhoe Lake cataclastic zone</u>

The outcrop south of the highway is on the western, high-grade side of the cataclastic zone and consists of migmatitic mafic gneiss with garnet-clinopyroxene-hornblende-plagioclase-quartz assemblages. It is transected by numerous small fault offsets and by one major cataclasite vein. In thin section, this black aphanitic material is seen to consist mainly of (recrystallized) fine actinolitic amphibole and of porphyroclasts of hornblende. A $^{40}Ar/^{39}Ar$ whole-rock analysis of material from this vein yielded an age plateau at 1720 Ma (Fig. 13). On the west side of the outcrop are thin (3 cm) rusty-weathering lamprophyre dykes.

108.7    STOP 3-9 - <u>Mafic metavolcanics, Abitibi subprovince</u>

This outcrop is on the eastern low-grade side of the ILCZ and is the westernmost exposure of metavolcanics of the Abitibi belt. It is a fine grained, layered hornblende-plagioclase ± clinopyroxene rock with local rusty-weathering patches.

The structural characteristics of the outcrop, including east-west strike of layering, vertical dip and steeply-plunging isoclinal small folds, are typical of the Abitibi belt. Chlorite and epidote are common to the east along strike, where the belt is wider, suggesting an easterly decrease in metamorphic grade.

Analyses of hornblende and plagioclase from this outcrop by the $^{40}Ar/^{39}Ar$ method (Fig. 13) show a plateau for hornblende at 2567 Ma and a saddle-shaped spectrum for plagioclase with a plateau at 1107 Ma. The plagioclase plateau may be due to argon loss resulting from a mild thermal event, possibly related to faulting.

Proceed east to Timmins.

## DAY 4 - GEOLOGY OF THE ABITIBI GREENSTONE BELT, TIMMINS AREA

The main emphasis will be on the rock types and stratigraphy of the volcanic and sedimentary rocks of the Abitibi Supergroup. Stops will be confined to exposures of volcanic rocks of the Tisdale Group and sedimentary rocks of the Porcupine Group.

### General Geology

With the exception of a few diabase dykes and Huronian sedimentary rocks of Proterozoic age, the bedrock of the area is of Archean age ( 2650 to 2750 Ma). The supracrustal rocks have been divided by Pyke (1982) into three groups, an older(?) metavolcanic Deloro Group, a younger metavolcanic Tisdale Group, and a metasedimentary Porcupine Group, considered to be stratigraphically equivalent to the upper part of the Deloro Group and the entire Tisdale Group (Fig. 16).

The Deloro Group is largely a calc-alkaline sequence, approximately 4500-5000 metres thick, and is composed mainly of andesite and basalt flows in the lower part and dacitic and rhyolitic flows and pyroclastics in the upper part. Iron formation is common at or near the top of the group. A major change in volcanic rock composition marks the lower contact of the Tisdale Group. The basal formation consists mainly of ultramafic volcanic rocks and basaltic komatiites. This is overlain by a thick sequence of tholeiitic basalts. The uppermost formation is largely volcaniclastic rocks of calc-alkaline dacitic composition. The total thickness of the Tisdale Group is almost 7000 metres.

The Porcupine Group consists of a lower turbiditic sequence of greywacke, siltstone and conglomerate, and an upper sequence of crossbedded fluviatile sandstone and conglomerate. The total exposed thickness of the group is approximately 3000 metres.
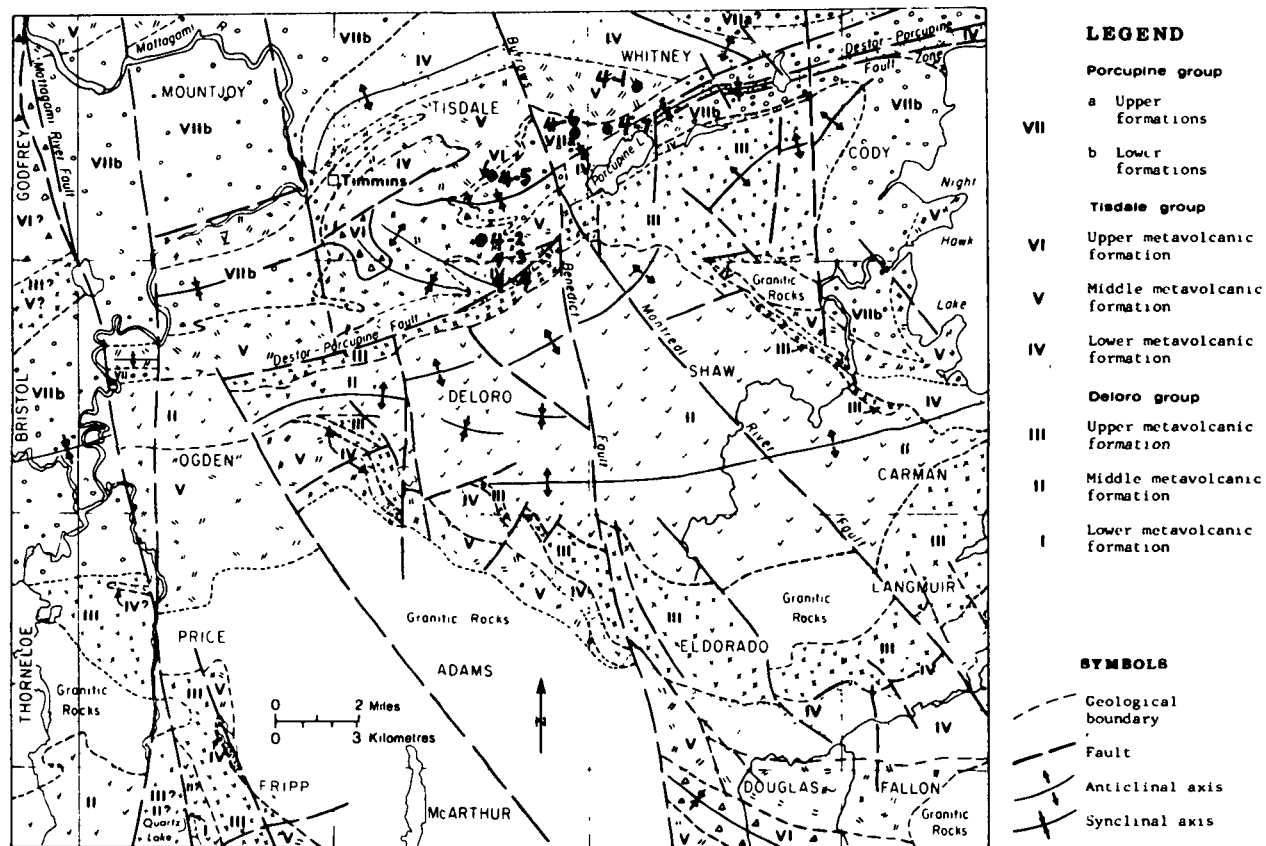
Fig. 16. Geology of the Timmins area (after Pyke et al., 1978).

Large sill-like bodies of dunite and lherzolite occur within the Deloro Group. Some of these bodies may be magma chambers for the overlying ultramafic volcanics of the Tisdale Group.

Small quartz-feldspar porphyry bodies are probably subvolcanic intrusions and may be part of intrusive-extrusive rhyolitic domes. Plutons and batholiths of trondhjemite, granodiorite, granite and syenite intrude the supracrustal rocks. Several of the massive, unmetamorphosed granodioritic intrusions yield U-Pb zircon ages close to 2680 Ma, thus approximately dating the termination of major orogenesis in this part of the Superior Province.

### STOP 4-1 - Mental Hospital Stop

This stop is located just north of the Resource Centre near the northeast end of Porcupine Lake. Travelling eastward from Timmins, turn north off Highway 101 on the road just before the Resource Centre. Then turn west (0.9 km) on the first road and proceed 100 m to the start of the trail.

This stop illustrates the upper part of formation IV of the Tisdale Group and the overlying basal portion of formation V. The contact between formations IV and V is placed at the first readily recognizeable iron-rich tholeiitic basalt. This Fe-tholeiite is possibly correlative with the 99 flow which underlies the V8 flow in Tisdale Township.

At this stop the upper part of formation IV consists of light grey Mg-tholeiitic basalt. That part of the lowermost flow included in the stop consists of pillow basalt and a thick sequence of pillow beccia. The overlying flow, also an Mg-tholeiite is massive at the base and pillowed in the upper part. Overlying this is a massive, medium grained, medium to dark green, Fe-tholeiitic basalt, which marks the base of formation V. This is overlain by variolitic pillow basalt, herein correlated with the V8 basalt in the Timmins gold camp. The overlying basalt is pillowed, locally contains very fine varioles and displays concentric cooling fractures. The overlying and uppermost flow on this stop is a
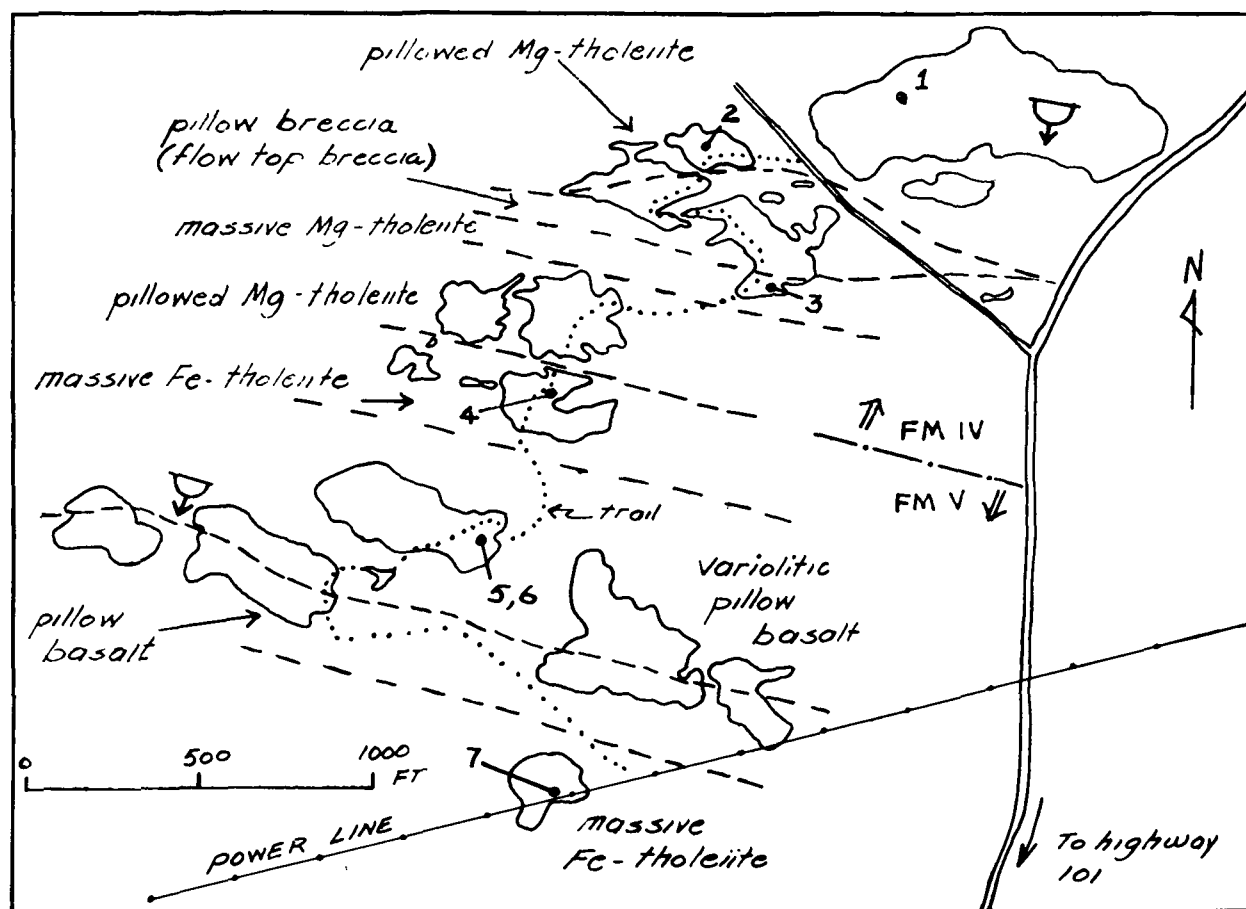
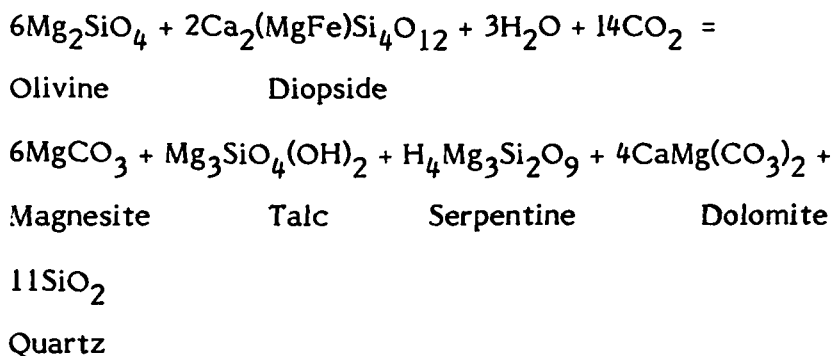Fig. 17.    Geology of Stop 4-1 (from Pyke et. al., 1978).

fine grained dark green Fe-tholeiitic basalt.  Although Mg-tholeiites are interlayered

with Fe-tholeiites in formation V in Whitney Township, they are absent in most of

Tisdale Township.

## STOP 4-2 - Carbonated Ultramafic

This stop is on the back road between Timmins and South Porcupine just south of

the Dome No. 3 head frame.

The outcrop is an altered peridotitic komatiite completely altered to carbonate minerals. Ultramafic flows such as this are almost wholly confined to the base of the Tisdale Group. The carbonatization has not destroyed the polysuturing which serves as an aid in the field to readily identify the rocks as being an ultramafic flow. This structure refers to the fracture pattern which is exhibited by the semi-equant polygons resembling mud cracks. Polysuturing is pervasive throughout ultramafic flows and probably represents some type of cooling phenomena, perhaps related to incipient pillow development.

A simplified equation for the alteration reaction of a komatiite might be:

$$6Mg_2SiO_4 + 2Ca_2(MgFe)Si_4O_{12} + 3H_2O + 14CO_2 =$$
Olivine     Diopside

$$6MgCO_3 + Mg_3SiO_4(OH)_2 + H_4Mg_3Si_2O_9 + 4CaMg(CO_3)_2 +$$
Magnesite   Talc   Serpentine    Dolomite

$$11SiO_2$$
Quartz

The irregular quartz stringers throughout the rock may be due to the released free silica being deposited in fractures.

## STOP 4-3 - Paymaster Porphyry

This stop is a few hundred metres east of the last stop.

The Paymaster porphyry is typical of quartz-feldspar porphyries of the area. It exhibits a light grey to buff weathering and is light grey to light green-grey on fresh surfaces. Quartz and albite crystals (phenocrysts or metacrysts?) are within a strongly foliated groundmass of fine albite, quartz and sericite.

The porphyries all occur within the lower formation of the Tisdale Group. Restriction to this stratigraphic interval suggest that they could represent rhyolitic

domes formed on the surface. This extrusive origin would explain the lack of offsets across contacts and the absence of evidence for forceful intrusion. If the porphyries within the Tisdale Group are not in part extrusive, there can be little doubt that they are high level intrusions of subvolcanic derivation.

### STOP 4-4 - Paymaster Mine

This stop is at the Porcupine Paymaster Mine, opposite Simpson Lake on the back road connecting Timmins and South Porcupine.

The stop illustrates some of the iron-rich tholeiitic basalts typical of formation V of the Tisdale Group. The most diagnostic feature of the Fe-tholeiites is the dark green colour as compared to the Mg-tholeiites previously visited at Stop 1. The variolitic basalts exposed here are interpreted to correlate with the variolitic basalts of Stop 1, Whitney Township. They form an important stratigraphic marker throughout the gold camp. At the Paymaster, a relatively dark green, vesicular, pillowed Fe-tholeiitic basalt (1) is exposed on the south side of the highway. Minor varioles are present and are a common feature to many of the flows in this part of the section. This pillowed flow also outcrops immediately north of the highway, as does a poorly exposed variolitic basalt (2). This in turn is overlain by an Fe-tholeiitic basalt (3) which is dominantly massive and contains minor blue quartz. This is in sharp contact with a variolitic pillow basalt. The pillows are commonly large (2-3.6 m in length) and the varioles are up to 4-5 cm in diameter, and may coalesce to form patches up to 45 cm in maximum dimension. Minor massive sections can be traced out within the dominantly pillowed variolitic basalt, and may represent intercalated flows. The coarse variolitic flow is overlain by a massive basalt (5) and this in turn by a variolitic pillow basalt (6). These flows are succeeded to the north by at least two and possibly four massive Fe-tholeiitic basalts. The northernmost basalt is highly carbonatized.
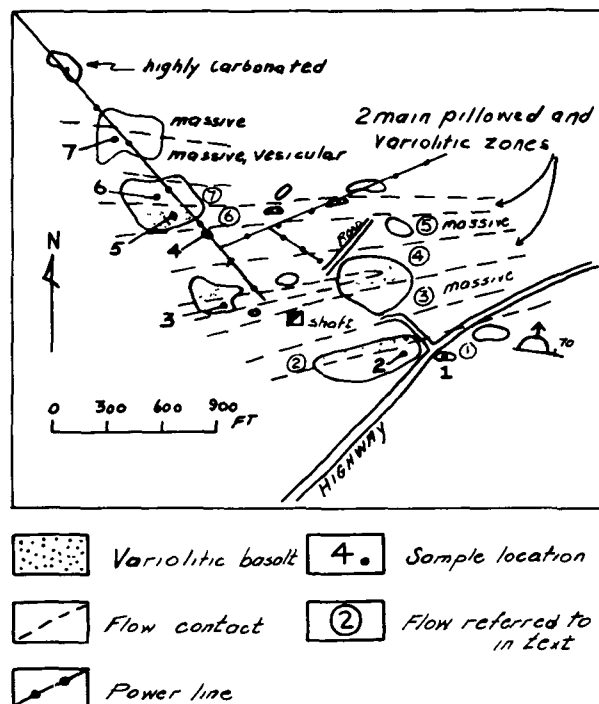
Fig. 18.   Geology of Stop 4-4 (from Pyke et al., 1978).


## STOP 4-5 - Krist Fragmental

This stop is approximately midway between Schumacher and South Porcupine, and is located at the large outcrop area adjacent to the south side of Highway 101 where the highway intersects a north-trending power line.

The Krist fragmental is a pyroclastic tuff breccia which forms the upper part (formation VI) of the Tisdale Group. The breccia is white-weathering, massive, and contains fragments of crystal tuff in a matrix of similar composition. The crystal fragments are dominantly white-weathering, subhedral, albitic feldspar averaging 1.5-2.5 mm in size; minor (5-10 per cent) quartz of comparable size is also present.

Although not readily obvious, the fragments are lineated and plunge eastward at approximately 45 degrees. Crystal tuff outcrops immediately adjacent to the highway, and rapidly passes into tuff-breccia farther down the power line. Most breccia fragments range from 2.5-15 cm in maximum dimension; the largest are about 45 cm. The Krist fragmental perhaps represents a glowing avalanche type of volcanic deposit.

## STOP 4-6 - Unconformity Outcrop

This stop is located 1.0 km north from Highway 101 along Crawford Street. Turn north at the Esso Service Station at the northeast end of South Porcupine. A trail west is just past the line between Concessions III and IV, Tisdale Township.

This stop illustrates the unconformity between what has classically been termed the "Keewatin" and "Timiskaming" sedimentary rocks in the Timmins area (Ferguson et al., 1968). The older "Keewatin" sediments, turbiditic wacke and siltstone, strike about 105°, dip steeply north, and face south conformably with the underlying volcanic rocks. Good examples of scouring, rip-up clasts, load casts and a Bouma sequence (A to D) may be seen in the northernmost exposures. A poorly sorted polymictic conglomerate approximately 60 m thick, trends 065 degrees and marks the base of the unconformity. Boulders are predominantly of basalt, siltstone, greywacke, and lesser gabbro and felsic volcanics. There does not appear to be any tectonic unconformity between the "Keewatin" and "Timiskaming sediments" as the structures within the two sedimentary sequences are the same.

## STOP 4-7 - Shovel Outcrop

This stop is at the first outcrop on Government Road, east of the turnoff to Stop 6.
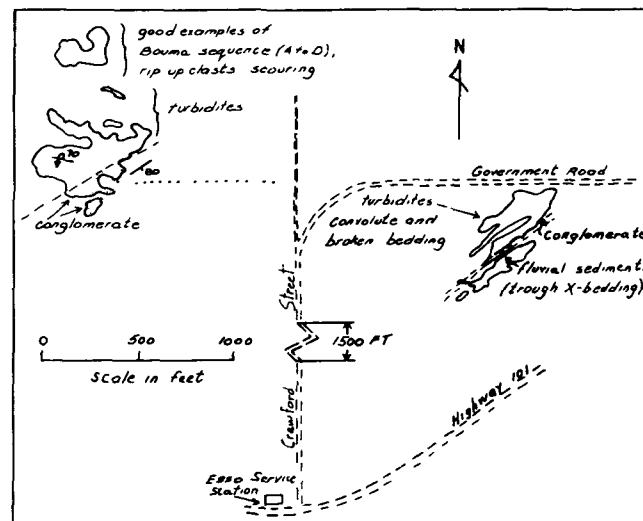
Fig. 19. Geology of Stops 4-6 and 4-7 (after Pyke et al., 1978).

The sediments at this stop perhaps contain the most significant sedimentary contact in the Timmins area. Here, there is a major change from turbidite to fluvial type of sedimentation. The contact between the two types of sedimentary rocks is not exposed, but the bedding is conformable. At the top of the turbidite sequence (base of the fluvial sediments?) is a narrow (1.8-2.4 m) polymictic conglomerate, conformable with the turbidites. Convolute bedding and disrupted and broken beds are common in the turbidites. The fluvial sediments are quartz-rich sandstones (lithic arenites) and display abundant trough crossbedding. This contact may warrant close scrutiny for gold mineralization.

Analysis (Wt%)

|  | 7 | 8 | 9 | 10 |
|---|---|---|---|---|
| $SiO_2$ | 54.7 | 43.9 | 72.7 | 55.6 |
| $Al_2O_3$ | 13.2 | 7.18 | 13.9 | 12.5 |
| $Fe_2O_3$ | 10.3 | 9.28 | 1.20 | 8.87 |
| FeO | 0.0 | 0.0 | 0.0 | 0.0 |
| MgO | 9.56 | 13.7 | 0.40 | 10.7 |
| CaO | 4.35 | 14.7 | 1.55 | 3.21 |
| $Na_2O$ | 3.02 | 0.61 | 6.50 | 1.76 |
| $K_2O$ | 0.62 | 1.77 | 1.25 | 0.50 |
| $TiO_2$ | 0.80 | 1.00 | 0.04 | 0.65 |
| $P_2O_5$ | 0.11 | 1.80 | 0.02 | 0.07 |
| S | 0.0 | 0.0 | 0.0 | 0.0 |
| MnO | 0.18 | 0.20 | 0.23 | 0.12 |
| $CO_2$ | 0.0 | 0.0 | 0.0 | 0.0 |
| $H_2O^+$ | 3.50 | 6.60 | 1.50 | 6.60 |
| $H_2O^-$ | 0.0 | 0.0 | 0.0 | 0.0 |
| Total | 100.3 | 100.7 | 99.3 | 100.6 |
| Cu | 74 | 192 | 20 | 66 |
| Zn | 90 | 81 | 64 | 118 |
| Ni | 340 | 73 | 24 | 620 |
| Cr | 840 | 236 | 32 | 840 |

7. Conglomerate
8. Basaltic komatiite
9. Calc-alkalic rhyolite
10. Argillite

### NORMS MOLECULAR

|  | 7 | 8 | 9 | 10 |
|---|---|---|---|---|
| AP | 0.238 | 4.013 | 0.042 | 0.157 |
| PO | 0.0 | 0.0 | 0.0 | 0.0 |
| IL | 1.152 | 1.483 | 0.056 | 0.967 |
| OR | 3.792 | 11.146 | 7.482 | 3.158 |
| AB | 28.037 | 5.831 | 59.060 | 16.876 |
| AN | 21.339 | 12.376 | 5.122 | 16.522 |
| C | 0.0 | 0.0 | 0.0 | 3.959 |
| AC | 0.0 | 0.0 | 0.0 | 0.0 |
| MT | 2.487 | 2.783 | 0.463 | 2.401 |
| HM | 0.0 | 0.0 | 0.538 | 0.0 |
| WO | 0.0 | 0.0 | 0.0 | 0.0 |
| EN | 27.220 | 1.705 | 0.106 | 31.548 |
| FS | 8.990 | 0.299 | 0.0 | 7.638 |
| Q | 6.556 | 0.0 | 25.106 | 16.776 |
| DI | 0.142 | 35.900 | 2.023 | 0.0 |
| FO | 0.0 | 15.461 | 0.0 | 0.0 |
| FA | 0.0 | 2.710 | 0.0 | 0.0 |
| NE | 0.0 | 0.0 | 0.0 | 0.0 |
| LC | 0.0 | 0.0 | 0.0 | 0.0 |
| KP | 0.0 | 0.0 | 0.0 | 0.0 |
| HE | 0.047 | 6.293 | 0.0 | 0.0 |
| CC | 0.0 | 0.0 | 0.0 | 0.0 |
| RU | 0.0 | 0.0 | 0.0 | 0.0 |
| NS | 0.0 | 0.0 | 0.0 | 0.0 |
| KS | 0.0 | 0.0 | 0.0 | 0.0 |
| CR | 0.0 | 0.0 | 0.0 | 0.0 |
| LN | 0.0 | 0.0 | 0.0 | 0.0 |

## DAY 5 - KOMATIITES AND SEDIMENTS OF THE ABITIBI BELT IN THE KIRKLAND LAKE AREA

On the return trip to Ottawa we will visit an outcrop area in the Abitibi belt southeast of Kirkland Lake. The aim of the stop will be to show the relationship of komatiites in Archean volcanic sequences and their relationships to other rock types.

In the Abitibi belt, komatiites form the initial phases of several major volcanic cycles. Each cycle consists of a lower komatiitic sequence followed in turn by tholeiitic, calc-alkalic, and, locally, alkalic phases. Two such cycles are present in the Kirkland Lake area. At this stop komatiites of the second cycle overlie rhyolites of the first cycle, indicating that these komatiites are not primitive crust nor were they deposited on primitive crust. It is possible that in the central parts of the basin komatiitic lavas were deposited directly on primitive Archean crust, but at this locality they overlap a pre-existing volcanic pile. Erosion of this volcanic pile occurred, and hence, turbiditic sediments with detritus from both the komatiites and rhyolites are interbedded with the volcanic rocks.

## STOP 5-1

Stop 1 shows pyroxene spinifex toward the top of a basaltic komatiite flow (analysis 8). Overlying the flow is massive calc-alkaline rhyolite tuff breccia(?). An amygdaloidal dyke cuts both the basaltic komatiite and the calc-alkalic rhyolite.

## STOP 5-2

Stop 2 shows the rhyolite tuff (analysis 9) interlayered with conglomerate composed mainly of ultramafic clasts. The conglomerates contain clasts of ultramafic and basaltic komatiite and magnesium-rich tholeiite, i.e. all the mafic volcanic rock-types found in the area. Clasts with olivine and pyroxene spinifex textures are present in the conglomerate. A large rhyolite-ultramafic komatiite boulder 3 x 2 m in size, occurs in the upper conglomerate. It may be noted mafic volcanic clasts are absent in the rhyolite tuff.

Many of the ultramafic conglomerates are difficult to distinguish from polysutured ultramafic flows and where slightly sheared and altered, are indistinguishable unless isolated spinifex clasts or clasts of rhyolite or trachyte can be observed.

STOP 5-3

Stop 3 shows a small peridotite intrusion cutting the sedimentary rocks. It is one of many found in the basinal area. Its texture and appearance are quite distinctive from the ultramafic komatiites.

STOP 5-4

Stop 4 shows a finely bedded turbiditic sequence. Flame structures and many other features associated with such rocks can be observed at this stop.

Some beds of the outcrop were disturbed during their deposition.

The chemical composition of the argillite (analysis 10) has the composition of basaltic komatiite. Analysis 7, is of a similar sedimentary rock found lower in the sequence.

STOPS 5-5 and 6

Stops 5 and 6 are coarser grained turbidites which have graded bedding, channel scouring; laminations can be seen at the tops of many beds. At stop 6, the sedimentary rocks range from boulder conglomerate to fine grained argillite.

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## REFERENCES

Attoh, K., 1980, Stratigraphic relations of the volcanic-sedimentary successions in the Wawa greenstone belt, Ontario; in Current Research, Part A; Geological Survey of Canada, Paper 80-IA, p. 101-106.

Ayres, L.D., 1969, Geology of Townships 31 and 30, Ranges 20 and 19; Ontario Department of Mines Geological Report 69, 100 p.

Bennett, G., Brown, D.D., George, P.T. and Leahy, E.J. 1967, Operation Kapuskasing; Ontario Department of Mines Miscellaneous Paper 10, 98 p.

Berry, M.J. and Mair, J.A., 1980, Structure of the continental crust: a reconciliation of the seismic reflection and refraction studies; in The Continental Crust and its Mineral Deposits (ed. D.W. Strangway); Geological Association of Canada, Special Paper 20, p. 149-180.

Binns, R.A., 1964, Zones of progressive regional metamorphism in the Willyama complex, Broken Hill District, New South Wales; Journal of the Geological Society of Australia II, p. 283-330.

Bohlen, S.R. and Essene, E.J., 1977, Feldspar and oxide thermometry of granulites in the Adirondack Highlands; Contributions to Mineralogy and Petrology 62, p. 153-169.

Burke, K. and Dewey, J.F., 1973, Plume-generated triple junctions: Key indicators in applying plate tectonics to old rocks; Journal of Geology, v. 81, p. 406-433.

Card, K.D., 1982, Progress report on regional geological synthesis, central Superior Province, in Current Research, Part A; Geological Survey of Canada Paper 82-1A, p. 23-28.

Carmichael, D.M., 1978, Metamorphic bathozones and bathograds; A measure of the depth of post-metamorphic uplift and erosion on the regional scale; American Journal of Science, v. 278, p. 767-797.

de Waard, D., 1965, The occurrence of garnet in granulite-facies terrain of the Adirondack Highlands; Journal of Petrology 6, p. 165-191.

Ellis, D.J. and Green, D.H., 1979, An experimental study of the effect of Ca upon garnet-clinopyroxene Fe-Mg exchange equilibria; Contributions to Mineralogy and Petrology 71, p. 13-22.

Ernst, R.E., 1981, Correlation of Precambrian diabase dike swarms across the Kapuskasing structural zone, northern Ontario; Unpublished M.Sc. thesis, University of Toronto, Toronto.

Forsyth, D.A. and Morel, P., 1982, Comparative study of the geophysical and geological information in the Timmins-Kapuskasing area (Abs.); Geological Association of Canada-Mineralogical Association of Canada, Program with Abstracts, v. 7, p. 49.

Fountain, D.M. and Salisbury, M.H., 1981, Exposed cross-sections through the continental crust; Implications for crustal structure, petrology, and evolution; Earth and Planetary Science Letters, v. 56, p. 263-277.

Fraser, J.A., Heywood, W.W. and Mazurski, M., 1978, Metamorphic map of the Canadian Shield; Geological Survey of Canada, Map 1475A, scale 1;3 500 000.

Gates, T.M. and Hurley, P.M., 1973, Evaluation of Rb-Sr dating methods applied to the Matachewan, Abitibi, Mackenzie and Sudbury dike swarms in Canada; Canadian Journal of Earth Sciences, v. 10, p. 900-919.

Gittins, J., MacIntyre, R.M. and York, D., 1967, The ages of carbonatite complexes in eastern Canada; Canadian Journal of Earth Sciences, v. 4, p. 651-655.

Goodwin, A.M., 1962, Structure, stratigraphy and origin of iron formation, Michipicoten area, Algoma district, Ontario, Canada; Geological Society of America Bulletin, v. 73, p. 561-586.

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Gorman, B.E., Pearce, T.H. and Birkett, T.C., 1978, On the structure of Archean greenstone belts; Precambrian Research, v. 6, p. 23-41.

Green, D.H. and Ringwood, A.E., 1967, An experimental investigation of the gabbro to eclogite transformation and its petrological applications; Geochimica et Cosmochimica Acta 31, p. 763-833.

Hall, D.H., and Brisbin, W. C., 1982, Overview of regional geophysical studies in Manitoba and northwestern Ontario; Canadian Journal of Earth Sciences, v. 19, p. 2049-2059.

Hillary, E.M. and Ayres, L.D., 1980, Trondhjemitic basement enclaves near the Archean Favourable Lake volcanic complex, northwestern Ontario; Canadian Journal of Earth Sciences, v. 17, p. 652-667.

Hyde, R.S., 1980, Sedimentary facies in the Archean Timiskaming Group and their implications, Abitibi greenstone belt, northeastern Ontario; Precambrian Research, v. 12, p. 161-195.

Jensen, L.S., 1981, A petrogenetic model for the Archaean Abitibi Belt in the Kirkland Lake area, Ontario; Unpublished Ph.D. thesis, University of Saskatchewan.

Jolly, W.T., 1978, Metamorphic history of the Archean Abitibi belt, in Fraser, J..A., and Heywood, W.W., eds., Metamorphism in the Canadian Shield; Geological Survey of Canada Paper 78-10, p. 63-78.

Krogh, T.E., 1982, Improved accuracy of U-Pb zircon ages by the creation of more concordant systems using an air abrasion technique; Geochimica et Cosmochimica Acta 46, p. 637-649.

Krogh, T.E., Davis, D.W., Nunes, P.D., and Korfu, F., 1982, Archean evolution from precise U-Pb isotopic dating (Abs.); Geological Association of Canada-Mineralogical Association of Canada, Program with Abstracts, v. 7, p. 61.

Kushiro, I. and Yoder, H.S. Jr., 1966, Anorthite-forsterite and anorthite-enstatite reactions and their bearing on the basalt-eclogite transformation; Journal of Petrology, v. 3, p. 337-362.

Lowden, J.A. and Wanless, R.K., 1963, Part 1 - Age determinations by the Geological Survey of Canada; in Age Determinations and Geological Studies; Geological Survey of Canada, Paper 62-17, p. 1-21.

Martingole, J. and Schrijver, K., 1971, Association of (hornblende) garnet-clinopyroxene (subfacies) of metamorphism and anorthosite masses; Canadian Journal of Earth Sciences, v. 8, p. 698-704.

Mattinson, J.M., 1978, Age, origin, and thermal histories of some plutonic rocks from the Salinian block of Californa; Contributions to Mineralogy and Petrology, v. 67, p. 233-245.

Milne, V.G., 1972, Geology of the Kukatush-Sewell area, District of Sudbury; Ontario Division of Mines, Geological Report 97, 116 p.

Newton, R.C., and Perkins, D., III, 1982, Thermodynamic calibration of geobarometers based on the assemblages garnet-plagioclase-orthopyroxene(clinopyroxene)-quartz; American Mineralogist, v. 67, p. 203-222.

Nunes, P.D., and Jensen, L.S., 1980, Geochronology of the Abitibi metavolcanic belt, Kirkland Lake area - Progress Report; in Pye, E. G., ed., Summary of geochronology studies, 1977-1979; Ontario Geological Survey Miscellaneous Paper 92, p. 40-45.

Nunes, P.D., and Pyke, D.R., 1980, Geochronology of the Abitibi metavolcanic belt, Timmins-Matachewan area - Progress Report, in Pye, E. G., ed., Summary of geochronology studies, 1977-1979; Ontario Geological Survey Miscellaneous Paper 92, p. 34-39.

Percival, J.A., 1981, Geological evolution of part of the central Superior Province based on relationships among the Abitibi and Wawa subprovinces and the Kapuskasing structural zone (Ph.D. thesis); Kingston, Queen's University, 300 p.

Percival, J.A., 1981b, Geology of the Kapuskasing Structural zone in the Chapleau-Foleyet area; Geological Survey of Canada, Open File Map 763.

Percival, J.A., 1983, High-grade metamorphism in the Chapleau-Foleyet area, Ontario; American Mineralogist, v. 68, p. 667-686.

Percival, J.A., and Coe, K., 1980, Geology of the Kapuskasing structural zone in the Chapleau-Foleyet area, Ontario, in Card, K. D., Percival, J. A., and Coe, K., Progress report on regional geological synthesis, Central Superior Province, in Current Research, Part A; Geological Survey of Canada, Paper 80-1A, p. 61-68.

Percival, J.A. and Coe, K., 1981, Parallel evolution of Archaean low- and high-grade terrane; a view based on relationships between the Abitibi, Wawa and Kapuskasing belts; Precambrian Research, 14, p. 315-331.

Percival, J.A. and Krogh, T.E., 1983, U-Pb zircon geochronology of the Kapuskasing structural zone and vicinity in the Chapleau-Foleyet area, Ontario; Canadian Journal of Earth Sciences, v. 20 (in press).

Percival, J.A.., Loveridge, W.D. and Sullivan, R.W., 1981, U-Pb zircon ages of tonalitic metaconglomerate cobbles and quartz monzonite from the Kapuskasing structural zone in the Chapleau area, Ontario; in Rb-Sr and U-Pb Isotopic age Studies, Report 4; in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 107-113.

Perkins, D. and Newton, R.C., 1981, Charnockite geobarometry based on coexisting garnet-pyroxene-plagioclase-quartz; Nature, v. 292, p. 144-146.

Poulsen, K.H., Borradaile, G.J. and Kehlenbeck, M.M., 1981, An inverted Archean succession at Rainy Lake, Ontario; Canadian Journal of Earth Sciences, v. 17, p. 1358-1369.

Pyke, D.R., MacVeigh, J.G. and Middleton, R.S., 1978, Volcanic stratigraphy and geochemistry in the Timmins mining area; in Toronto '78 Field Trips Guidebook (eds. R.L. Currie and W.O. Mackasey); Geological Association of Canada Guidebook, p. 160-184.

Pyke, D.R., 1982, Geology of the Timmins area, District of Cochrane; Ontario Geological Survey Report 219, 141 p.

Riccio, L., 1981, Geology of the northeastern portion of the Shawmere anorthosite complex, District of Sudbury; Ontario Geological Survey, Open File Report 5338, 113 p.

Ringwood, A.E., 1975, Composition and Petrology of the Earth's Mantle; McGraw-Hill, New York, 618 p.

Sage, R.P., 1980, Wawa area, District of Algoma, p. 47-50; in Summary of field work, 1980 (ed. V.G. Milne, O.L. White, R.B. Barlow, J.A. Robertson, and A.C. Colvine), Ontario Geological Survey, Miscellaneous Paper 96.

Schwerdtner, W.M. and Lumbers, S.B., 1980, Major diapiric structures in the Superior and Grenville provinces of the Canadian Shield; in The continental Crust and its Mineral Deposits (ed. D.W. Strangway); Geological Association of Canada, Special Paper 20, p. 149-180.

Schwerdtner, W.M., Stone, D., Osadetz, K., Morgan, J. and Stott, G.M., 1979, Granitoid complexes and the Archean tectonic record in the southern part of northwestern Ontario; Canadian Journal of Earth Sciences, v. 16, p. 1965-1977.

Simmons, E.C., Hanson, G.N., and Lumbers, S.B., 1980, Geochemistry of the Shawmere anorthosite complex, Kapuskasing structural zone, Ontario; Precambrian Research, v. 11, p. 43-71.

Smithson, S.B. and Brown, S.K., 1977, A model for lower continental crust; Earth and Planetary Science Letters, v. 35, p. 134-144.

Smithson, S.B., Brewer, J., Kaufman, S., Oliver, J., and Hurich, C., 1978, Nature of the Wind River thrust, Wyoming, from COCORP deep-reflection data and from gravity data; Geology, v. 6, p. 648-652.

Stevens, R.D., Delabio, R.N. and Lachance, G.R., 1982, Age determinations and Geological Studies, K-Ar isotopic ages, Report 15; Geological Survey of Canada, Paper 81-2, 56 p.

Sutcliffe, R. and Fawcett, J.J., 1979, Petrological and geochronological studies on the Rainy Lake granitoid complex, northwestern Ontario; a progress report; in Current Research, Part A, Geological Survey of Canada, Paper 79-1A, p. 377-380.

Thurston, P.C. and Breaks, F.W., 1978, Metamorphic and tectonic evolution of the Uchi-English River subprovince; in Metamorphism in the Canadian Shield; Geological Survey of Canada, Paper 78-10, p. 49-62.

Thurston, P.C., Siragusa, G.M., and Sage, R.P., 1977, Geology of the Chapleau area, Districts of Algoma, Sudbury and Cochrane; Ontario Division of Mines Geological Report 157, 293 p.

Turek, A., Smith, P.E., and Van Schmus, W.R., 1982, Rb-Sr and U-Pb ages of volcanism and granite emplacement in the Michipicoten belt, Wawa, Ontario; Canadian Journal of Earth Sciences, v. 19, p. 1608-1625.

Turek, A., Smith, P.E. and Van Schmus, W.R., 1983, The evolution in time of the Wawa-Gamitagama plutonic-volcanic terrain, Superior Province - Northern Ontario (Abs.); Geological Association of Canada-Mineralogical Association of Canada, Program with Abstracts, v. 8, p. A70.

Turner, F.J., 1981, Metamorphic Petrology; Mineralogical, Field and Tectonic

Aspects, second edition; McGraw-Hill, New York, 524 p.

Van Schmus, W.R., 1975, On the age of the Sudbury dike swarm; Canadian Journal

of Earth Sciences, v. 12, p. 1690-1692.

Watson, J., 1980, The origin and history of the Kapuskasing structural zone,

Ontario, Canada; Canadian Journal of Earth Sciences, v. 17, p. 866-876.

Whitney, P.R., 1978, The significance of garnet "isograds" in granulite facies rocks

of the Adirondacks; in Metamorphism in the Canadian Shield; Geological Survey

of Canada, Paper 78-10, p. 357-366.

Wilson, H.D.B., and Brisbin, W.C., 1965, Mid-North American ridge structure

(Abs.); Geological Society of America Special Paper 87, p. 186-187.

Winkler, H.G.F., 1979, Petrogenesis of Metamorphic Rocks, 5th edition; Springer-

Verlag, Berlin, Heidelburg, New York.