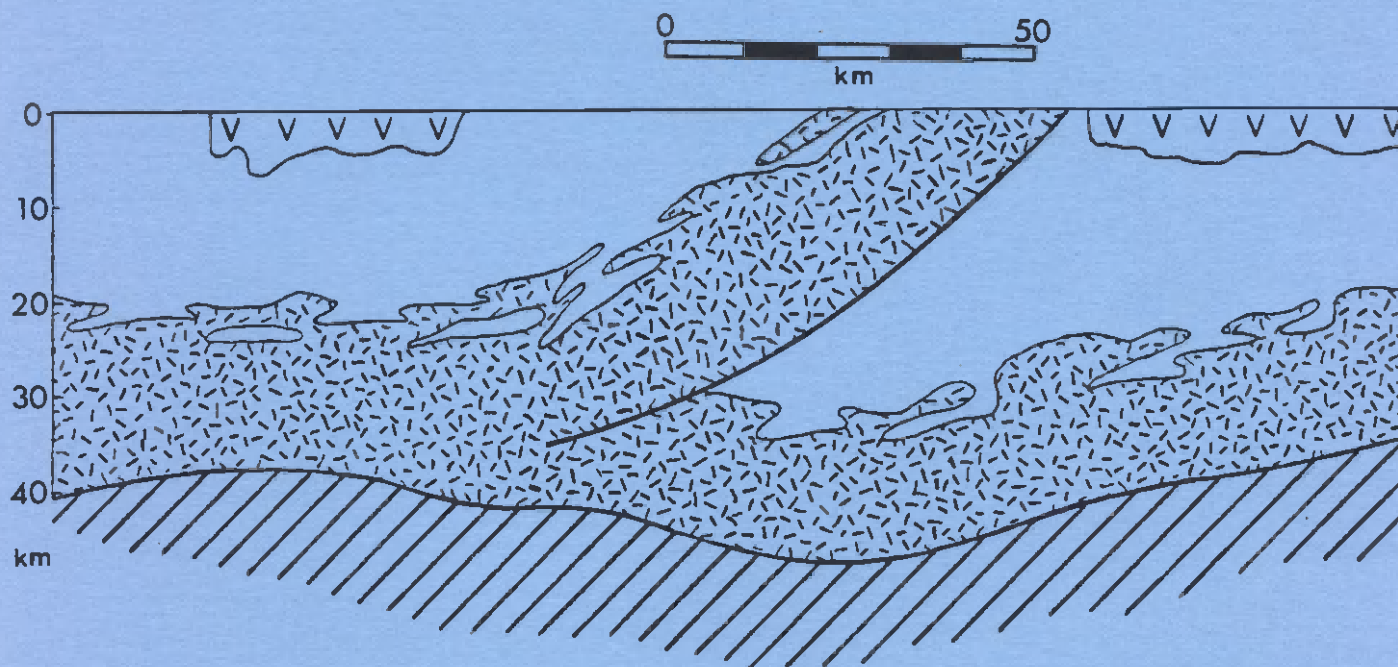




Institute on Lake Superior Geology Thirty-Third Annual Meeting

THE KAPUSKASING UPLIFT: ARCHEAN
GREENSTONES AND GRANULITES



Wawa, Ontario.

Vol. 33, Part 5

**THE KAPUSKASING UPLIFT: ARCHEAN
GREENSTONES AND GRANULITES**

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PART I: GEOLOGICAL FRAMEWORK OF THE KAPUSKASING UPLIFT

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, and contact relationships between the Michipicoten supracrustal rocks and intrusions of the Wawa domal gneiss terrane. On day two, the Wawa-Kapuskasing boundary will be examined, as well as features of the Kapuskasing structural zone, including the Shawmere anorthosite complex, high-grade gneisses, and the Ivanhoe Lake cataclastic zone separating rocks of the Kapuskasing zone from those of the Abitibi Belt.

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). At its southern end, 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 (Percival and Card, 1983; 1985) (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 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 Proterozoic sinistral transcurrent movement. The low-to-high-grade transition at the southern end of the structure has been interpreted as an oblique cross-section of the crust uplifted along an east-verging thrust (Percival and Card, 1983). Recent earthquakes in the region indicate that the structure is still active (Forsyth and Morel, 1982; Forsyth et al., 1983).

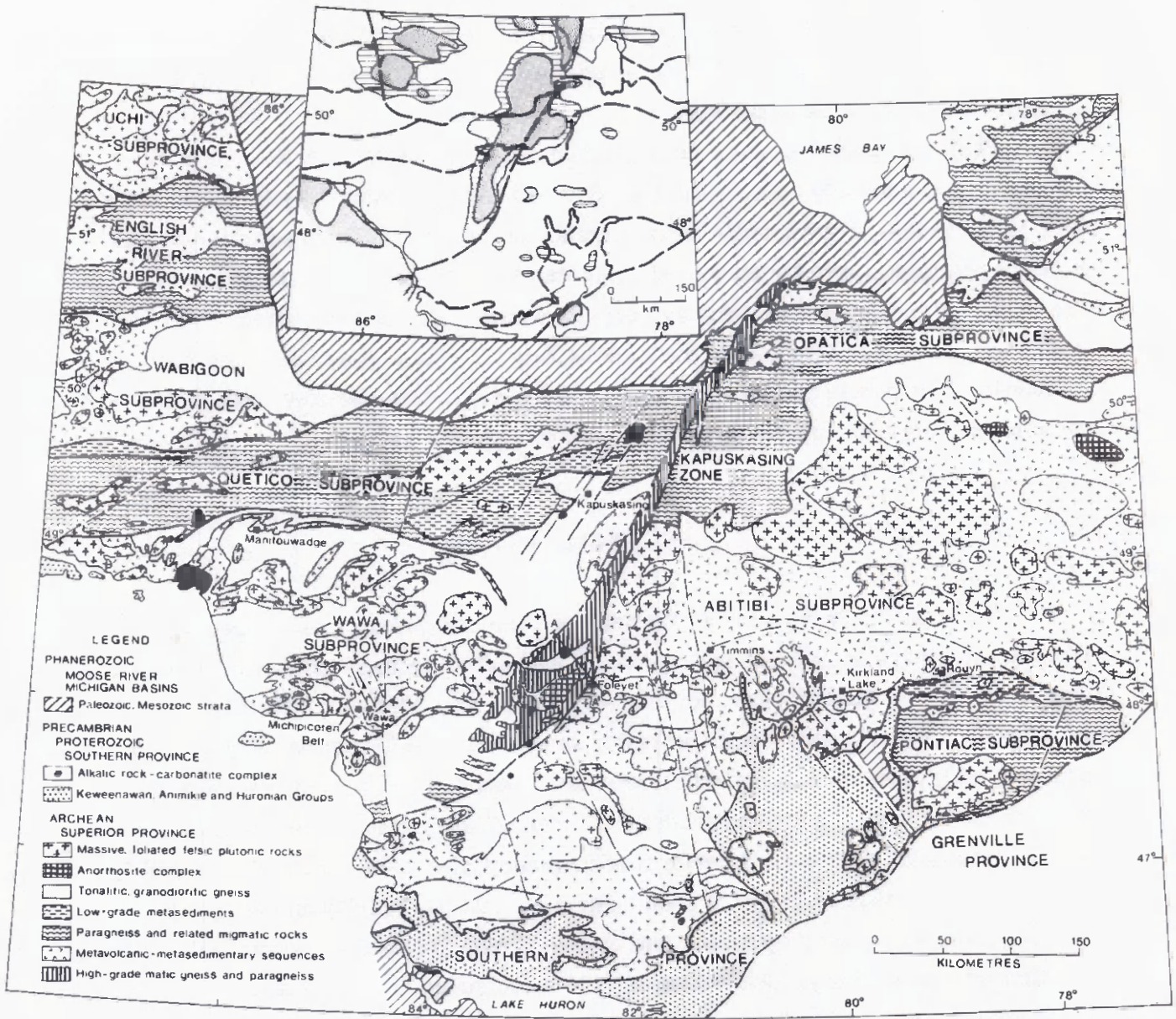


Figure 1. Geology of the central Superior Province. Inset: positive gravity anomalies dark stipple (>-25 mGal); lines (-25 mGal to -35 mGal); blank (<-35 mGal).

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

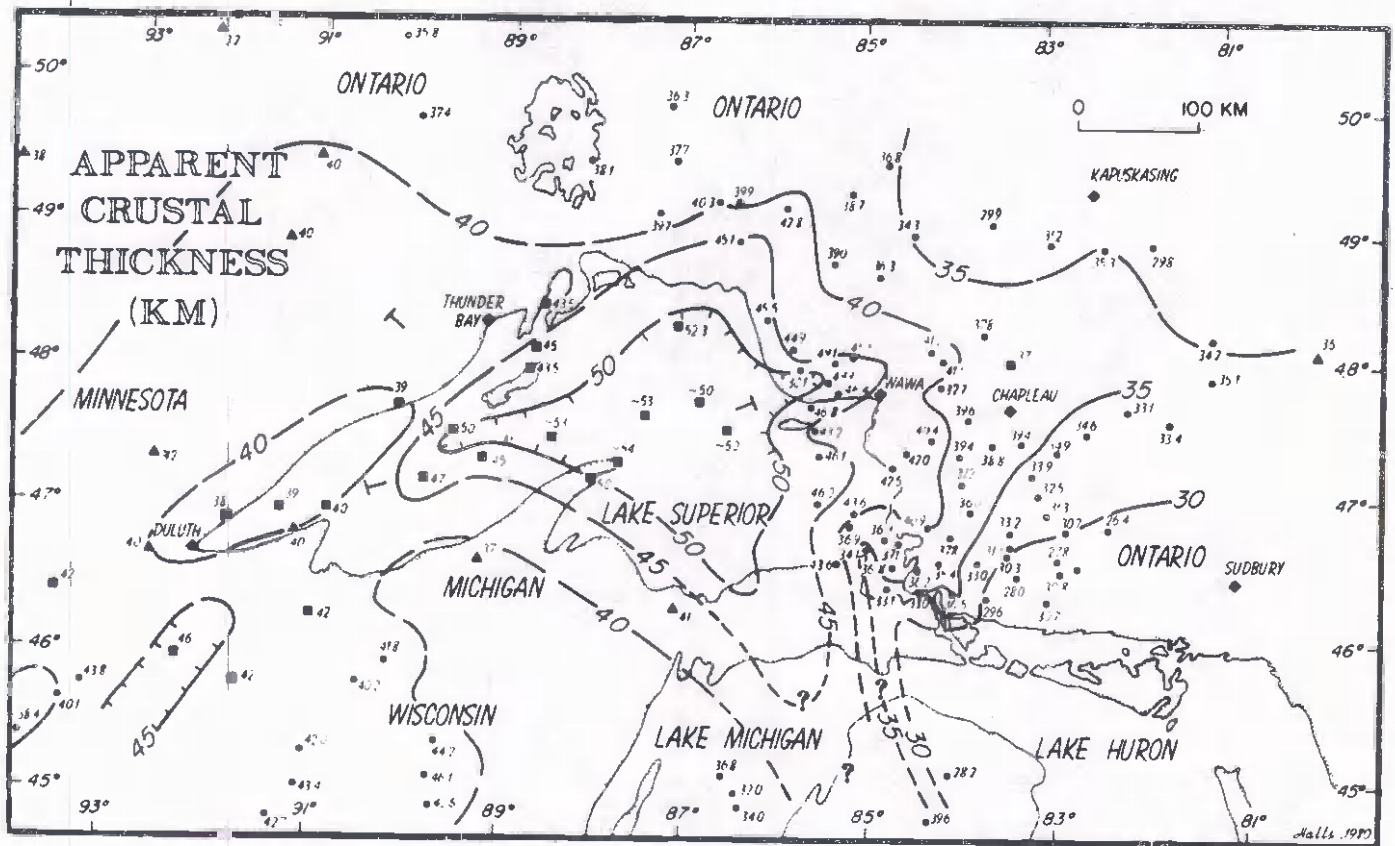


Figure 3 Apparent thickness of the crust in the Lake Superior area (after Halls, 1982). Sources of data are listed by Halls and extreme caution in using map was advised.

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.

Results of a major refraction study, conducted in 1984 to examine crustal structure beneath the Kapuskasing zone, are still in preliminary stages of interpretation (Northey et al., 1985; Northey and West, 1986; Boland and Ellis, 1986). Initial

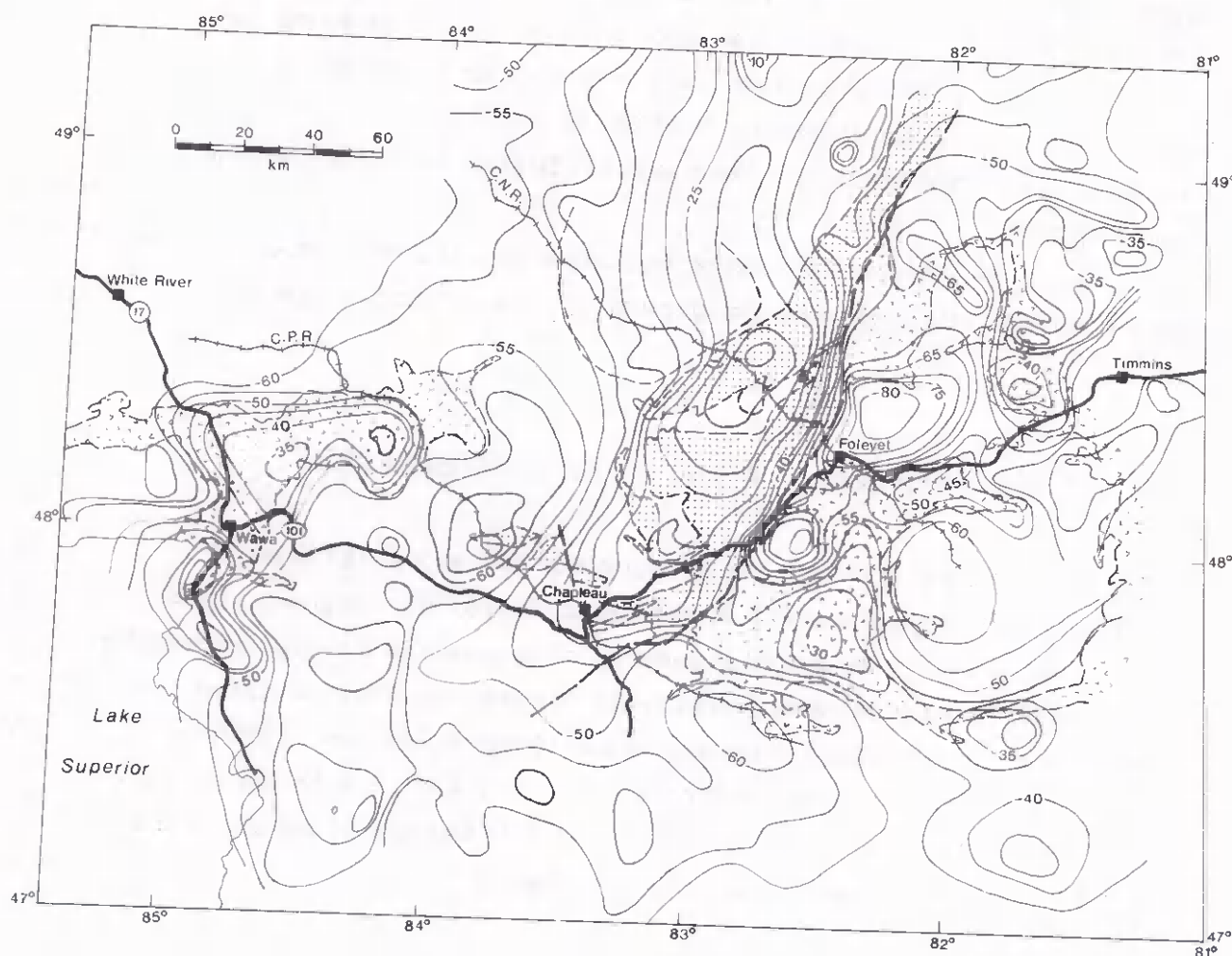


Figure 4. Bouguer gravity map superimposed on simplified geology. (Gravity values in mGal), with 5 mGal contour interval are from Earth Physics Branch maps 44078, 44084, 48078 and 48084). Checks - greenstone belts; dots -Kapuskasing gneiss; unpatterned-undivided granitoid rocks.

interpretation indicates that the crust is at least 48 km thick beneath the Kapuskasing zone and thins abruptly to about 40 km to the southeast; high seismic velocities characterize the high-grade rocks of the Kapuskasing zone (D. Northey, oral communication, 1985; Cook, 1985; Fountain and Salisbury, 1986).

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. 12) shows a paired high-low anomaly. The trough of the low is coincident with the fault at the eastern boundary of the Kapuskasing zone.

To the north, the positive Kapuskasing anomaly broadens as it coalesces with the east-west gravity high associated with the Quetico-Opatika metasedimentary subprovince.

GENERAL GEOLOGY OF THE WAWA - ABITIBI REGION

Three distinct types of terrane are recognized in this part of the Superior Province: 1) greenstone-granite belts, 2) regions dominated by orthogneiss in the amphibolite facies, and 3) heterogeneous gneisses in the granulite to upper amphibolite facies. Differences between terranes in metamorphic grade and pressure suggest that the terrane types represent crustal-scale megalayers (Percival and Card, 1985), the components of the upper and middle continental crust. The Wawa and Abitibi belts are greenstone-granite terranes; the Wawa gneiss terrane is an example of terrane type 2 and the Kapuskasing zone represents type 3 terrane (Fig. 2).

1) Greenstone-Granite Terranes

The Michipicoten belt, part of the volcanic-rich Wawa subprovince, is composed mainly of metavolcanic rocks of ultramafic, mafic and felsic composition (Goodwin, 1962), with intercalated greywacke, conglomerate, chert and iron formation, mainly siderite. 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 supracrustal rocks of the southern part of the belt were divided into three major cycles by Goodwin (1962). 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 associated Jubilee Stock, a high-level subvolcanic intrusion, was emplaced within a caldera structure (Sage, 1980). 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. Recent work by Sage (pers. comm., 1986) indicates only two cycles within the main volcanic pile.

The Abitibi subprovince is dominated by a thick sequence of volcanic and sedimentary rocks of the Abitibi greenstone belt (Jensen, 1981; 1985). 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, probably has both 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), with late alkaline volcanics (Timiskaming Group) at 2685 Ma, and in the Michipicoten belt, from 2,749 to 2,696 Ma, with synvolcanic plutons at 2737, 2744 (Turek et al., 1982) and 2745 Ma (Sullivan et al., 1985). 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; Frarey and Krogh, 1986). Thus the main 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; Ernst and Halls, 1984). 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).

2) Wawa Gneiss Terrane

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. Tonalite gneiss has a minimum U-Pb zircon age of 2707 Ma, partly reset by the intrusion of granodiorite sheets at 2677 Ma (Percival and Krogh, 1983) (Fig. 5). The layers of granodiorite on the cm to km scale in the gneiss terrane can be correlated by zircon geochronology with discordant plutons of 2680 Ma age in the greenstone belts, suggesting that the plutons have deep roots in the gneissic terrane.

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 (Fig. 6). 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 structural domes may be related to the emplacement of plutons, possibly by diapiric rise. 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.

A geobarometer for calc-alkaline igneous rocks is based on the Al content of hornblende (Hammarstrom and Zen, 1986):

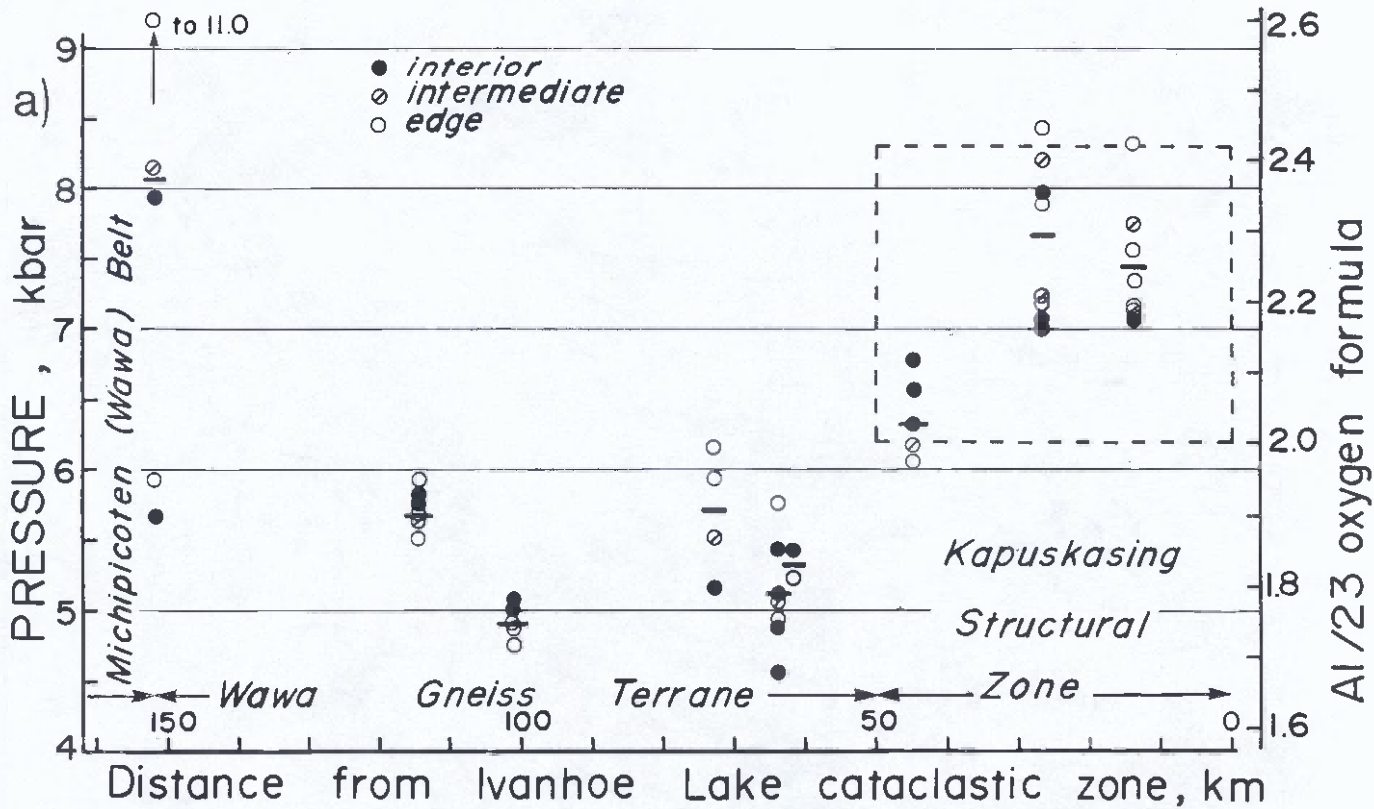
$$P = -3.92 + 5.03 \text{ Al}^{\text{Total}}$$

Application of the barometer to a suite of tonalites from the Wawa gneiss terrane suggests that the pressure of igneous crystallization increases from about 5 kbar in the central part of the terrane to over 6 kbar near the Kapuskasing zone (Fig. 5). These results are intermediate between independent pressure estimates for the Michipicoten belt of 2-3 kbar, based on sphalerite-pyrrhotite geobarometry (Studemeister, 1983) and for the Kapuskasing zone of 6-8 kbar based on garnet-pyroxene-plagioclase-quartz barometry (Percival, 1983). A value of 8 kbar near Wawa is derived from a biotite tonalite which contains hornblende only adjacent to contacts with amphibolite; the hornblende is probably xenocrystic.

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 6). 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, 1983; 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. 6) may represent the eastern extremity



Al in Hornblende Barometer

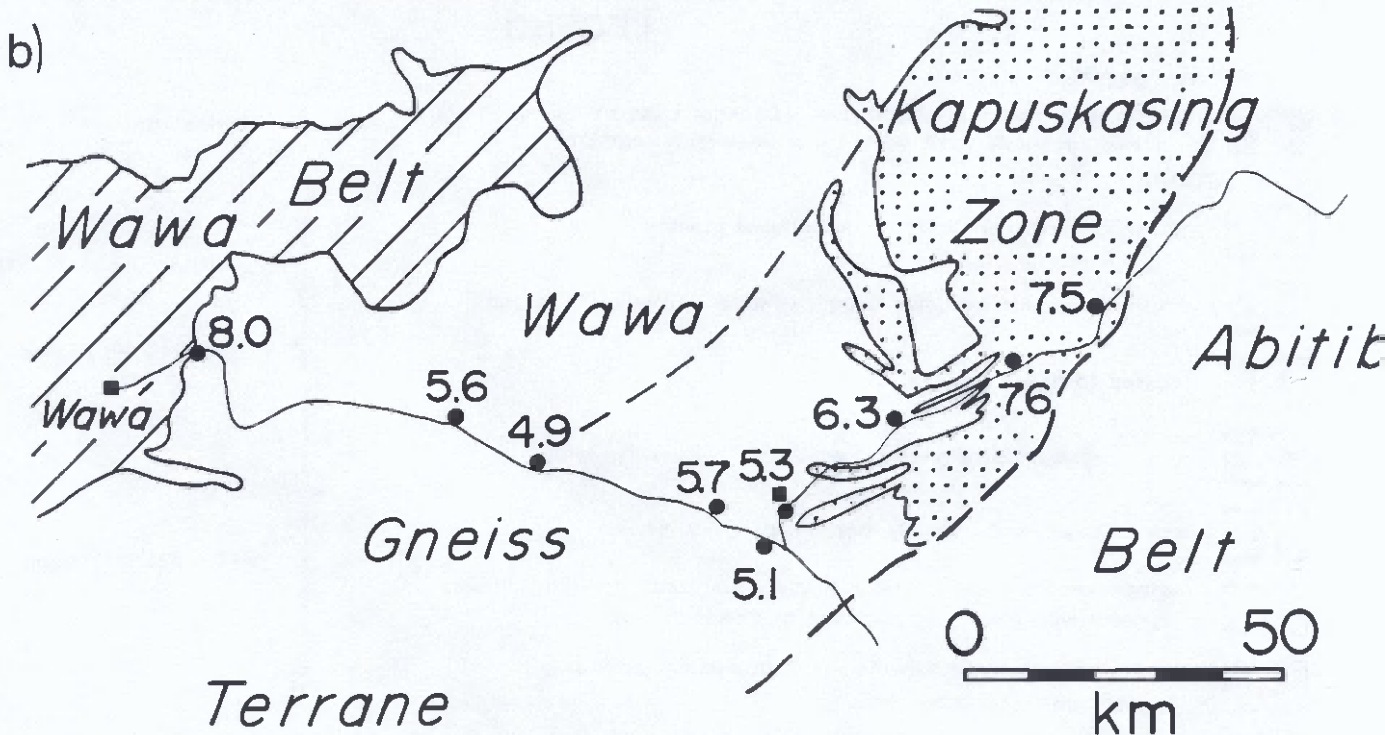
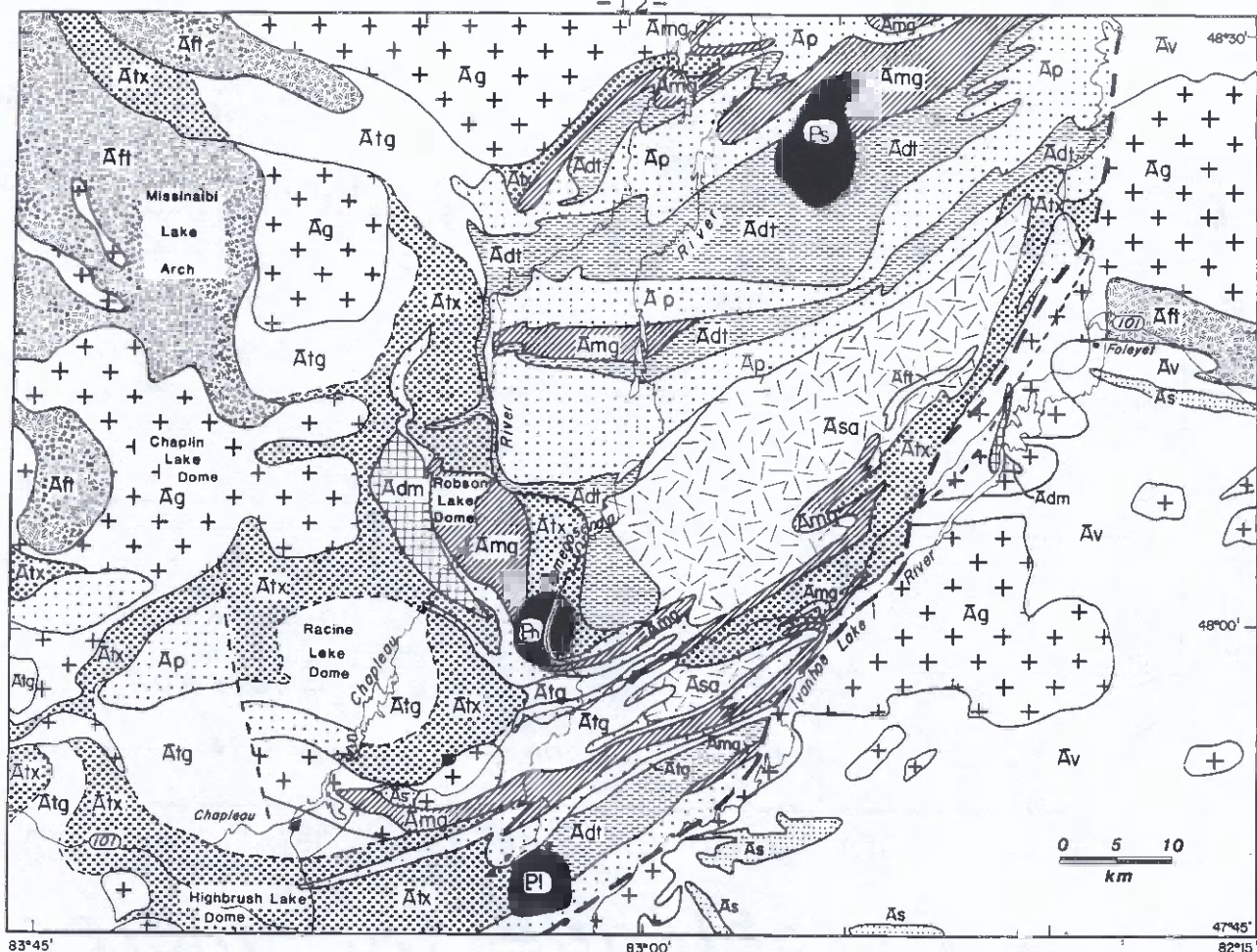


Figure 5. a) Plot of hornblende composition vs spatial distribution. Vertical array of points represents within-sample compositional variation. Hornblende at far left is probably xenocrystic, from adjacent amphibolites of Michipicoten belt. Dashed-line box represents independent pressure estimates from granulites of the Kapuskasing zone.
b) Map showing average pressure in kbar, based on hornblende geobarometry in the Wawa-Chapleau area.



LEGEND

Proterozoic



Alkalic rock-carbonatite complex: l: Lackner Lake complex;
n: Nemegosenda Lake complex; s: Shenango complex

1100 Ma intrusions

Archean



massive granite, granodiorite, with minor tonalite



diorite-monzonite intrusive complex; minor hornblendite, granite



foliated to flaser tonalite



tonalite-granodiorite gneiss; xenolithic



metavolcanic rocks, mainly metabasalt



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

2749-2696 Ma sequence



flaser diorite to mafic tonalite - includes minor gabbro,
hornblendite, granodiorite



Shawmere anorthosite complex: metamorphosed gabbroic anorthosite,
anorthosite, gabbro, minor tonalite



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

pre-2765 Ma sequence



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



fault; Ivanhoe Lake cataclastic zone

Figure 6. Geology of the Kapuskasing structural zone and vicinity.

of a first-order dome of 75-100 km diameter, of which the individual structural domains are higher-order domes of similar scale and spacing to those of the Wabigoon subprovince (Schwerdtner and Lumbers, 1980).

Dome development can be temporally related 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, possibly relating to late diapiric rise.

3) 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 6).

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_±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% Al₂O₃) basalt

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).

	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 ₂ O ₃	15.5	15.6	17.2	13.4	16.2	17.0
Fe ₂ O 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 ₂ O	2.0	2.4	2.8	1.6	2.3	2.8
K ₂ O	0.25	0.41	0.12	0.58	0.33	0.2
H ₂ O	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					
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	
HE	12.80	10.90	8.28	6.41	10.58	18.9
EN	6.05	3.64	5.80	5.39	4.96	
FS	7.83	3.95	7.89	3.89	4.43	9.4
FO		3.01		9.26	1.54	
FA		3.60		7.36	1.21	9.3
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). 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 mafic sedimentary parentage for the rock type.

In the area of Figure 6, four linear, northeast-striking bodies of foliated diorite and mafic tonalite occur dominantly within paragneiss terranes. These

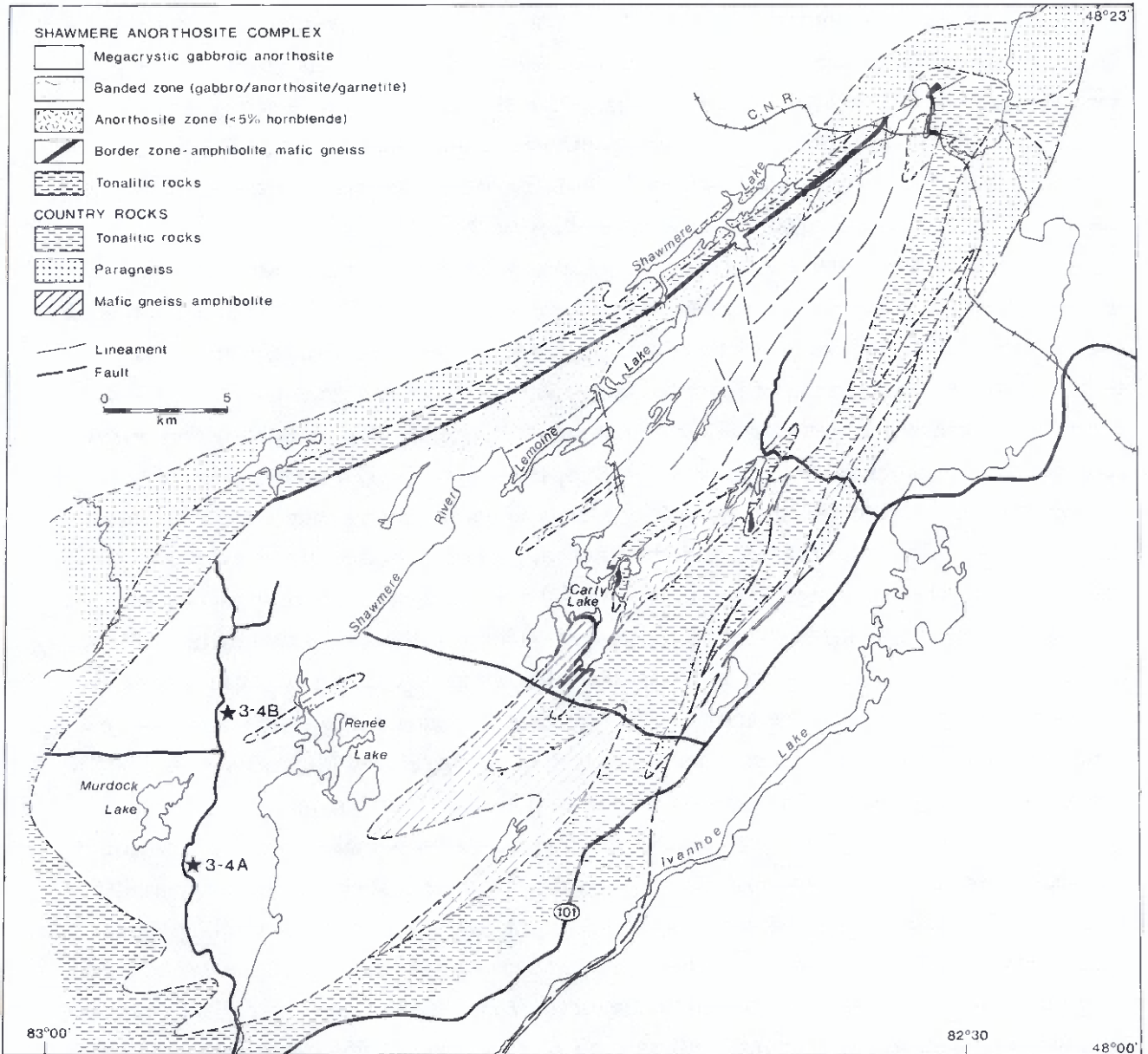


Figure 7. Geology of the Shawmere anorthosite complex (after Riccio, 1981 and Percival, 1981).

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.

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. 7): (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 monoclinial flexures. Axial surfaces are rarely accompanied 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 100 m scale. Both regional and local plunge reversals can be related to a gently southeast-plunging warp axis.

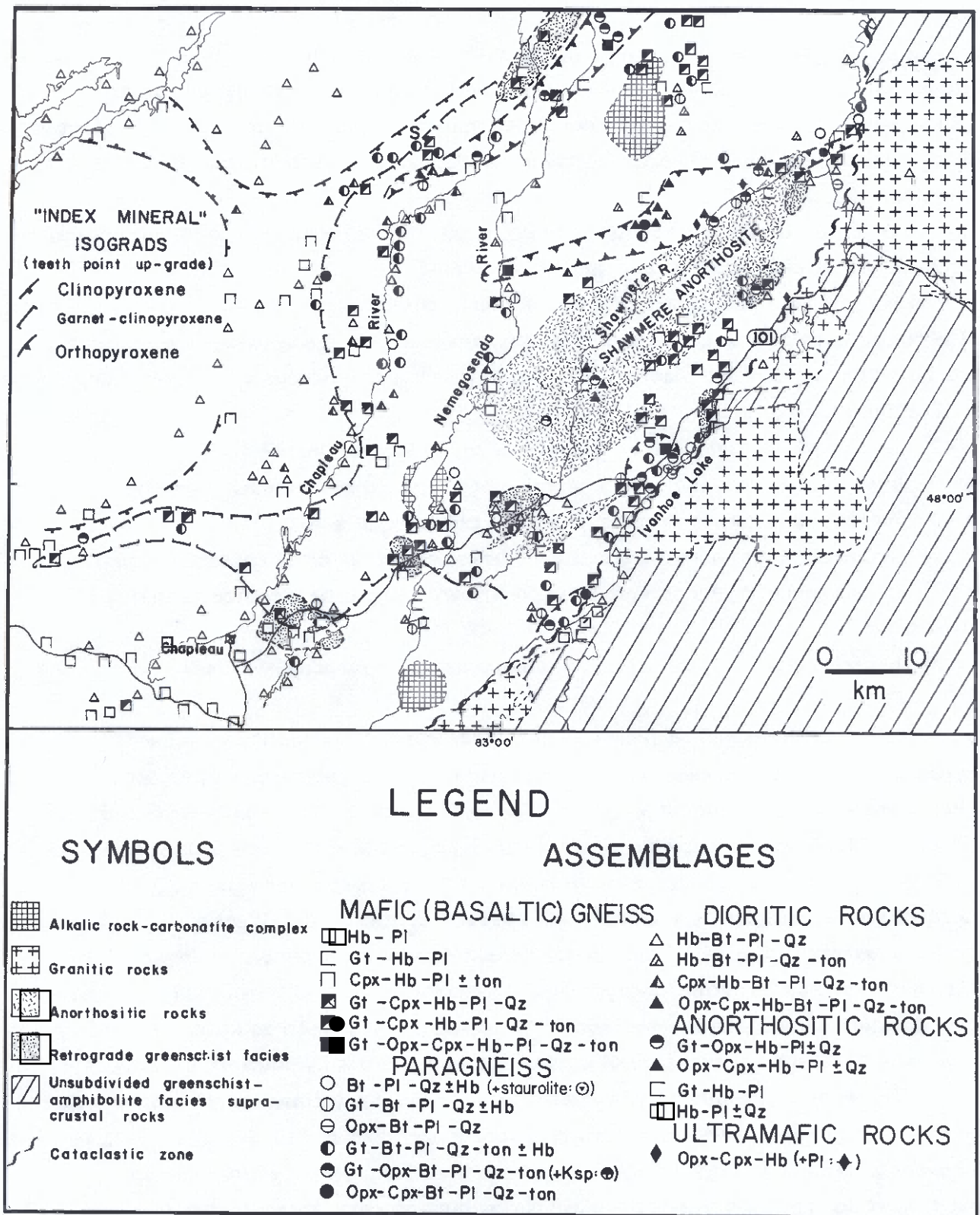


Figure 8. Metamorphic mineral assemblages and index mineral isograds for part of the Chapleau-Foley area. Gt - garnet; Opx - orthopyroxene; Cpx - clinopyroxene; Hb - hornblende; Bt - biotite; Pl - plagioclase; Ksp - feldspar; Qz - quartz; ton - tonalitic segregations. (after Percival, 1983)

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. 8; Percival, 1983).

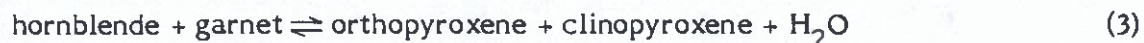
A continuous reaction resulting in decomposition of hornblende in mafic rocks to produce garnet and clinopyroxene may be written:



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. 9):



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



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:



A P-T diagram summarizing continuous reactions in the mafic system and apparent metamorphic conditions based on various mineral geothermometers and geobarometers, is presented in Figure 9. Apparent pressures, based on Newton and Perkins' (1982) garnet-clinopyroxene-plagioclase quartz barometer, are plotted on a map in Figure 10 and have an average value of 6.3 kbar. This barometer apparently underestimates pressure by 1-1.6 kbar (Newton and Perkins, 1982; Ghent et al., 1983) and hence an average value of 7.8 kbar is suggested. Apparent temperatures, based on the Ellis and Green (1979) garnet-clinopyroxene thermometer (Fig. 9) are in the range 700-800°C. Metamorphic fluids were probably depleted in H₂O, based on water barometry (Percival, 1983) and the presence of carbonic fluid inclusions (Rudnick et al., 1984).

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) suggested 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,

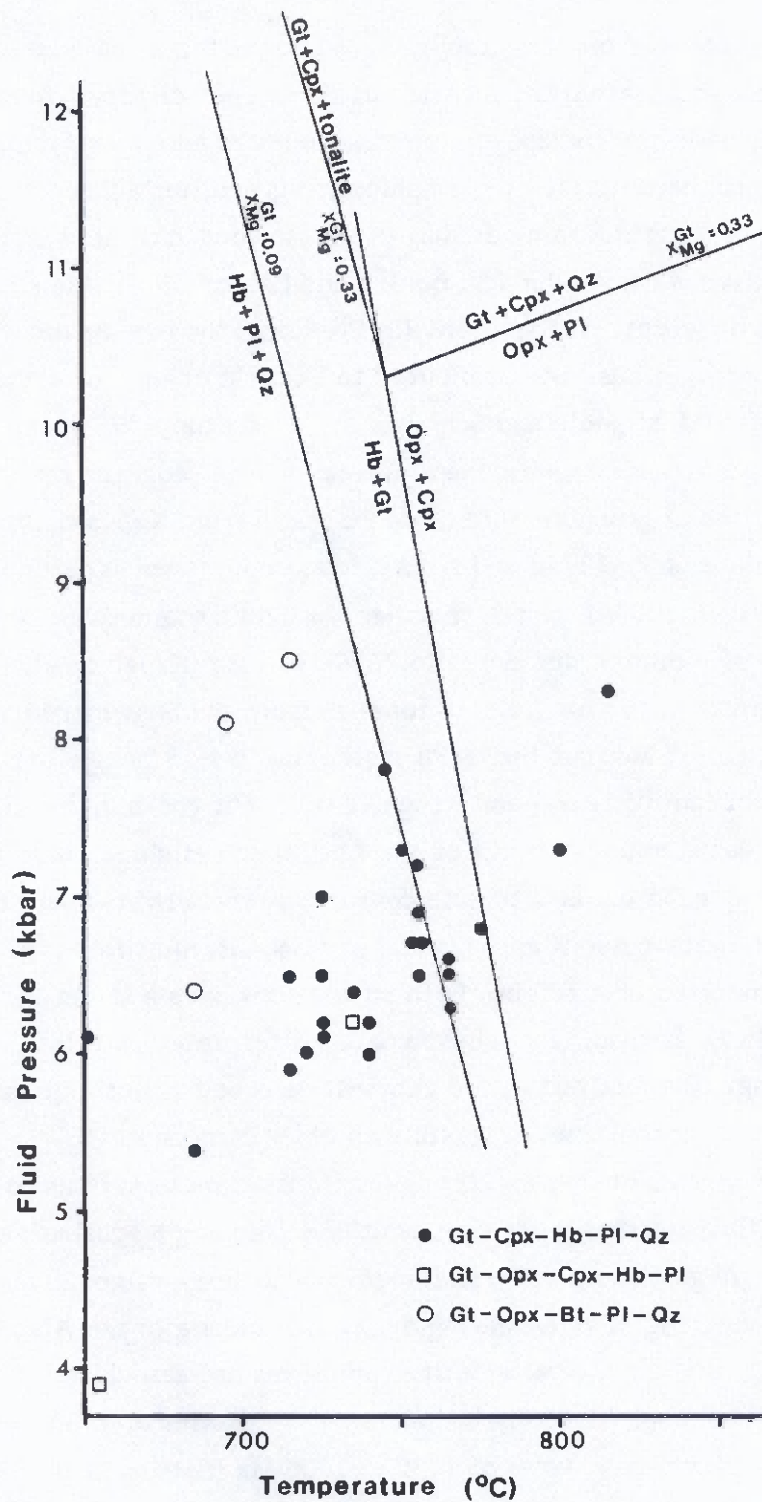


Figure 9. Summary of reactions applicable to mafic rocks and metamorphic pressure-temperature estimates. Temperatures are derived from the garnet-clinopyroxene thermometer (Ellis and Green, 1979) and pressures from garnet-pyroxene-plagioclase-quartz barometers (Newton and Perkins, 1982).

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. 11). Further work on metamorphic zircon has extended the range of metamorphic dates to 2696-2616 Ma (Krogh et al, 1986). 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.

Quartz-bearing gabbroic anorthosite contains zircon with two habits: red-stained resorbed grains with meta-igneous appearance, and equant, multifaceted, colourless grains of probable metamorphic origin. Both populations have a U-Pb age of 2649 Ma (J.A. Percival and R.W. Sullivan, unpublished data), interpreted as the time of metamorphic cooling. The analysis of the coarsest resorbed grains plots slightly to the right of the 2649 Ma discordia line, suggesting an older component.

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.

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-Foley area gave a K-Ar date of 1144 ± 31 Ma (Stevens et al., 1982).

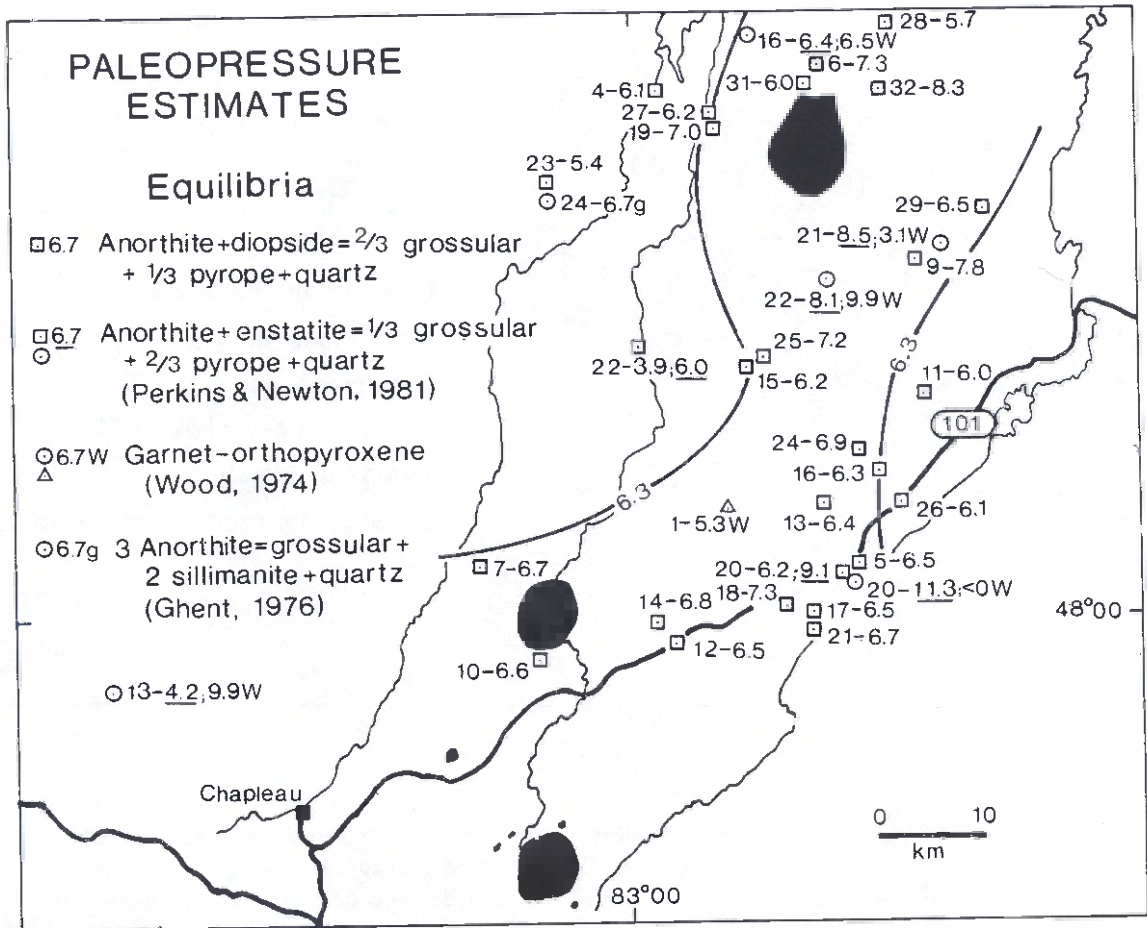


Figure 10. Paleopressure map of the Chapleau-Foley area. Symbols represent rock type (circles - paragneiss; squares - mafic gneiss; triangles - orthogneiss). Numbers to the right of the dash are pressure estimators (kbar) keyed to the equilibrium used to derive the value. The 6.3 kbar reference line is based on garnet-clinopyroxene-plagioclase-quartz equilibrium. (after Percival, 1983)

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

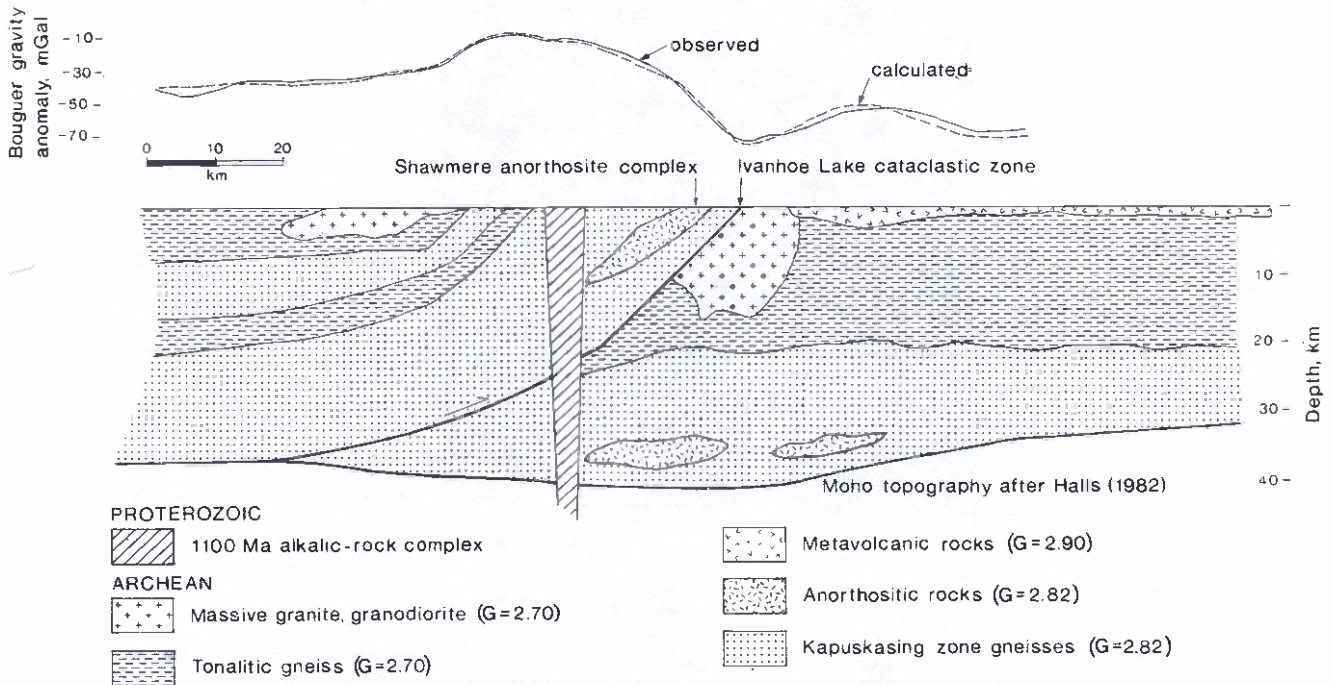


Figure 12. Generalized west-east cross-section from the Wawa domal gneiss terrane, through the Kapuskasing structural zone into the Abitibi subprovince, showing gross crustal structure. The gravity model based on the average rock densities: tonalitic gneiss and granite: 2.70; metavolcanics: 2.90; Kapuskasing structural zone and lower crust: 2.82 g/cm³.

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 Ivanhoe Lake cataclastic zone is the surface expression of a northwest-dipping thrust fault (Fig. 12). A short (10 km) seismic reflection survey over the zone indicates a reflector in the appropriate position with a true northwesterly dip of 33-40° (Cook, 1985; Fig. 13).

The Wawa-Kapuskasing boundary varies in character over its length. North of Bonar Lake, it is a fault, with distinct aeromagnetic expression, which diverges westward into Wawa tonalites toward the southwest. South of Bonar Lake, the boundary has

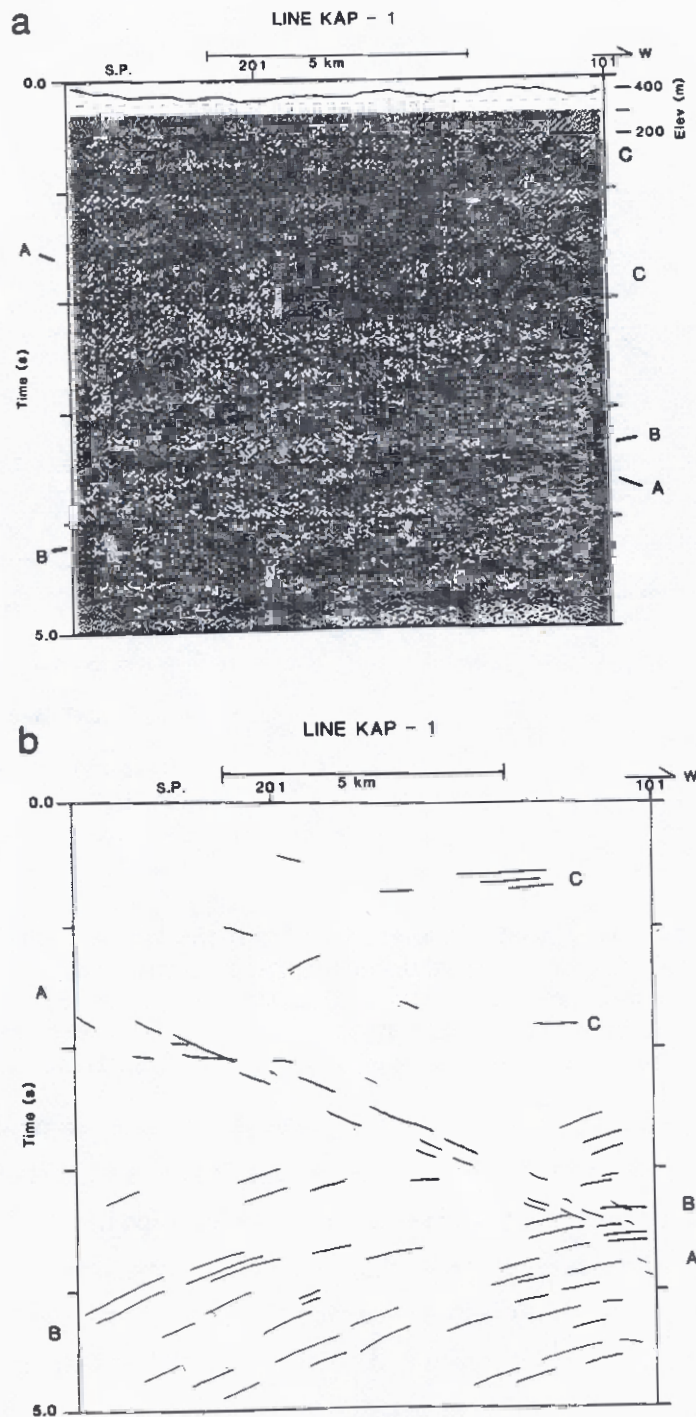


Figure 13. a) Seismic reflection profile over the eastern Kapuskasing zone. The data are plotted with east on left because the continuity of reflections is better in this orientation. Reflection A is likely from the Ivanhoe Lake cataclastic zone; reflections B are from within the Abitibi subprovince.

b) Line drawing interpretation of a) (after Cook, 1985).

gradational 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 between terranes with contrasting structural styles, 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 (Percival, 1981 a,b). Cataclastic veinlets characterize the faulted contact between mafic gneiss and tonalitic gneiss southwest of Kapuskasing Lake. To the south, 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 established prior to metamorphism and doming, and that rock units of the Kapuskasing zone locally occur structurally below the Wawa tonalite-granodiorite gneiss. Based on the change in average rock density across this diffuse subhorizontal boundary, Percival (1986) suggested that it could represent an exposed mid-crustal (Conrad) discontinuity.

STRUCTURE OF THE KAPUSKASING CRUSTAL CROSS-SECTION

The transition from the Michipicoten belt to the eastern boundary of the

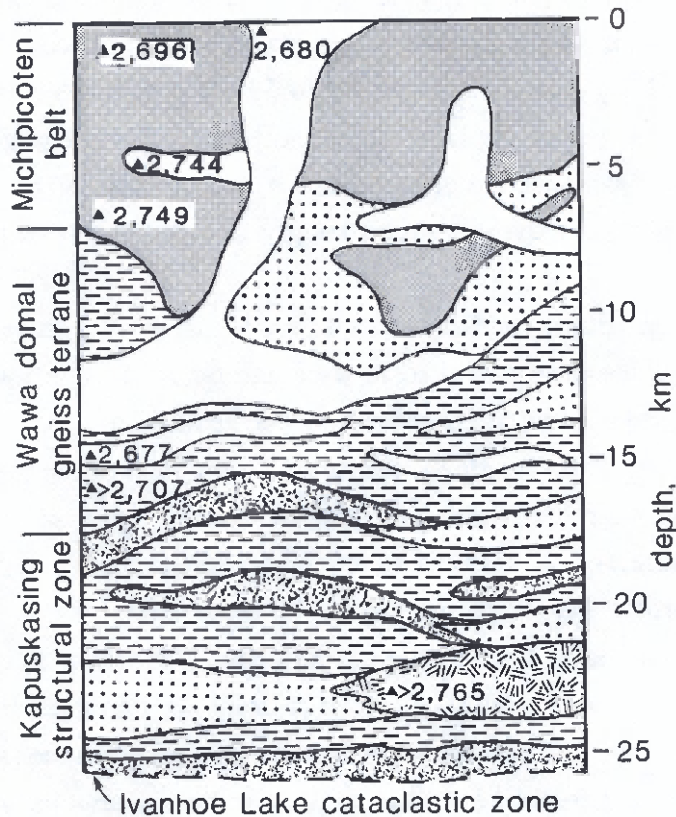


Figure 14. Restored vertical section through the Michipicoten belt, domal gneiss terrane and Kapuskasing zone. Numbers are zircon dates (± 4 Ma) on igneous and meta-igneous rocks (after Percival and Card, 1983).

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 (Studemeister, 1983) 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. 12); and 5) rocks with seismic velocities typical of the upper crust are not present

in the Kapuskasing zone. The three major terrane types recognized in the Abitibi-Wawa region can be related to depth zones in the crust based on metamorphic evidence and consistent with known seismic velocity characteristics of the crust of the Superior Province. Thus the uppermost crust is made up of supracrustal rocks of the greenstone belts and discordant plutonic rocks. Beneath is a megalayer made up of variably deformed felsic to intermediate plutonic rocks, with large-scale domal geometry. With increasing depth within this layer, the attitude of gneissosity changes from sub-vertical, near greenstone contacts, to sub-horizontal, near the Kapuskasing zone. The lowermost exposed megalayer is represented by the Kapuskasing zone, made up of a heterogeneous lithological assemblage at high metamorphic grade. Moderate dips of lithological layering are interpreted as the dominant sub-horizontal lower crustal attitude rotated passively during uplift.

Construction of a generalized crustal cross-section (Fig. 14) 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). Similar pressures, in the 2-3 kb range, are based on sphalerite-pyrrhotite geobarometry on rocks from Gutcher Lake, 30 km northwest of Wawa (Studemeister, 1983). 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 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 is reversed northwest of the Michipicoten belt where Ernst (1981, p. 87; 1983) reported consistent 85° SW dips of Matachewan dykes. Therefore, an intermediate dip value of 10° perpendicular to the fault was chosen for construction of Figure 14. 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 eastern Wawa subprovince. In view of the complex relationships described and uncertainties involved, these simplifications 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. 14), 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 and separated by an analogue of the Conrad discontinuity (Percival, 1986) is an estimated 10 to 15 km thickness 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.

Woods (1985) studied electrical conductivity in the Kapuskasing region with a large-scale magnetometer array. Although the lower crust is anomalously conductive in the area (Duncan et al., 1980), there is no conductivity anomaly associated with the Kapuskasing zone of mid- to lower crustal origin. This was interpreted to indicate that the conductivity anomaly at depth is the result of in situ fluids which were lost during uplift of the Kapuskasing structure.

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. 11) is given by a nearly concordant point and is therefore probably close to the age of crystallization. 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. The tonalitic intrusions have imposed amphibolite-facies aureoles on metavolcanic host rocks; considering the volume of tonalite, the heat from these magmas was probably sufficient to account for most of 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 (cf Wells, 1979). 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 of the main pile, and 2680 Ma, the approximate age of late- to post-tectonic plutons (Frarey and Krogh 1986). 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 relative tectonic quiescence 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 2696 to 2616 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

2696 to 2616 Ma ago, up to 60 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, an anomalously young zircon date of 2552 Ma on a trondhjemitic cobble from the weakly metamorphosed >2696 Ma Doré conglomerate near Wawa, reported by Turek et al. (1984), suggests that the date on the Borden Lake conglomerate cobble may not represent the age of crystallization of the source pluton. In addition, 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. It is more likely that a single protracted metamorphic event was responsible for producing the observed characteristics.

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 $700 \pm 50^\circ\text{C}$ 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 later fluid circulation and possible 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-plunging 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 promoted resetting of 2700-2680 Ma zircons to ages down to 2616 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-Opatika 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 2,650-2,250 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 $^{40}\text{Ar}/^{39}\text{Ar}$ analysis of blastomylonite from the Ivanhoe Lake cataclastic zone gave a date of 1,720 Ma (Percival, 1981; Fig. 15).

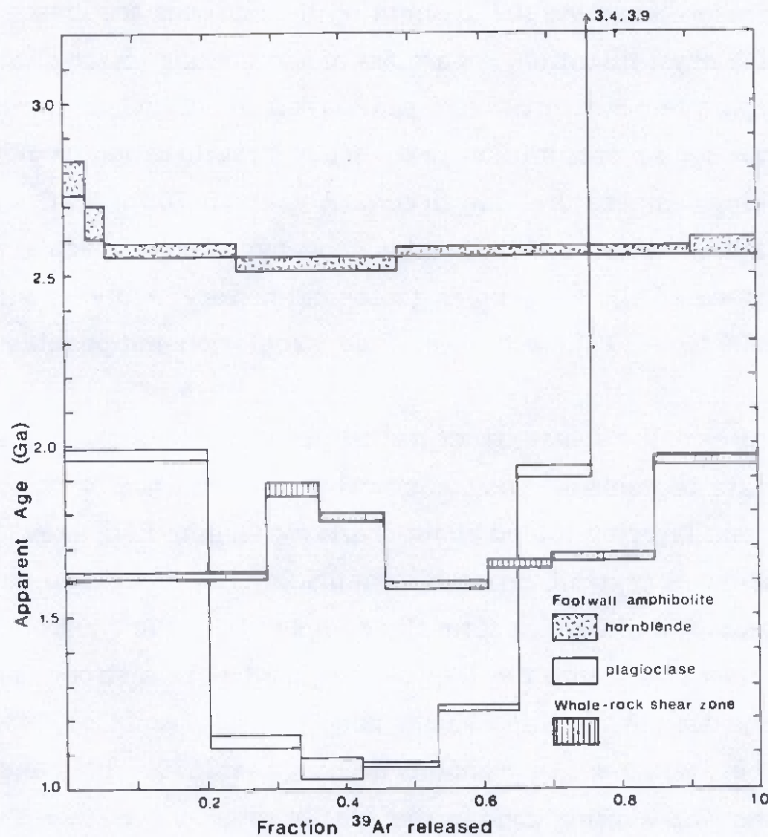


Figure 15. $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra for hornblende, plagioclase and whole-rock separates from near the Ivanhoe Lake cataclastic zone. Increasingly high release temperature from left to right. Width of bars are 2σ uncertainty estimates.

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 $^{40}\text{Ar}/^{39}\text{Ar}$ plateau at 1107 Ma (Fig. 15), 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.

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 compressional event, possibly an early Proterozoic collision in the Churchill Province to the northwest (Percival and McGrath, 1986).

PART II: ROAD LOG

THE KAPUSKASING UPLIFT: ARCHEAN GREENSTONES AND GRANULITES

SUMMARY

Various structural levels within the central Superior Province will be examined to demonstrate their characteristics and interrelationships. Starting with the lowest-grade rocks in the Michipicoten belt of the Wawa area, we will progress up-grade through an unbroken oblique crustal cross-section into tonalitic gneisses and granulites of the Kapuskasing zone.

DAY 1

Geology of the Wawa Subprovince, Wawa to Chapleau

The first day begins in well-preserved supracrustal rocks of the Michipicoten belt and ends in the lowest structural levels of the Wawa gneiss terrane (Fig. 2). Exposures near Wawa will demonstrate lithological, structural and metamorphic characteristics of the low-grade terrane. To the east, the transition to gneissic plutonic rocks and internal characteristics of the Wawa gneiss terrane will be examined.

STOP 1-1: Doré Conglomerate (E. and W. sides of Hwy 17)

The Doré is a thick, coarse polymictic conglomerate unit that is overlain by metavolcanic rocks of cycle III, dated at 2696 Ma (Turek et al., 1982) and underlain by older 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° eastward 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. A single trondhjemitic boulder yielded discordant zircons with an apparent age of 2552 Ma (Turek et al., 1984), approximately 150 Ma younger than the stratigraphic age of the unit based on zircon dates of bracketing volcanic rocks. It is likely that the source of the trondhjemitic boulders and cobbles is syn-volcanic plutons of 2744-2735 Ma age.

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 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).

The iron formation caps the first volcanic cycle in the Michipicoten greenstone belt and from the stratigraphic bottom to top consists of five lithologic rock types that are gradational into each other. Upper and lower contacts 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 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 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, in less abundant quantities.

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 Spherulitic rhyolite (100-200 m south of gate on McLeod Mine road)

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 well developed 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.

Return to Hwy 101

00.0 km - Junction of Highway 101E and Broadway Ave, Wawa. Proceed east on Highway 101.

20.9 km - Stop 1-4: Mafic gneiss - tonalite contact zone (N. and S. sides of Hwy 101)

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 hornblende-plagioclase gneiss is considered to represent deformed, metamorphosed Michipicoten volcanics and metagabbro. In this exposure 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.

45.1 km Jct. Hwy. 651. Continue east.

68.5 km Stop 1-5: Xenolithic tonalite gneiss at Budd Lake (N. and S. sides of Hwy 101)

This complex outcrop consists of several phases: 1) xenoliths of mafic gneiss, interpreted as rafts of the Michipicoten metavolcanic sequence, 2) gneissic tonalite with small, wispy mafic xenoliths, cut by 3) foliated to gneissic hornblende-epidote-biotite-sphene tonalite, all cut by 4) pink-white granitic pegmatite. Hornblende from foliated tonalite (3) contains 10.4 wt% Al_2O_3 , indicating crystallization at approximately 5.6 kbar (Hammarstrom and Zen, 1985).

Continue east on 101

94.4 km Stop 1-6: Tonalite gneiss and mafic dykes (N. and S. of Hwy 101)

Tonalite gneiss is cut by northwest- and northeast-trending mafic dykes with chilled margins. The older northwest-striking Hearst dykes (Ernst and Halls, 1980, 1984) 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 ± 75 Ma (Gates and Hurley, 1973). Ernst and Halls (1980, 1984) 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, 1983; Percival, 1981). The tonalitic gneiss is thinly layered and has sparse mafic xenoliths. Gneissosity appears to have chaotic orientation but is subhorizontal on average. There is evidence for at least two sets of structure: an older gneissosity is reoriented by younger subhorizontal foliation to give complex sigmoidal patterns.

Continue east

122.2 km Stop 1-7: Highbrush Lake Dome (N. side of Hwy 101)

Small domes exposed in roadcuts here are typical of the local and possibly 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 with a rolling east-west mineral and stretching lineation. 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. Late sub-horizontal high-strain zones are common in the vicinity of the transition zone between Wawa and Kapuskasing gneisses.

Continue east

131.6 km - Junction of Highway 101 and Highway 129. Follow 101E, 129N toward Chapleau

132.9 km Stop 1-8: Xenolithic tonalite with horizontal shear zones (W. side of Hwy 101-129)

Xenolith alignment and gneissosity have sub-vertical orientations in the central part of this road-cut. Layering becomes horizontal and mafic xenoliths become thin mafic layers in 1/2 m-thick horizontal shear zones near the top and bottom of the outcrop.

139.5 km - Junction of Highways 101E and 129N: 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.

178.3 km - Stop 1-9: 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.

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.

179.3 - Foliated and lineated diorite with gabbro and melagabbro layers

Igneous clinopyroxene with hornblende overgrowths and granitic leucosome layers are present.

183.2 km - Stop 1-10: Granulite gneiss, Robson Lake dome

The easternmost exposure of the igneous complex is a foliated to augen-textured, xenolithic hornblende porphyritic granite with leucogranite veinlets parallel to foliation on the 1 mm-2 cm scale. The layering is folded about northwesterly-plunging axes. Parts of the outcrop are 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 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.

Return to Chapleau

DAY 2

Geology of the Kapuskasing Structural Zone in the Chapleau-Foleyet Area

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.

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

13.0 km Stop 2-1: Borden Lake conglomerate (S. side of Hwy 101)

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 hornblende 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.

An aggregate of tonalitic cobbles extracted from the metaconglomerate 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. Rather than recording the original crystallization age of the source pluton for the cobbles the zircons probably date a later deformation-metamorphic event. The source pluton for the cobbles may be similar to that which provided material for the Doré conglomerate.

Continue east on 101

25.9 km Stop 2-2: Mafic gneiss xenoliths with amphibolitic margins (N. side of Hwy 101)

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, the dominant rock type 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 magma 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 as low as 2616 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 at least 2616 Ma ago.

Continue east

37.0 km Stop 2-3: Thinly-layered tonalitic gneiss and diatreme breccia (N. side of Hwy 101)

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 ~1100 Ma age (Stevens et al, 1982) 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. These include tonalitic gneiss, spinel hercynite and 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 model, lies vertically below at a depth of ~15 km.

Continue east on Hwy 101

52.1 km Stop 2-4: Kapuskasing Gneiss (S. side of Hwy 101)

Layered mafic gneiss with concordant in situ tonalitic leucosome, and cross-cutting tonalitic and pegmatitic dykes. Layering on the 5 to 10 cm scale is given by alternating hornblende-rich and garnet-pyroxene-rich layers (see analyses of similar layers in Table 1). Metre-scale blocks of mafic gneiss in breccia give parts of the outcrop a chaotic appearance. These structurally complex panels are separated by m-scale high-strain zones with gently rolling, north-dipping, pronounced foliation.

Continue east on Hwy 101

59.8 km Stop 2-5: Kapuskasing gneisses (N. and S. sides of Hwy 101)

There are several features of interest at this outcrop (Fig. 23):

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 identified by microprobe analyses, form barely-visible coronas around garnet, clinopyroxene and hornblende. Analyses of the symplectite minerals, at

the lower size limit of microprobe resolution, are reported along with those of the other minerals in the rock, in Table 2. 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.

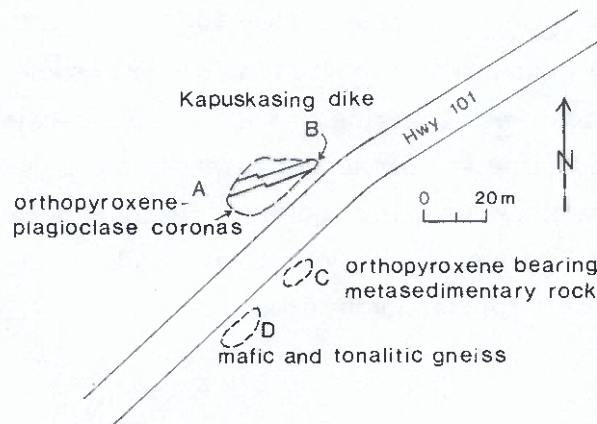


Figure 23. Location of outcrops at Stop 2-5.

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., 1982). Hanes et al, (1986) estimated an age of 2040 - 2200 Ma based on $^{40}\text{Ar}/^{39}\text{Ar}$ analyses of a Kapuskasing dyke and its baked country rock.

C. Homogeneous metasedimentary rock

South of the road is a flat outcrop of medium grained rock with the assemblage

Table 2. Microprobe analyses of minerals in coronitic mafic gneiss, Stop 2-5

	1	2	3	4
SiO ₂	38.01	51.57	49.06	42.29
TiO ₂	0.00	0.34	0.03	2.03
Al ₂ O ₃	20.99	2.92	4.75	12.98
Cr ₂ O 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 ₂ O	0.27	0.74	0.52	1.95
K ₂ O	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
Aliv	0.00	0.092	0.115	1.748
Alvi	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)

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

* Total iron as FeO; Fe³⁺ by stoichiometry
Specimen also contains quartz and plagioclase (An_{33,57} in matrix,
An₈₈ in symplectite)

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. Application of the garnet-orthopyroxene-plagioclase-quartz geobarometer yields values in excess of 11 kbar by both Newton and Perkins (1982) and Perkins and Chipera (1985) calibrations.

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. Thin mylonite seams are sub-parallel to gneissic layers.

Continue east

67.0 km Stop 2-6: Xenolithic tonalitic gneiss (S. side of Hwy 101)

This outcrop consists of medium- to coarse-grained tonalite made up of garnet, hornblende, biotite, plagioclase and quartz. A variety of xenoliths includes mafic gneiss (garnet-clinopyroxene-plagioclase-quartz), amphibolite, biotite-rich schists and spinel pyroxenite. Amphibole-rich rims characterize the high-grade inclusions.

Continue east

69.2 km Stop 2-7: Ivanhoe Lake cataclastic zone (S. side of Hwy 101)

The outcrop 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 $^{40}\text{Ar}/^{39}\text{Ar}$ whole-rock analysis of material from this vein yielded an age plateau at 1720 Ma (Fig. 15). On the west side of the outcrop are thin (3 cm) rusty-weathering lamprophyre dykes. Tonalitic rocks cut by cataclasite are exposed 150 m north of the highway on an overgrown road.

Continue east

70.5 km Stop 2-8: Mafic metavolcanics, Abitibi subprovince (N. side of Hwy 101)

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 $^{40}\text{Ar}/^{39}\text{Ar}$ method (Fig. 15) 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.

72.3 km Turn north on logging road follow main road.

87.2 km Stop 2-9: Shawmere gabbroic anorthosite (W. side of road)

The outcrop is mainly coarse-grained gabbroic anorthosite with hornblende and rare garnet as mafic minerals. Ultramafic layers consisting of ortho- and clinopyroxene with hornblende rims and sparse plagioclase to 3 cm, occurs in layers and pods up to 4 m thick. Mafic and ultramafic layers are locally folded into shallow NE-plunging structures with a prominent lineation. One part of the outcrop is a spectacular coronitic gabbroic anorthosite with football-sized plagioclase megacrysts. Coronas have orthopyroxene or clinopyroxene cores; orthopyroxene has successive hornblende and pale garnet rims; clinopyroxene has hornblende rims and rare orthopyroxene cores.

Return to Hwy 101.

134.1 km END OF DAY 2. Return to Wawa.

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