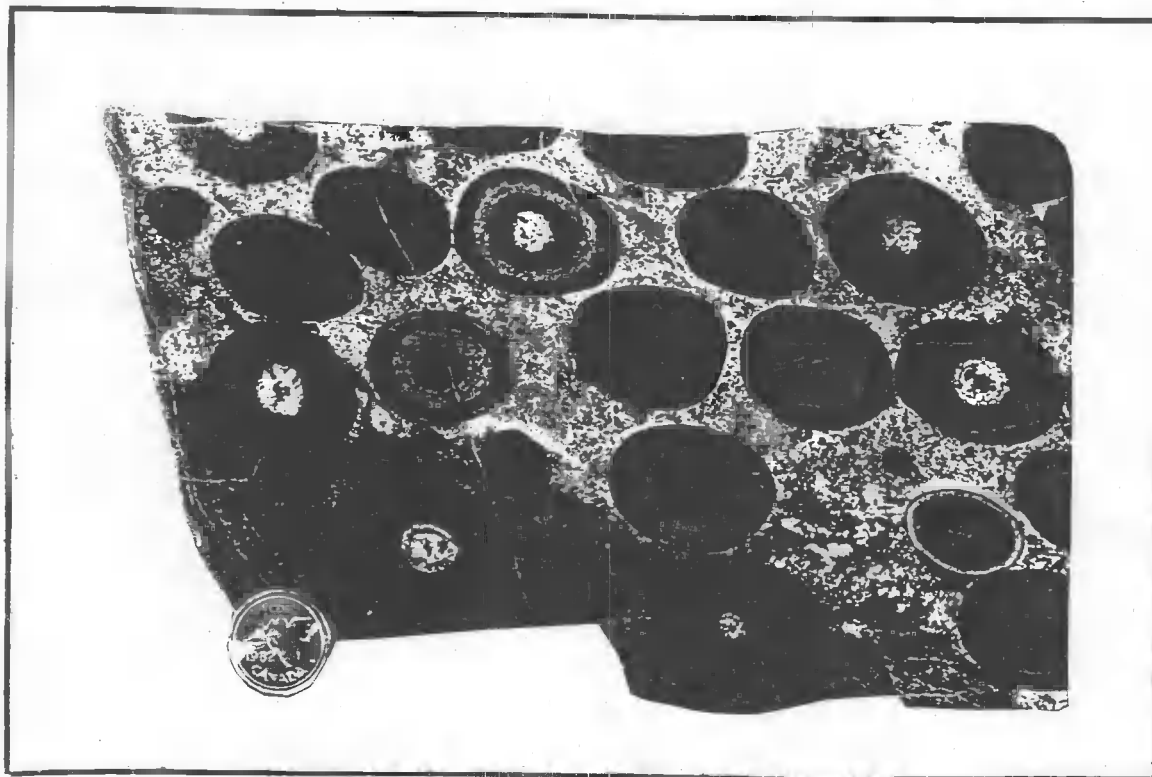
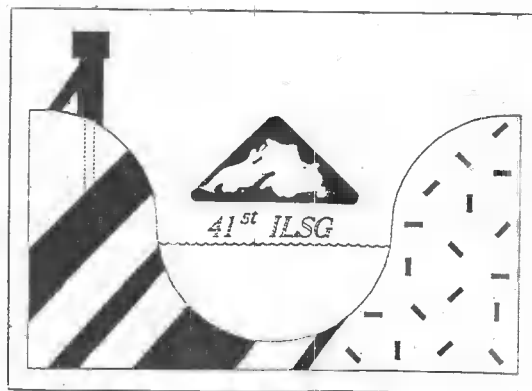

Alkalic Rocks of the Midcontinent Rift



**Institute on Lake Superior Geology
41st Annual Meeting, May 13-18, 1995
Marathon, Ontario**

**Proceedings Volume 41: Part 2a
Field Trip Guidebook**



Alkalic Rocks of the Midcontinent Rift

by

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Frontispiece: Orbicular ijolite from the Prairie Lake carbonatite, the world's only known occurrence of this texture in a rock of this composition (photograph courtesy of John Scott)

CHAIRMAN'S NOTE:

Due to sudden illness, co-author and field trip co-leader Dr. David Watkinson could not take part in these field trips that feature the alkalic rocks associated with the Midcontinent rift in the vicinity of Marathon. Dr. Watkinson's continuing research into the petrology and metallogeny of these rocks dates back twenty-five years. He has expressed his regrets about not being able to attend but has wished all meeting and trip participants the best of luck during the 41st I.L.S.G.. I'm sure that you will join me in extending our best wishes for a speedy and complete recovery to Dr. Watkinson.

In the absence of Dr. Watkinson, Dr. Cliff Shaw (Post-Doctorate Fellow, University of Western Ontario) has graciously offered to co-lead these field trips with Dr. Ron Sage. Dr. Shaw has just completed his Ph.D. thesis on the gabbros of the Coldwell complex and has taken part in the recent 1:20 000 scale mapping project of the complex. This knowledge will definitely be an asset during these trips. On behalf of the I.L.S.G. organizing committee, I would like to express my gratitude to Dr. Shaw for stepping in on such short notice and wish all of you a great field trip!

*Mark Smyk
Chairman, 41st I.L.S.G.
April 23, 1995*

ACKNOWLEDGEMENTS

The authors were helped in preparation of trails and sites by Mr. Duncan Michano of Heron Bay, Ontario. Permission to examine outcrops of the Dead Horse Creek Diatreme was granted by Mr. John Ternowesky, Prospector, Thunder Bay, Ontario. Mr. J. McGoran, President, Fleck Resources Ltd., Vancouver, agreed to display diamond drill core and samples from the Two Duck Lake intrusion within the Port Coldwell Complex.

The Ontario Geological Survey donated copies of Geological Studies 27, 45 and 46 and Geological Report 264 to the Institute of Lake Superior Geology in support of the alkalic rock field trip. These reports provide the background data and more detailed maps for the Slate Islands, Killala Lake Complex, Prairie Lake Carbonatite and Diatreme structures found in the region.

VIEW LOOKING EAST: Carbonatite intrusion from the Prairie Lake
Complex. The intrusion is a large, irregularly shaped body of
intrusive material, composed primarily of
silica-rich, crystalline material.

ALKALIC ROCKS OF THE MIDCONTINENT RIFT

Introduction

Alkalic rock intrusions occur in an area that extends from the north shore of Lake Superior in a north to east of north direction for approximately 140 km. The diatreme structures found on the Slate Islands (Sage, 1991) and the carbonatite dikes with associated fenite in the Geraldton area (Sage, 1985) will be excluded from the present tour due to difficulty of access (Slate Islands) or distance (Chipman Lake) (Figure 1). An attempt will be made to examine the major rock types in the alkalic rock, carbonatite and diatreme structures near Marathon, Ontario. By necessity the visits to individual outcrops will be brief. Whenever possible, economic aspects of the alkalic rocks will be emphasized. Some optional stops are included in the guidebook for those who wish to spend additional time on their own in the area examining exposures which time does not permit this tour to visit.

Regional setting

The Midcontinent Rift is represented by a horseshoe shaped gravity and aeromagnetic anomaly open to the south that is exposed in the Lake Superior region (Figure 2) and lies east of the Kapuskasing Structural Zone (KSZ) and northwest of the Grenville Front Tectonic Zone (GFTZ) (Figure 3). The western limb of this gravity and aeromagnetic anomaly strikes southwest into Kansas and it has been postulated by Adams and Keller (1994) to extend as far south as west Texas and eastern New Mexico. The eastern arm of the Midcontinent Rift strikes southeast into southern Michigan and has been extended into Ohio and Kentucky by Dickas et al. (1992). The rift has been interpreted to be a triple junction formed by a rising mantle plume (Burke and Dewey (1973); however, Green (1983) concluded that a triple junction has not developed in the usual sense due to the U-shape of the rift and the abrupt termination of activity along it. The model of decompression melting over an upwelling mantle plume in an extensional lithosphere environment is the favored model at present (Nicholson and Shirley, 1990; Hutchinson et al. 1990). Nicholson and Shirley (1990) indicated that large volumes of magma of dominantly tholeiitic composition extruded over a short period is consistent with the mantle plume model. Tre'hu et al. (1991) stated that the Midcontinent Rift resembles a passive continental margin which is in marked contrast to many models of active and extinct Phanerozoic rift zones. Behrendt et al. (1988) have also noted that the Midcontinent Rift is different from

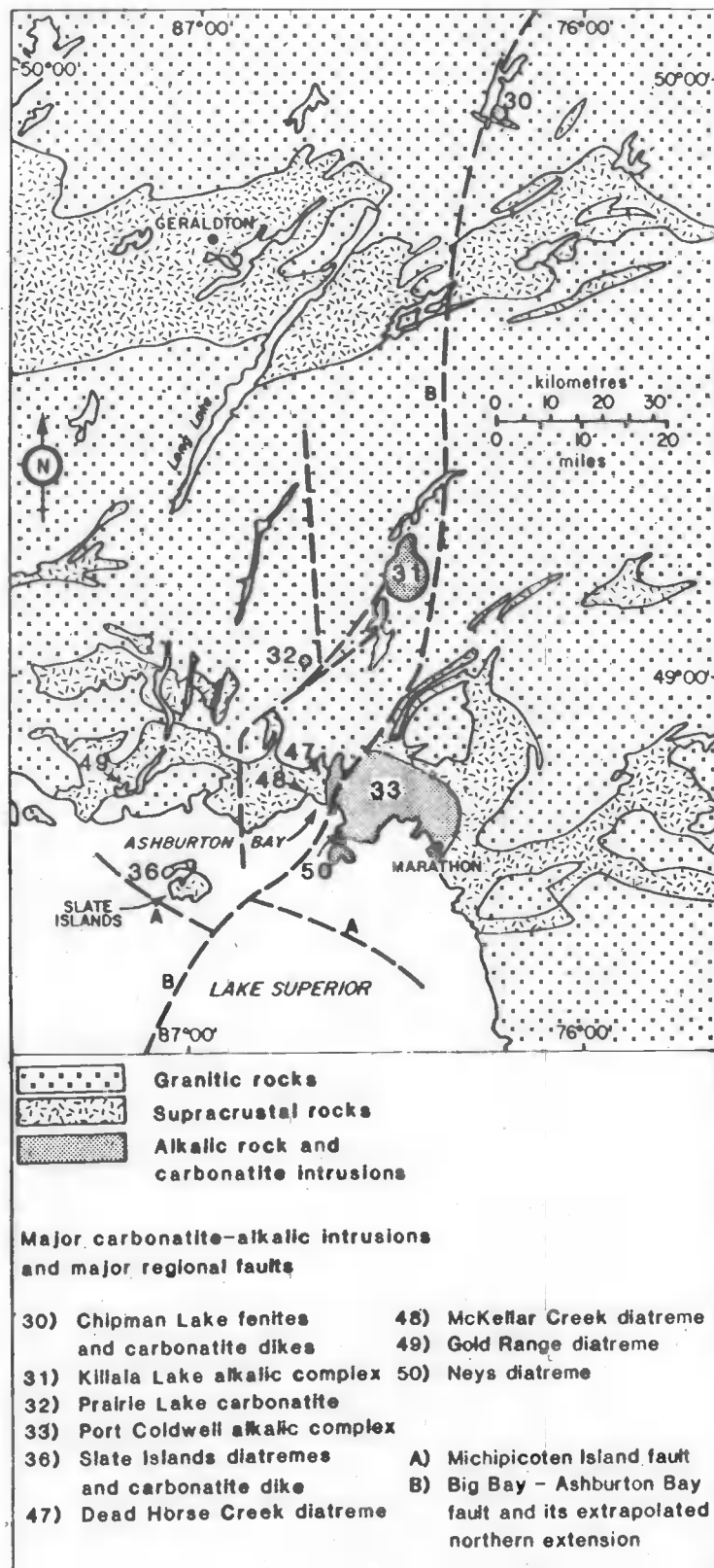


Figure 1: Index map showing the location of features associated with Keweenaw alkaline magmatism (from Sage, 1986).



Figure 2: Index map showing general location of the Midcontinent Rift System (Shaded Area). Shaded area includes that part of the gravity high associated with the rift that is greater than -20 milligals; the portion extending into Kansas is drawn along the -40 milligal contour. The gravity high roughly, but not uniquely, defines location of dense crust associated with the Midcontinent Rift System. The rift extends beneath Lake Superior, although the gravity high is not continuous in this area because of structural complexities. Modified from Klasner et al. (1982).

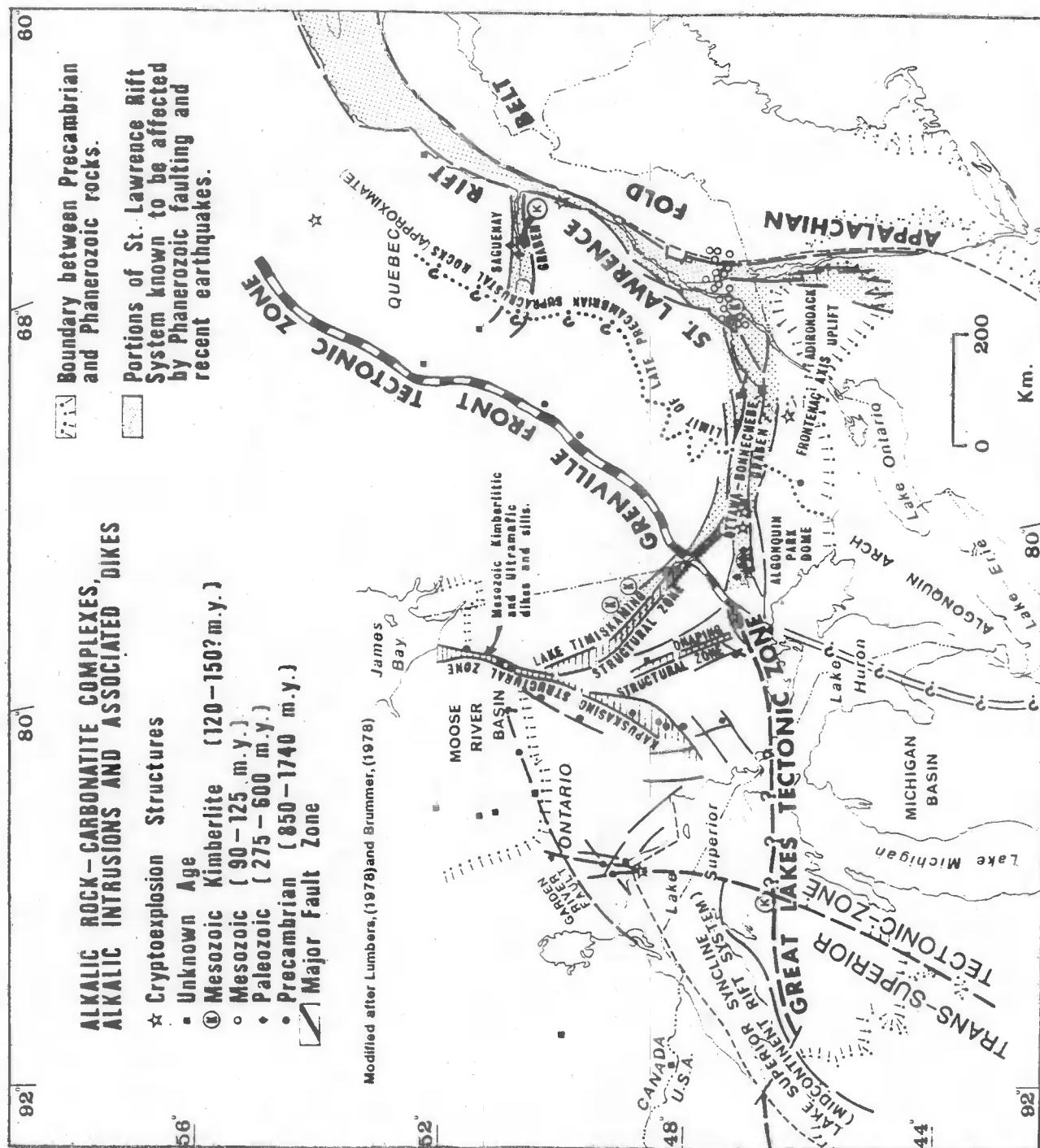


Figure 3: Index map showing relationship of major regional structural features to the Midcontinent Rift System (Modified from Lumbers, 1978; Brummer, 1978; from Sage, 1986).

Phanerozoic rifts in that it has a crustal thickness equal to or greater than the surrounding unextended terrane.

A high-velocity lower crustal layer is thickest beneath the rift graben under Lake Superior and is interpreted to be magmatic material underplated during rifting (Behrendt et al., 1990; Tre'hu et al., 1991; Shay and Tre'hu, 1993). Gravity modeling requires large buried masses of mafic igneous rocks of assumed Keweenawan age beneath the Midcontinent Rift whose distribution is asymmetric (Mariano and Hinze, 1994b; Thomas and Teskey (1994)). This asymmetry implies that the northern margin of the rift is dominated by plutonic rocks and the southern margin by volcanic rocks (Thomas and Teskey, 1994). The Port Coldwell Complex and Duluth Complex are manifestations of the dominant plutonic northern margin. Allen et al. (1992) suggested that a topographic dome centered on Lake Superior represents the effect of the mantle plume active in the Keweenawan time of rifting which is now maintained isostatically by changes to the lithosphere. These changes consist of magmatic underplating and depletion of the upper mantle. Paces and Miller (1993) compiled existing U-Pb geochronological and paleomagnetic data for volcanic and intrusive rocks within the Lake Superior Basin. A simplified version of their data is given in Table 1 which demonstrates that the Duluth Complex and Port Coldwell Complex have similar U-Pb isotopic ages and were emplaced just before and after a paleomagnetic reversal; from reverse polarity (Port Coldwell) to normal polarity (Duluth Complex). Samson and West (1994) reported that this reversal occurred at approximately 1097 Ma. During this polar reversal the composition of the volcanic rocks changes from those displaying crustal contamination (Osler series) to those displaying typically uncontaminated asthenospheric mantle-derived melts (Mamainse Point) (Lightfoot et al. 1994). Both volcanic series evolved from high-Mg primitive basal members to those of tholeiitic composition; the earlier flows (Osler series) may represent deeper partial melts than the later series (Mamainse Point; Lightfoot et al. (1994)).

TABLE 1
(Simplified from Paces and Miller, 1993)

Rock Unit	Polarity	Age (Ma)	Reference
Unnamed	R	1097.5 $\pm$ 3	1
Copper City Flow	N	1096.2 $\pm$ 1.8	2
Greenstone flow	N	1094.0 $\pm$ 1.5	2
Lake Shore Trap	N	1087.2 $\pm$ 1.6	2
Nathan's Layered Series	R	1106.9 $\pm$ 0.6	3
Duluth Complex	N	1099.0 $\pm$ 0.5	3
Beaver Bay Complex	N	1096.0 $\pm$ 1	3

Logan Sills	R	1108.8 +4 -2	4
Osler Porphyry	R	1107.5 +4 -2	4
Agate point Rhyolite	R	1097.6 +- 3.7	4
Port Coldwell	R	1108 +- 1	5
Complex			
Michipicoten Island	N	1086.5 +13 -3.0	6
Formation			

1) Van Schmus et al.(1990); 2) Davis and Paces(1990); 3) Paces and Miller(1993); 4) Davis and Sutcliffe(1985); 5) Headman and Machado(1992); 6) Palmer and Davis(1986)

In the Lake Superior region separation along the rift axis was on the order of 50 to 60 km(Klasner et al. 1982; Chandler, 1983; Shay and Tre'hu, 1993) and the resulting graben filled with 30 to 32 km of volcanic and sedimentary rocks(Tre'hu et al., 1991; Behrendt et al., 1988; Cannon et al. 1989). Gravity modeling is consistent with the Keweenaw, Isle Royal, Thiel(Big Bay-Ashburton Bay), Douglas and Michipicoten Island Faults being growth faults controlling rift development(Thomas and Teskey, 1994)(Figure 4). Cannon et al.(1989) interpreted regional seismic data to suggest that the central graben beneath Lake Superior changes attitude along strike and thus the rift is segmented and requires accommodation zones between the segments. Sexton and Henson(1994), using an interpretation of seismic data, questioned the modeling of accommodation zones and suggested that upon closer inspection zones of accommodation may not be required to account for changing attitudes along strike.

From seismic data, the depth to the Moho beneath western Lake Superior is 37-46 km, central Lake Superior approximately 55 km and in eastern Lake Superior 42-49 km(Behrendt et al. (1988). Tre'hu et al. (1991) and Shay and Tre'hu(1993) used seismic data to interpret the crustal thickness beneath the Midcontinent Rift as diminishing rapidly to the south from 55 to 60 km at the centre of the rift to 35 km and gradually to the north to 40 km.

Bisecting the gravity anomaly of the horseshoe-shaped Midcontinent Rift is the Trans-Superior Tectonic Zone that contains the Thiel or Big Bay-Ashburton Bay Fault(Sage, 1978, Klasner et al., 1982)(Figures 3, 5, 6, and 7). This zone of deformation is represented by changes in isomagnetic and isomilligal contouring of aeromagnetic and gravity data(Figures 6 and 7) and is also represented as a topographic high on lake bottom contours(Hough, 1958, Wold et al. 1982)(Figure 8). The Slate Island Diatremes and the Port Coldwell Complex occur where this zone impacts on the north shore of Lake Superior and is proximal to the Proterozoic-Archean contact. Extrapolating faults related to this zone to the region of Chipman Lake(Sage, 1985), a distance of approximately 140 km, reveals a change in the

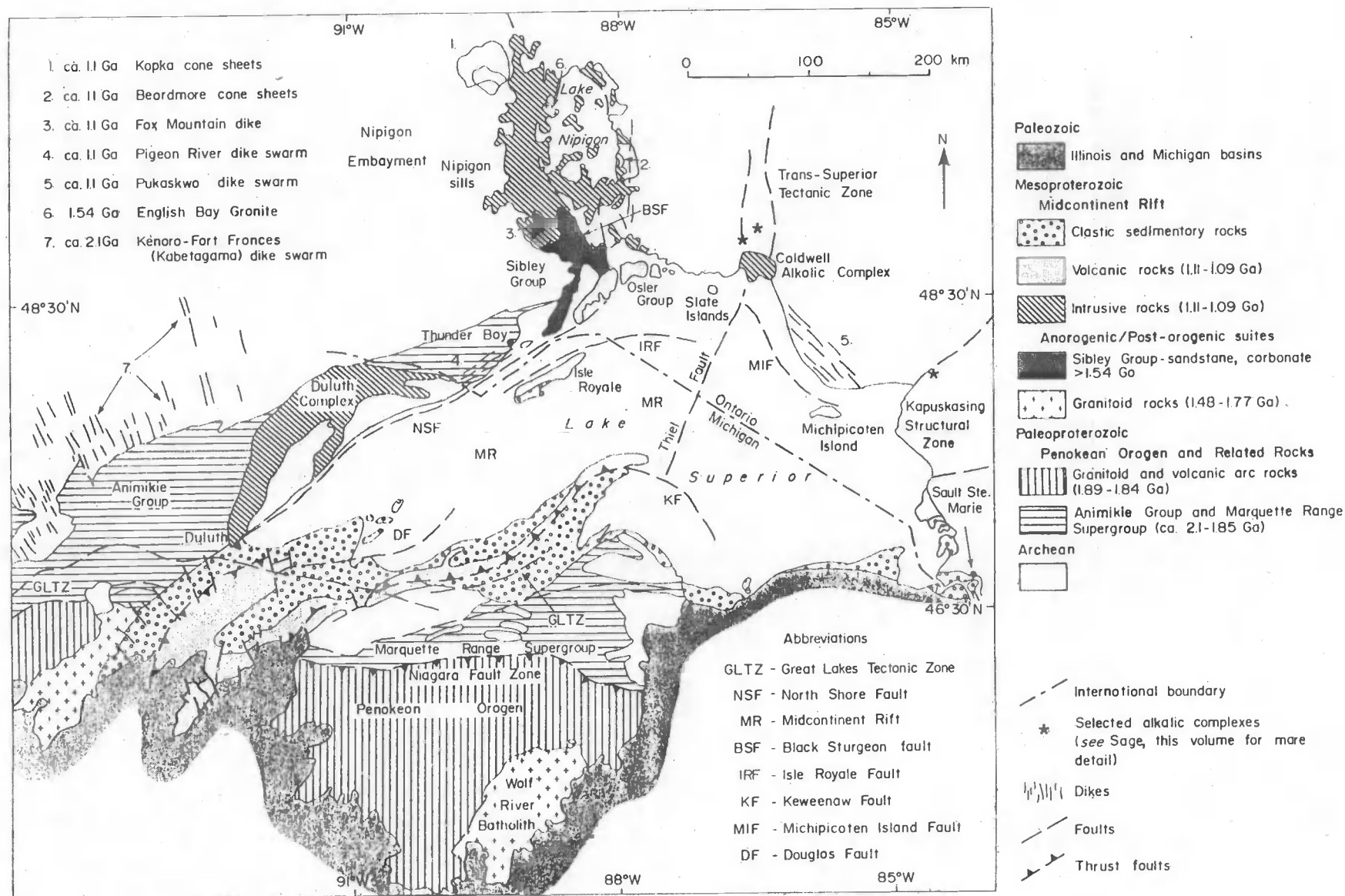


Figure 4: Major Proterozoic geological features found in the Lake superior region(from Sutcliffe, 1991).

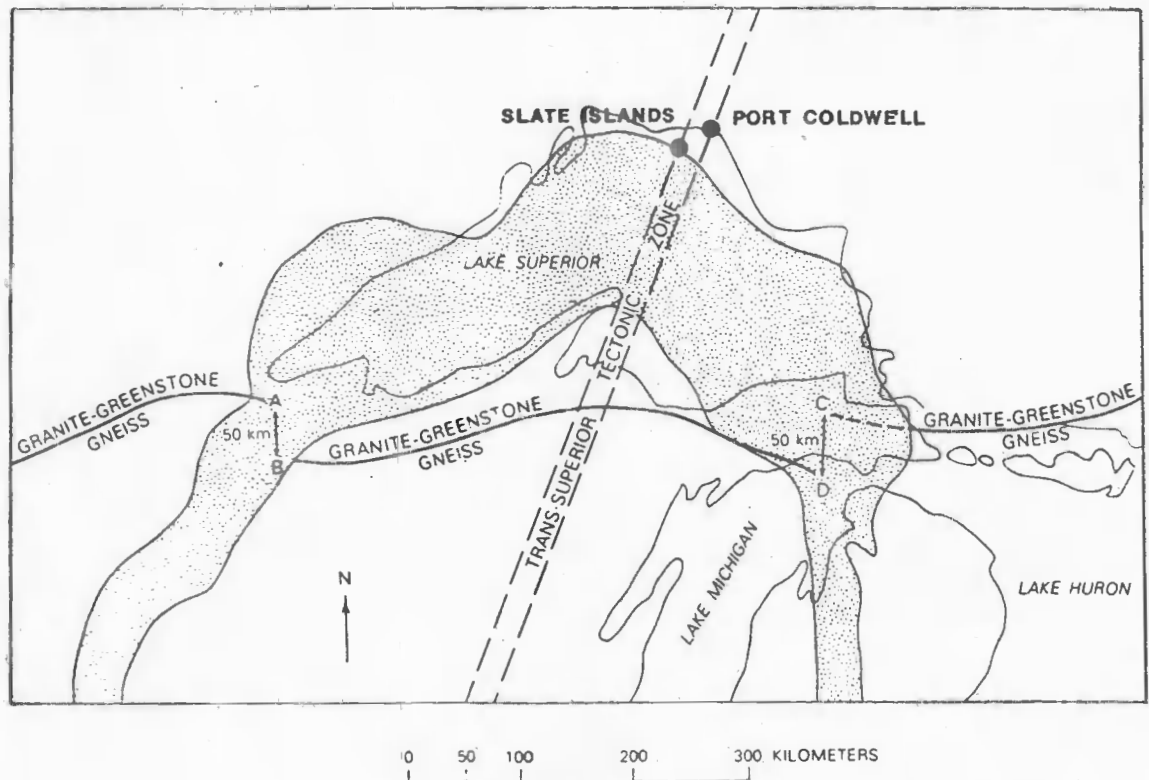


Figure 5: Diagram indicating the amount of separation that may have occurred during development of the Midcontinent Rift System, location of the Trans-Superior Tectonic Zone and the position of two alkaline features associated with the tectonic zone(modified from Klasner et al., 1982).

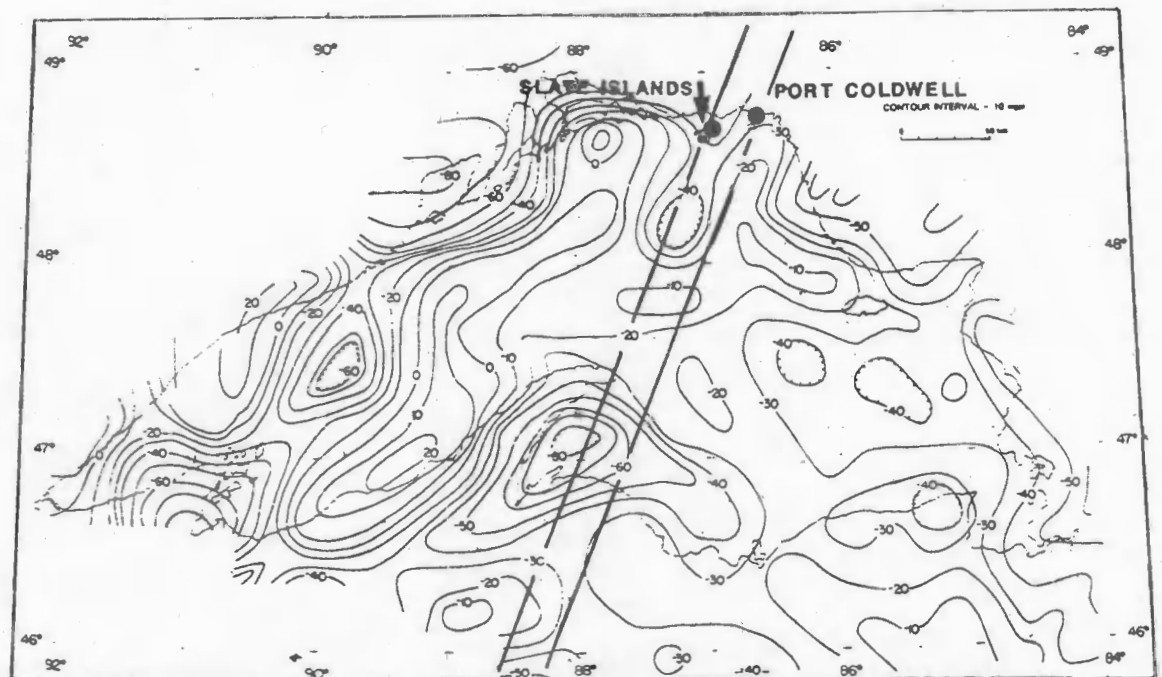


Figure 6: Regional Bouguer gravity anomaly map over Lake Superior with the Trans-Superior Tectonic Zone and two sites of alkaline magmatism superimposed(modified from Hinze et al., 1982).

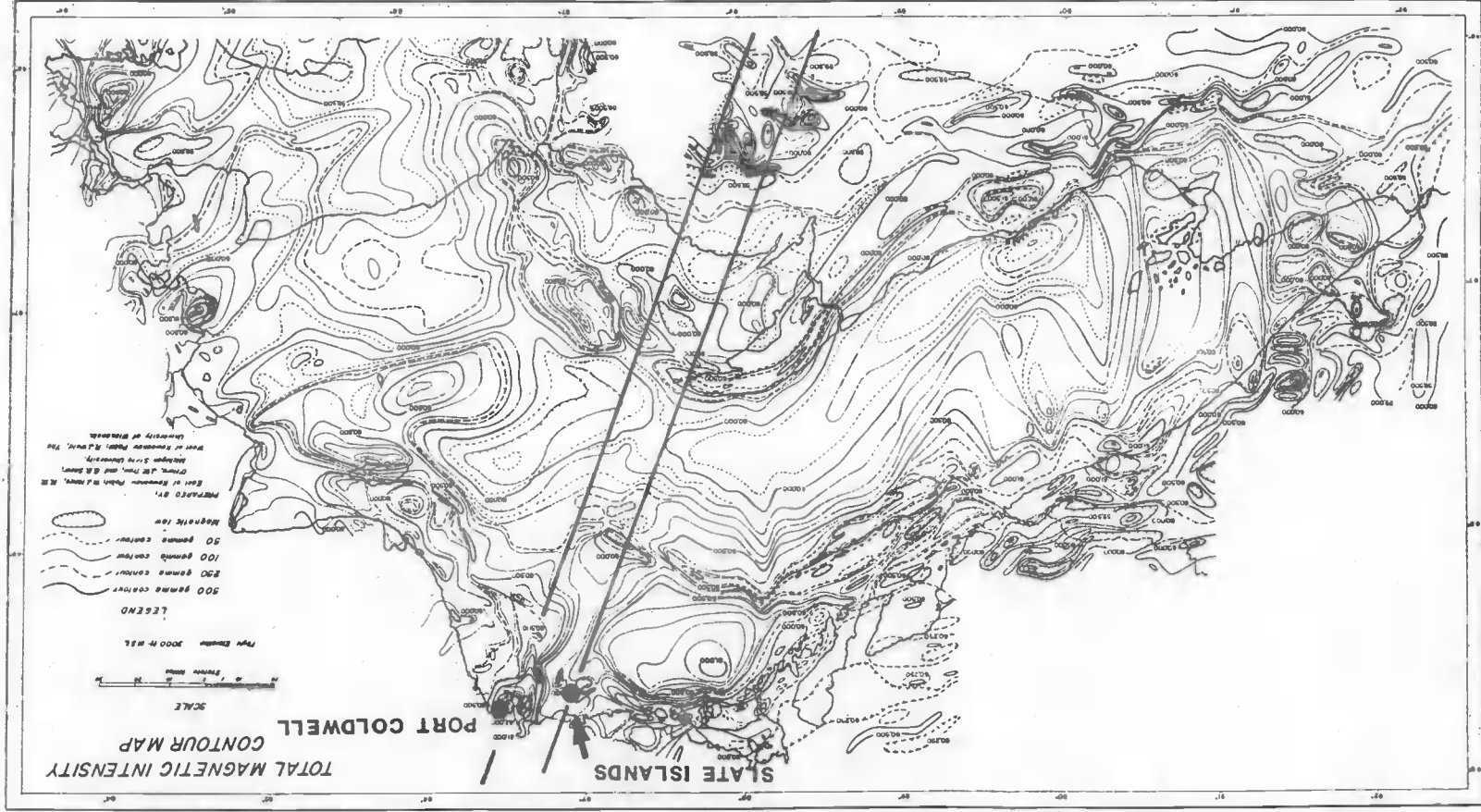


Figure 7: Total magnetic intensity map of Lake Superior region with the Trans-Superior Tectonic Zone and two sites of alkaline magmatism superimposed(modified from Hinz et al., 1966).

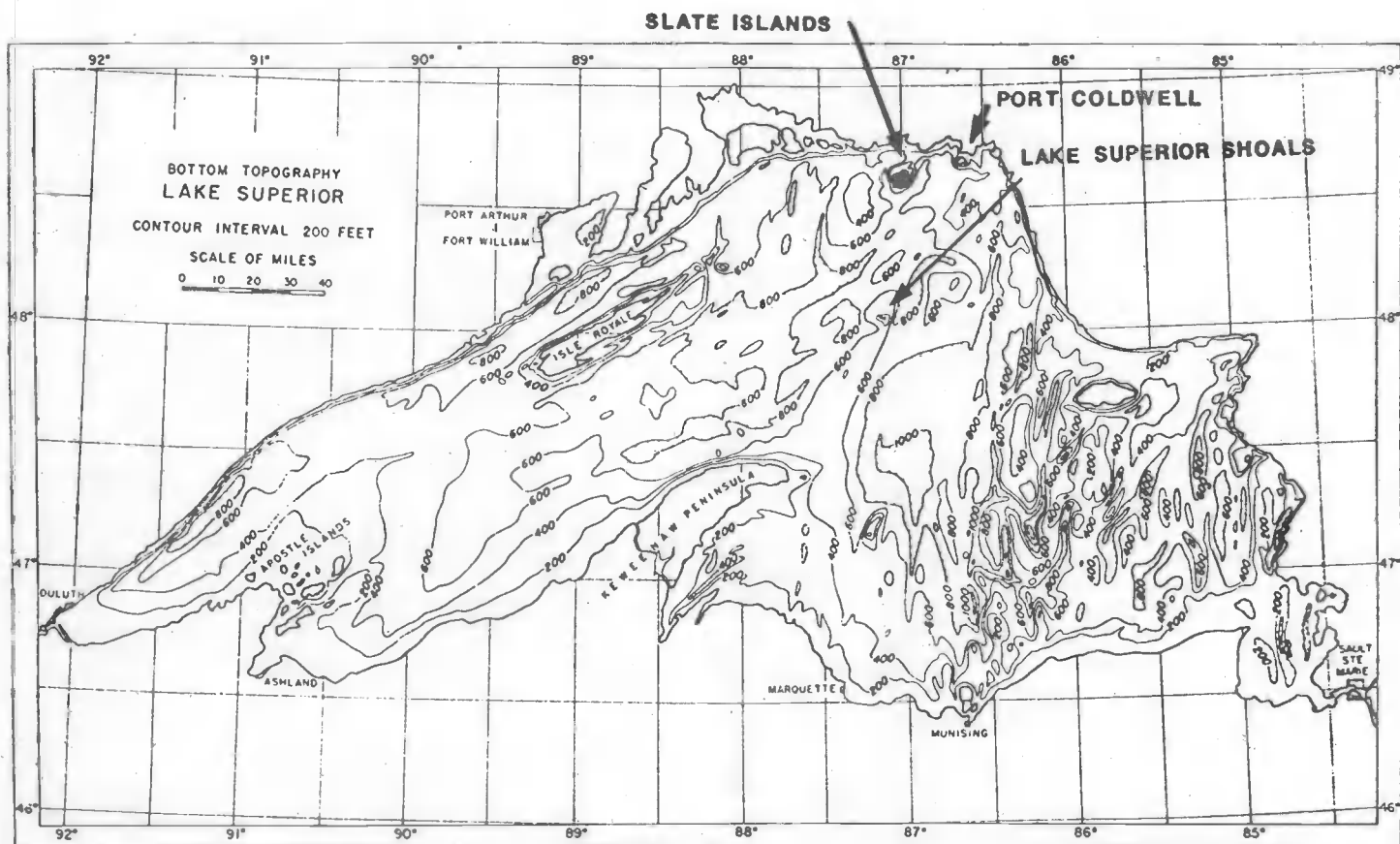


Figure 8: Contour map of lake bottom topography of Lake Superior (From Hough, 1958; Wold et al., 1982).

scale of alkalic magmatic activity. The Port Coldwell complex consists of 3 superimposed ring complexes exposed along the shore of Lake Superior; the smaller Killala Lake Complex, 40 km north of the lake, consists of a single ring complex. The much smaller Prairie Lake Carbonatite is 30 km north of the lake and the Chipman Lake alkalic units, at a distance of 140 km, consist of carbonatite dikes and fenite. What structural or tectonic significance this change from alkalic rock to carbonatite magmatism, and presumably much lower degrees of partial melting at greater depths, may have is unclear.

Interpretation of the aeromagnetic data in eastern Lake Superior by Hinze et al. (1966) indicates that the Thiel or Big Bay-Ashburton Bay fault offsets the Michipicoten Island fault left-laterally for approximately 5.0 miles (8.0 km). This left-lateral offset has been recognized along the northern extension of this structure in the Chipman Lake area where offset of approximately 0.8 km has been mapped (Sage, 1985). The age relationships between the younger Thiel or Big Bay-Ashburton Bay fault and the older Michipicoten Island fault are consistent with recent structural interpretations using seismic data (Mariano and Hinze, 1994).

Linear trends in the topography approximately 1.5 km southeast and southwest of the Slate Islands estimated to be approaching 240 metres in depth, are consistent with the fault trends established by Hinze et al. (1966) and Sage (1991). These trends likely represent subsidiary faulting related to the two regional fault trends. The Michipicoten Island fault follows the margin of the Lake Superior Basin and is likely to have been active during basin formation. The water-filled-linear southeast of the Slate Islands can be extrapolated to the Port Coldwell Complex where it joins with the Little Pic River lineament, a zone of possible faulting. This linear feature is therefore likely to be related to the Thiel or Big Bay-Ashburton Bay fault system which cuts the Port Coldwell Complex, dated by U-Pb techniques as 1108 ± 1 Ma (Heaman and Machado, 1992). Some movement along the Thiel or Big Bay-Ashburton Bay fault thus postdates the emplacement of the Port Coldwell Complex and likely represents reactivation of earlier regional structural patterns.

Soon after extension and rifting, between 1109 and 1094, closing of the rift began (Cannon, 1994). The southwest-trending arm of the Midcontinent Rift closed by an estimated 30 km, the central portion of the graben was inverted along thrust faults and the southeast-trending arm was dominated by strike-slip motion (Cannon, 1994). Rifting of the lithosphere caused weakening in an otherwise strong continental lithosphere and focused this deformation into the Midcontinent Rift. The graben in eastern Lake Superior

is 30 km wider than in western Lake Superior, consistent with thrusting in the west and strike-slip faulting in the east (Cannon, 1994). Tectonic transport was perpendicular to the Grenville Front with the transition between thrusting and strike slip movement occurring where the Midcontinent Rift changes trend (Cannon, 1994). The abundance of east-north-east faults and displacements in the Midcontinent Rift suggests accommodation perpendicular to the axis of the rift (Mariano and Hinze, 1994a). The Thiel Fault or Big Bay-Ashburton Bay Fault is the most notable transform fault separating distinct segments of the rift: symmetrical graben in the east, asymmetrical half-graben in the west (Mariano and Hinze, 1994a). This accommodation zone is the Trans-Superior Tectonic Zone into which alkalic magma was emplaced north of Lake Superior. Mariano and Hinze (1994a) indicated that this compression is demonstrated by drag faults along the Keweenaw Fault, folds and anticlinal structures in the main basin, and reverse faults to the north and south of central Lake Superior. Mariano and Hinze (1994a) report the presence of an anticlinal structure in eastern Lake Superior bisecting and paralleling the trend of the rift which has as much as 5 km of relief on the basalt layer. The rapid evolution of the Midcontinent Rift from an extensional to a compressional feature occurred at approximately 1080 Ma (Cannon, 1994).

Midcontinent Rift and the Grenville Front Tectonic Zone (GFTZ)

It was suggested by Gorden and Hempton (1986) on the basis of isotopic ages that the Midcontinent Rift was formed as a result of convergence related to the synchronous Grenville orogeny. Cannon (1994) noted that the Midcontinent Rift developed adjacent to the Grenville Front Tectonic Zone and that the evolution from extension to compression at approximately 1080 Ma was coincident with renewal of northwest-directed thrusting in the Grenville Province, probably in response to continent-continent collision. A reversal of movement occurred along the major graben faults of western Lake Superior, and in eastern Lake Superior the compression was taken up in reverse faults oriented normal to the Midcontinent Rift axis and parallel to the Grenville Front Tectonic Zone (Manson and Halls, 1994). In most cases these faults reactivated pre-existing faults within the Archean basement (Manson and Halls, 1994). Cannon (1994) has interpreted geophysical data in southern Michigan as suggesting that the Grenville Front Tectonic Zone is thrust over the Midcontinent Rift (Figure 9).

Midcontinent Rift and the Kapuskasing Structural Zone (KSZ)

The Kapuskasing Structural Zone (KSZ) strikes north of east from the east shore of Lake Superior, but quickly becomes more northerly in strike east of the Lake. This more

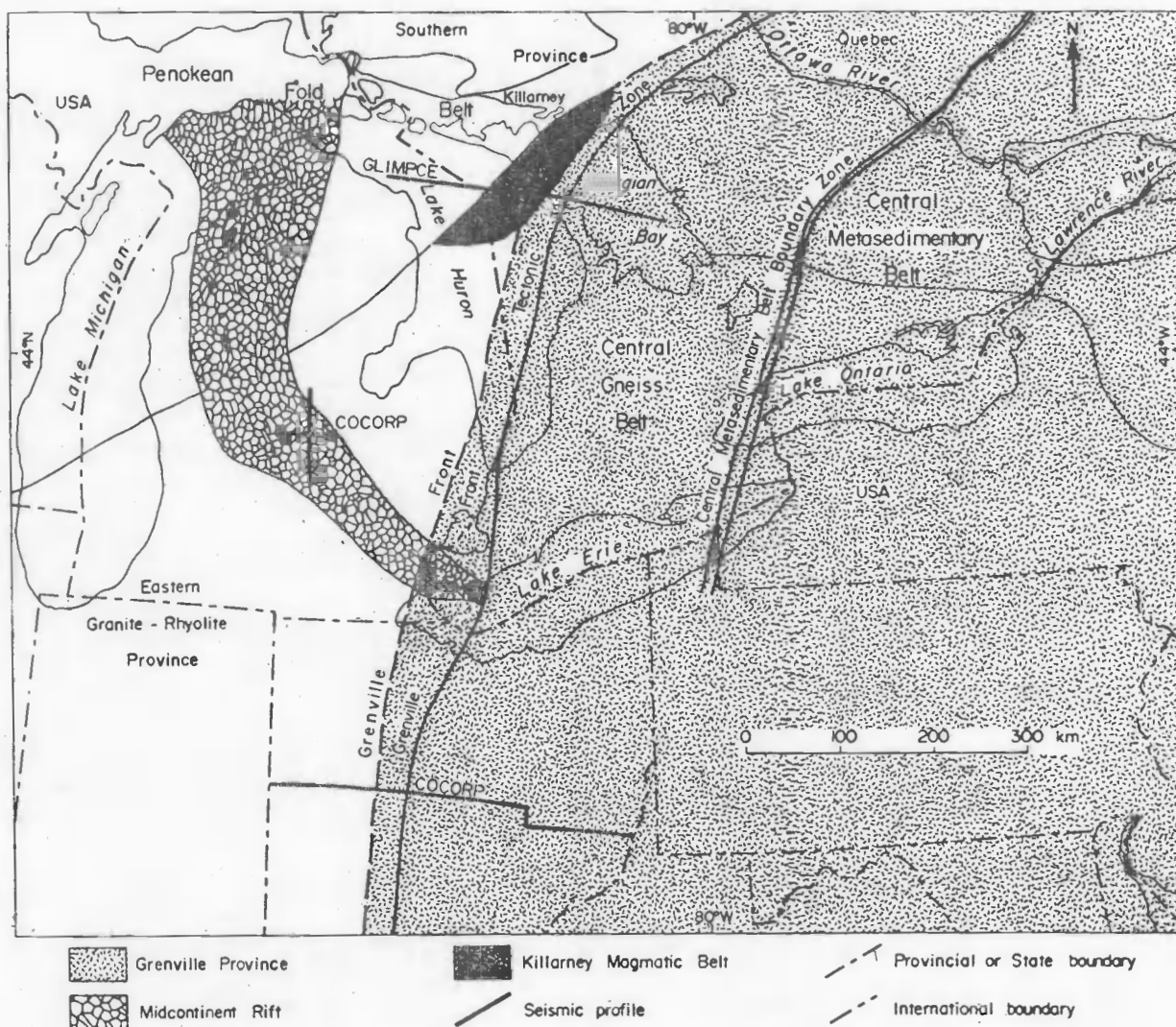


Figure 9: Regional sketch map showing the positions of the Grenville Front, Grenville Front Tectonic Zone, Central Metasedimentary Belt Boundary Zone. The southeastern extension of the Midcontinent Rift in areas covered by Paleozoic are geophysical interpretations. (From Easton, 1992).

northern strike is closely parallel to the trend of the Trans-Superior Tectonic Zone (Figures 3, 10). The KSZ is a broad zone of faulting at the east coast of Lake Superior and a very narrow zone near Hudson Bay to the north. The zone is characterized by positive aeromagnetic and gravity anomalies and west-dipping zones of high refraction velocities (Percival and West, 1994). The zone contains west-dipping faults exposing an approximately 20 km thick section of Archean crust (Percival and Card, 1983) related to transpressive tectonics (Percival and West, 1994). Uplift took place prior to 2.45 Ga (Bursnall et al. 1994; Heaman, 1988), since relatively undeformed Matachewan dikes cut the structure, and perhaps as early as 2630 Ma (Krogh and Moser (1994)). Granulite-facies metamorphism and other deformations took place prior to this KSZ event (Bursnall et al., 1994; Krogh and Moser, 1994). The Kapuskasing Structural Zone is an intracratonic uplift representing 27 km of crustal shortening through brittle upper crustal thrusting (Percival and West, 1994). Within the KSZ the Moho occurs at a depth of about 53 km compared to values near 45 km near Wawa and approximately 35 km near Timmins (Percival, 1990).

Manson and Halls (1994) traced faults from Mamainse Point to the Keweenaw Peninsula and suggested that the Michipicoten Island Fault may be traced into Grindstone Point Fault. Grunsky (1991) mapped a large northeast-trending fault along the Montreal River which separated the Chapleau Gneiss Domain from the Batchawana Volcanic Domain-Ramsy Gneiss Domain. This fault contained gouge, breccia and quartz veins and was correlated with faulting associated with the KSZ. Other northeast-trending faults with a brittle style of deformation occur along the Batchawana and Goulais rivers but their relationship to the KSZ is not as clear (Grunsky, 1991). Sage (1994) correlated the Wawa-Hawk-Manitowik Lake Fault with the Kapuskasing Structural Zone and suggested that it was rooted in Archean tectonics. The extension of this fault toward Lake Superior may account for the straight coastline of Lake Superior west of Wawa. Approximately 15 km east of the shoreline of Lake Superior the Wawa-Hawk-Manitowik Lake Fault is cut by a smaller local fault known as the Firesand River Fault (Sage, 1994). At this intersection, the mantle-derived Firesand River Carbonatite was emplaced, which on the basis of imprecise age dating is likely coeval with the Prairie Lake Carbonatite (Sage and Watkinson, 1991). Numerous alkalic rock and carbonatite complexes occur along the Kapuskasing Structural Zone which were emplaced at about 1.1 Ga (Sage and Watkinson, 1991; Bursnall et al. 1994; Percival and West, 1994). Within the constraints of existing isotopic age dating, alkalic magmatism along the Trans Superior Tectonic Zone and Kapuskasing Structural Zone appear to be contemporaneous. The alkalic rock and carbonatite complexes along the Kapuskasing Structural Zone should also be considered

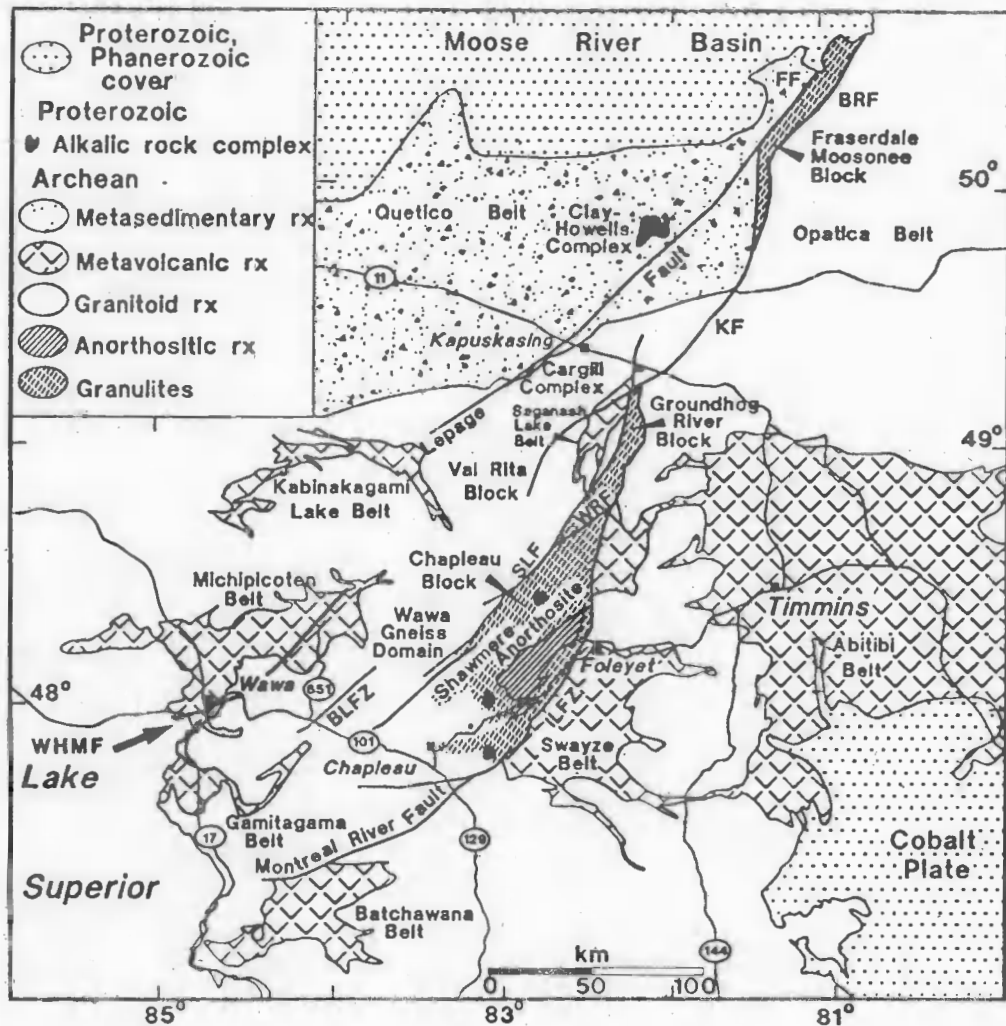


Figure 10: Geologic sketch map of the Kapuskasing Structural Zone and surrounding parts of the central Superior Province, showing major geological and geographical features. BLFZ, Budd Lake fault zone; BRF Bad River fault; FF, Foxville fault; ILFZ, Ivanhoe Lake fault zone; KF, Kineras fault; SLF, Saganash Lake fault; WRF, Wakusimi River fault; WHMF, Wawa-Hawk-Manitowik Lake fault; rx, rock types(modified from Percival and West, 1994).

manifestations of tectonic activity associated with the development of the Midcontinent Rift. It is also likely that late Archean faulting associated with the KSZ was reactivated at 1.1 Ga and played a major role in the development of the Midcontinent Rift in the Lake Superior region.

Summary of Regional Setting

The locations of alkalic rock intrusions north and northeast of Lake Superior are not haphazard or random. The intrusions occur within the Trans-Superior Tectonic Zone and its northern extension with the largest intrusion occurring at the Proterozoic-Archean contact close to the intersection of the Michipicoten Island and the Big Bay-Ashburton Bay Faults. The Trans-Superior Tectonic Zone bisects the Midcontinent Rift and has aeromagnetic, gravity and topographic expression. The Trans-Superior Tectonic Zone is an accommodation zone separating the Lake Superior basin into two contrasting tectonic domains. Development of the Midcontinent Rift is intimately related to developments along the GFTZ and KSZ and likely strongly influenced by early structures in the Archean basement.

The unique positions of the diatreme structures on the Slate Islands is more consistent with the tectonic development of the Midcontinent Rift than meteorite impact (Sage, 1978; Halls and Grieve, 1976).

DAY 1

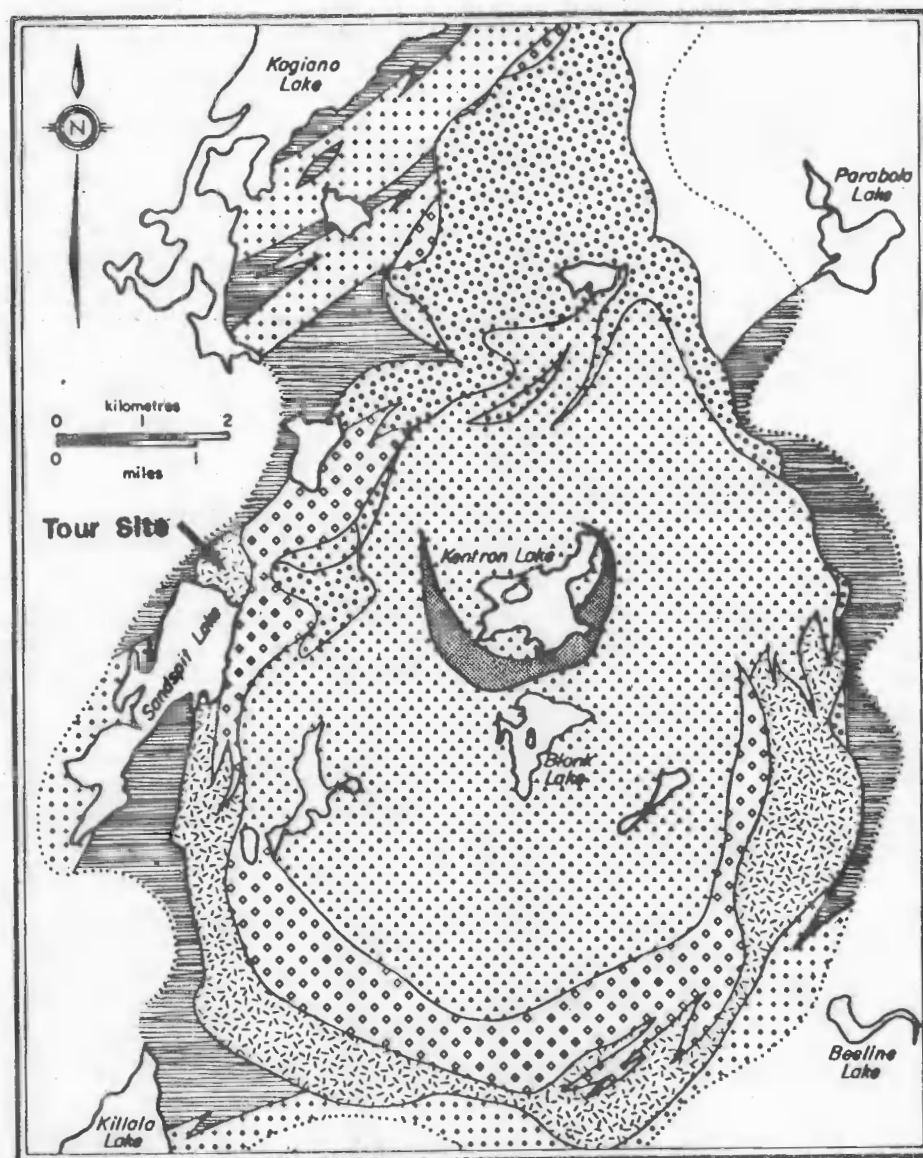
KILLALA LAKE COMPLEX

The Killala Lake Complex is an alkalic complex consisting of an outer ring of troctolite followed inward by nepheline syenite and then an inner core of amphibole syenite (Figure 11). The interior syenites are coarse-grained, equigranular, grey to buff on fresh surface in the southern part of the complex and red to reddish brown, coarse-grained, inequigranular, seriate to porphyritic, in the northern part. Several minor lithologies are also locally present. In the centre of the intrusion on the shores of Kentron and Blank Lakes mafic banding is very well developed in the syenites (Sage, 1988). The size and number of xenoliths increase to the north as do the number of septa of enclosing wall rocks projecting into the syenite mass.

The Killala Lake complex has been dated by Rb-Sr techniques as being 1050 ± 35 Ma (Bell and Blenkinsop, 1980). This age is essentially coeval with more precise U-Pb isotopic age dating at the Port Coldwell Complex.

The Killala Lake Complex has a surface area of approximately 110 km^2 . The large number of xenoliths and prevalence of wall rock septa suggest that the northern part of the ring complex is exposed at a higher structural level than is the southern part and the intrusion which may plunge south to southwest toward the Lake Superior Basin. The isomagnetic contours on aeromagnetic maps of the complex outline a circular pattern in contrast to the tear-drop shape defined by mapping (ODM-GSC 1963f, g; Coates, 1970; Sage, 1988) (Figure 12). The northern projection of the ring complex has been interpreted as a sheet-like or petal-like laccolithic mass extending north of, and overlying the outer gabbro ring which on the basis of aeromagnetic data likely completely encloses the complex (Coates, 1970; Sage 1988). The arcuate, open-to-the-north horseshoe shaped lithologic distribution suggests the possibility of a former caldera structure (Sage, 1988). Coates (1967, 1970), Bathe (1977), Wanless (1976) and Sage, (1988) described detailed geology of the Killala Lake Complex.

The Killala Lake Complex lies at the intersection of topographic and aeromagnetic linears. A prominent aeromagnetic linear connects the Killala Lake Complex with the Prairie Lake Carbonatite, of similar age, to the southwest. Williams (1989) has identified a zone several kilometres wide of deformed rocks along this aeromagnetic



KILLALA LAKE ALKALIC COMPLEX

PROTEROZOIC

-  Brown syenite
-  Nepheline syenite
-  Buff syenite
-  Syenodiorite
-  Gabbro
-  Diabase

ARCHEAN

-  Quartz monzonite to granodiorite
-  Metasediments

Figure 11: Geologic sketch map of the Killala Lake Alkalic Rock Complex (from Sage, 1986, 1988)



1.6 km

Figure 12: Aeromagnetic map of the Killala Lake Alkaline Rock Complex (ODM-GSC, 1963e, f).

trend which he referred to as the Killala Lake Deformation Zone. A prominent topographic linear along the east side of the complex has been interpreted by Coates(1970) as a fault(Boomerang Lake Fault) and this fault may be interpreted from air photographs to join with the Little Pic River lineament on the west side of the Port Coldwell Complex. The Killala Lake Complex lies within the northern part of the Lake Superior Tectonic Zone and along the interpreted extensions of the Big Bay-Ashburton Bay Fault(Sage, 1978) or Thiel Fault(Klasner et al. 1982). The trend of this tectonic zone parallels approximately the Kapuskasing Structural Zone. The topographic linear connecting the Port Coldwell and Killala Lake complexes may be extrapolated northward to the Chipman Lake area where fenite and carbonatite dikes are widespread(Sage, 1985)(Figure 13). Offset of lithologies across this lineament at Chipman Lake is 0.8 km(Sage, 1985).

Economic Geology

The Killala Lake complex has been prospected for its copper and nickel content in the early 1950s(Sage, 1988). Bath(1977) examined some of the drill core remaining from this effort and recognized pyrrhotite, chalcopyrite, cubanite, pentlandite, and valleriite. The complex remains relatively untested for PGE. Within the area of Blank Lake, near the centre of the intrusion, zircon and pyrochlore are present within narrow pegmatite dikes(Sage, 1988). Pyrochlore and zircon are of mineralogical interest only. The complex is currently undergoing examination as a possible source of building stone.

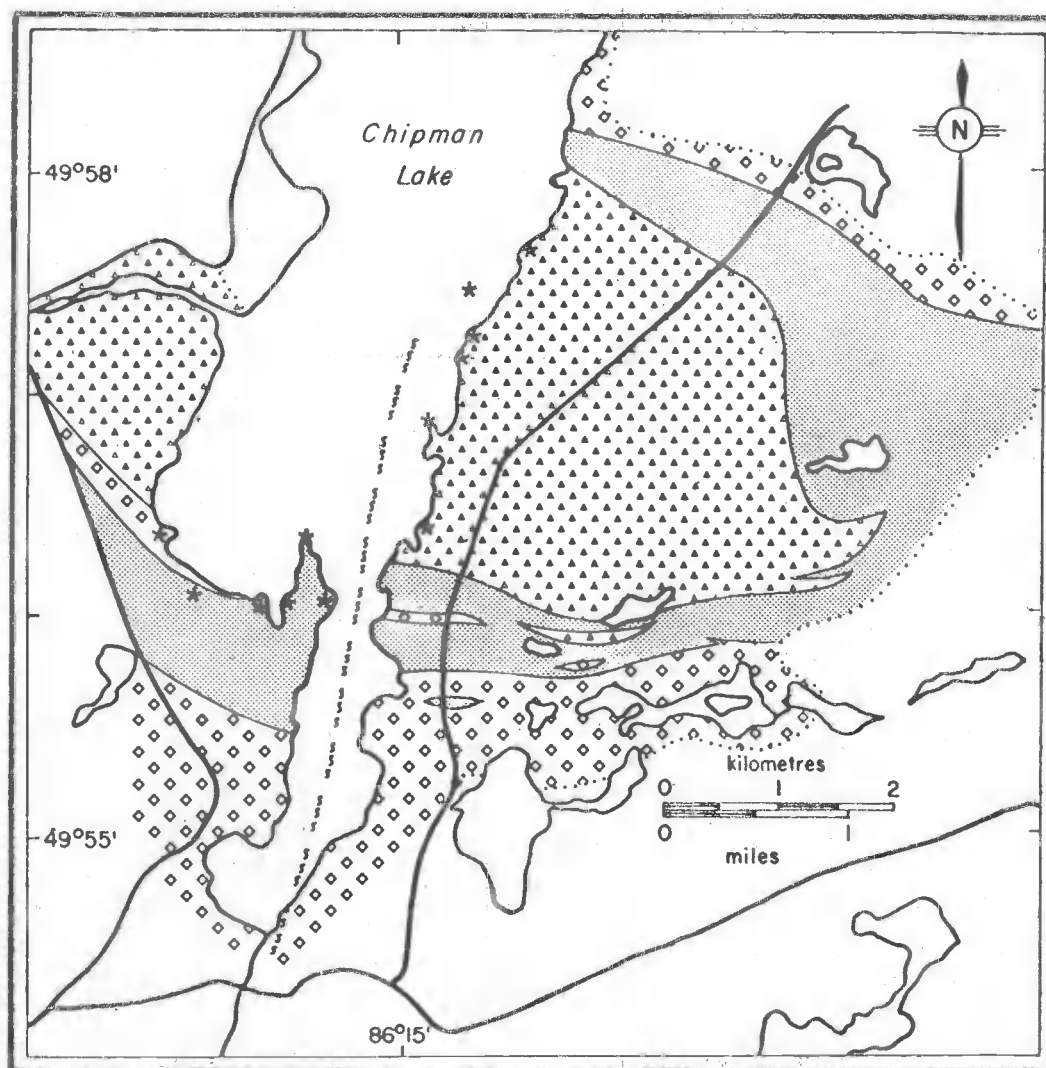
The road log is from the intersection of Highway 17 and the Dead Horse Creek access road. Recent logging activities in the region now permit access to the margins of the Killala Lake Complex.

@ 18.4 Km Jackpine Road, turn right

@ 33.7 Km Prairie Lake Carbonatite intersection on left. Continue north and cross the Little Pic River bridge and stay to the right hand branch of the road.

@ 55.2 Km Turn right on skid road. This road is located just north of Sandspit Lake.

@ 0.40 Km Park in cleared area.



CHIPMAN LAKE AREA

PROTEROZOIC

- ★ Fenite and carbonatite dikes

ARCHEAN

- ◊ Diorite to quartz monzonite intrusive rocks
 ▲ Diorite to syenodiorite intrusive rocks
 ■ Mafic metavolcanics

--- fault

— road

Figure 13: Geologic sketch map of the Chipman Lake area showing faulting along the long axis of Chipman Lake. This fault may be the northern extension of the Boomerang Lake fault that passes along the east side of the Killala Lake Alkaline Rock Complex (from Sage, 1985, 1986).

Stop 1a Walk approximately 470 m along skid road in a SE direction and up hill(See Figure 14; Sage, 1988).

Medium- to coarse-grained equigranular olivine gabbro to troctolite occurs along the north side of road as rounded grey outcrops. Outcrop is cut by dikes of pink to red syenite and nepheline syenite. Bathe(1977) described samples from similar gabbro outcrops as being composed of 45 to 50 % plagioclase(An56 - An65), augite clinopyroxene 20 to 25 %, olivine 5 to 30 % and accessory apatite, biotite, green amphibole, oxides and sulphides. Noritic gabbro and larvikite are described by Bathe(1977) as being associated with the olivine gabbro to troctolite phase. In thin section the olivine-plagioclase have a corona structure where in contact, attributed to a reaction with water under middle to upper amphibolite-facies conditions(Laderoute, 1984).

Stop 1b approximately 130 m east of 1a

Small outcrop of coarse-grained nepheline syenite with up to 10 to 15 % nepheline. The nepheline syenite has been estimated to contain approximately 10 to 15 % nepheline, 0 to 25 % dark green to green brown amphibole, 30 to 70 % perthite and 0 to 65 % sodic plagioclase(Sage, 1988). Minor amounts of clinopyroxene and biotite are present with accessory sphene, magnetite, apatite, sericite and carbonate.

Stop 1c approximately 70 m east of 1b.

Nepheline syenite with lower nepheline content.

Stop 1d approximately 145 m east of 1c

Coarse-grained equigranular buff to pink syenite. Some pitting of the surface may be due to weathering of nepheline but most is due to the mafic mineral content. The buff syenite contains approximately 5 to 45 % plagioclase(An28-An39), 5 to 25 % amphibole, and 15 to 95 % perthite(Sage, 1988). Minor amounts of olivine 0 to 1 %, biotite 0 to 10 % and clinopyroxene 0 to 5 % may also be present in some specimens. The accessory minerals are apatite, magnetite, sericite, carbonate, zircon, sphene, chlorite and quartz(Sage, 1988).

Stop 1e approximately 55 m east of 1d

Small outcrop as previously described.

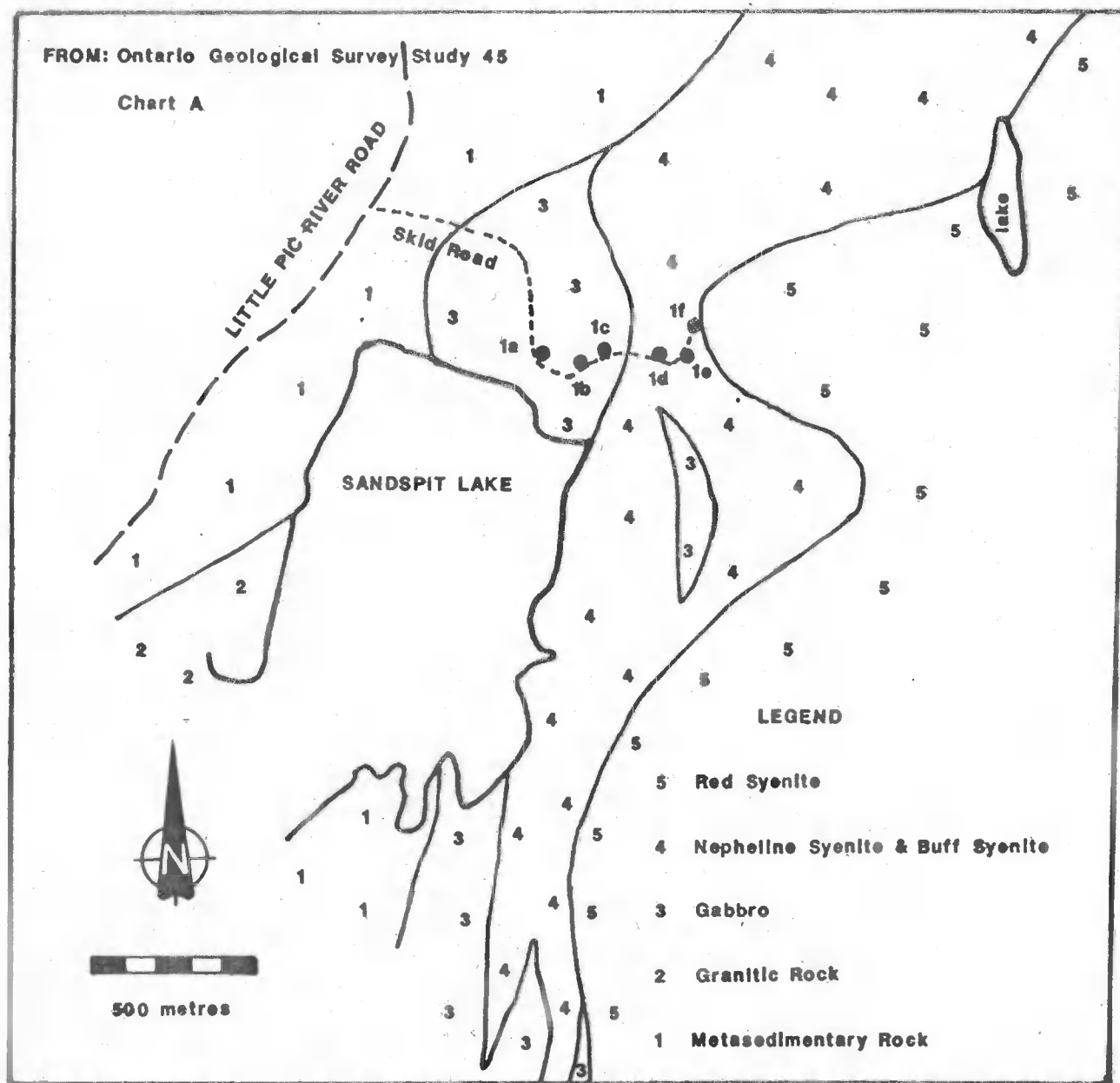


Figure 14: Geologic sketch map showing field trip sites, Killala Lake Alkalic Rock Complex. Base map from Sage(1988, Chart A).

Stop 1f approximately 125 m east of 1e; skid road has deteriorated.

Coarse-grained equigranular amphibole syenite. Feldspar somewhat lath-like and mafic mineral content approximately 15 to 20 %. Outcrop similar to 1d.

Return to parking area and go back to the main road. Turn left back the way you came. Drive 21.5 km to intersection of a skid road and Jackpine Lake Road.

Turn right onto skid road and continue to 25.6 km.

Turn right onto partially overgrown skid road and continue until odometer reads 26.6 km. At this point the road splits. Take the right-hand fork and continue to the odometer reading of 26.7 km. An overgrown skid road leads off to the right up a steep hill. Park at the road junction or drive the short distance to the base of the hill and park.

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PRAIRIE LAKE CARBONATITE

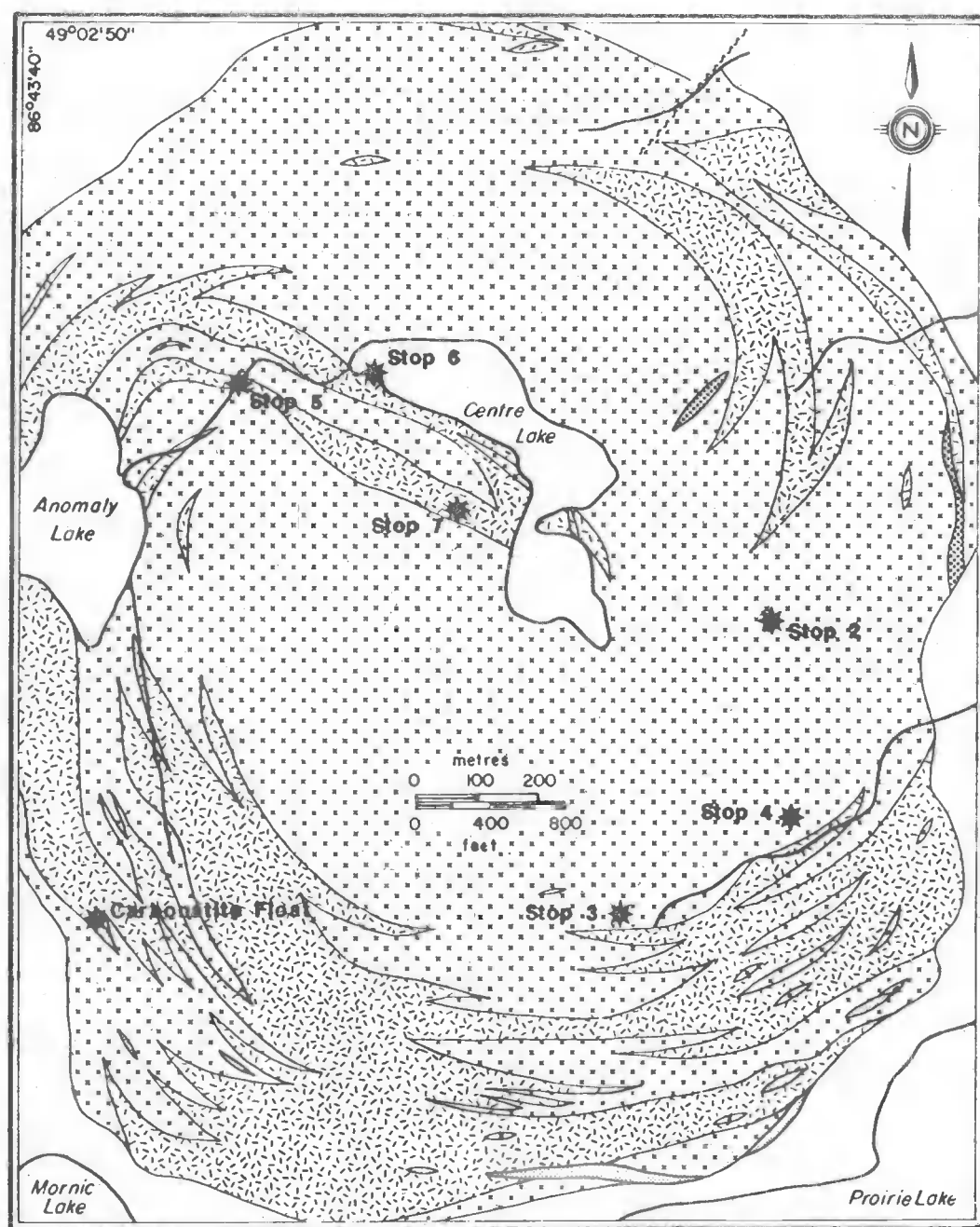
The Prairie Lake Carbonatite consists of arcuate units of sovite, silicocarbonatite, ijolite and wollastonite ijolite(Figures 15, 16a,b). Minor amounts of biotitite(glimmerite) may locally be present. The complex forms a prominent hill, which, in spite of its topographic relief, has only rare outcrops. Weathering is deep and samples of fresh rock may be obtained only through diamond drilling, trenching or in logging skid roads where lumps resistant to weathering have been exposed by heavy equipment. The carbonatite complex has a surface area of approximately 8.8 km².

The complex has been dated at 1033 $\pm$ 59 Ma by the Rb-Sr technique(Bell and Blenkinsop, 1980) and is considered to be essentially coeval with the Killala Lake and Port Coldwell Alkalic rock complexes. The Prairie Lake Complex has a prominent aeromagnetic anomaly approximately 2.5 km in diameter(ODM-GSC, 1963a). The aeromagnetic anomaly is approximately 1400 gammas absolute total field above the surrounding background area which is underlain by granitoid rocks(Sage, 1987)(Figure 17).

North-trending topographic linears from the north shore of Lake Superior intersect northeast-trending topographic and aeromagnetic linears at the site of the Prairie Lake Carbonatite. The northeast trending linear connects with the Killala Lake Complex. Williams(1989) recognized a deformation zone between these two complexes which he referred to as the Killala Lake deformation zone. This zone is several kilometres wide and defines the boundary between the Wawa and Quetico subprovinces(Williams, 1989). The complex lies within the northern part of the Trans-Superior Tectonic Zone(Klasner et al, 1982) and along linear trends subparallel to the Big Bay-Ashburton Bay Fault or Theil Fault trend(Sage, 1978; Klasner et al, 1982). The geology and mineralogy of the Prairie lake Carbonatite have been described by Watkinson,(1971, 1973, 1976), Melnik(1984) and Sage(1987). Sage(1987) cites numerous unpublished company reports on the complex, most of which can be obtained from the Assessment Files, Resident Geologist's Office, Thunder Bay.

Economic Geology

Uranium mineralization was discovered at "Jim's Showing" by J. Gareau working for Newmont Mining Corporation of Canada Ltd. in 1968 and the company completed radiometric, magnetic and geochemical surveys over the intrusion. At "Jim's



PRAIRIE LAKE CARBONATITE

PROTEROZOIC

- | | | | |
|--|------------------------|--|---------|
| | Alkaline Dike Rocks | | Archean |
| | Sovite Silicocarbonate | | |
| | Ijolite | | |
| | Fenite | | |

fault stream

Figure 15: Geological sketch map of the Prairie Lake Carbonatite showing field trip sites and potential sites. (Modified from Sage, 1986, 1987).

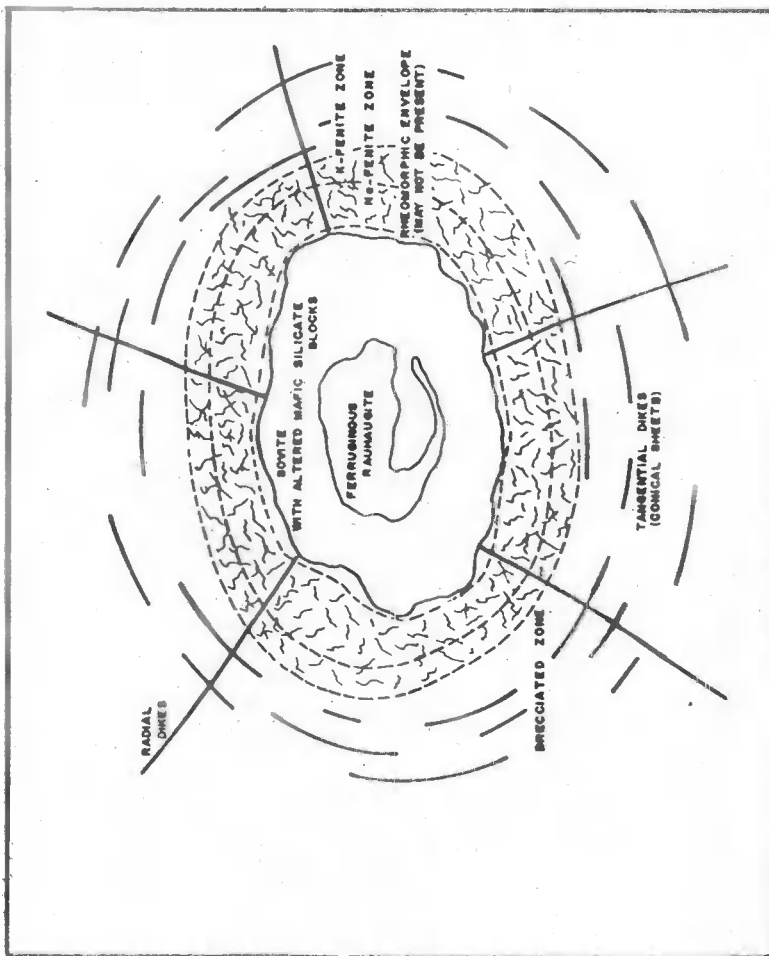


Figure 16a: Idealized plan view of a carbonatite intrusion showing the idealized distribution of rock types (from Sage, 1986).

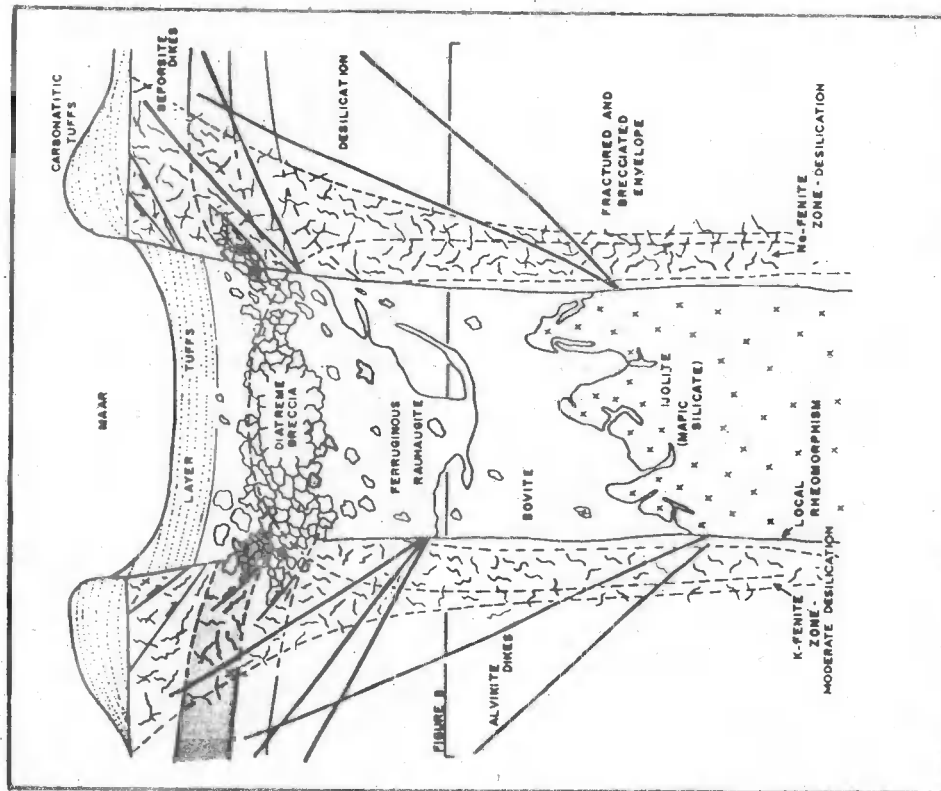


Figure 16b: Idealized cross section through a carbonatite intrusion illustrating an idealized distribution of rock types (from Sage, 1986).

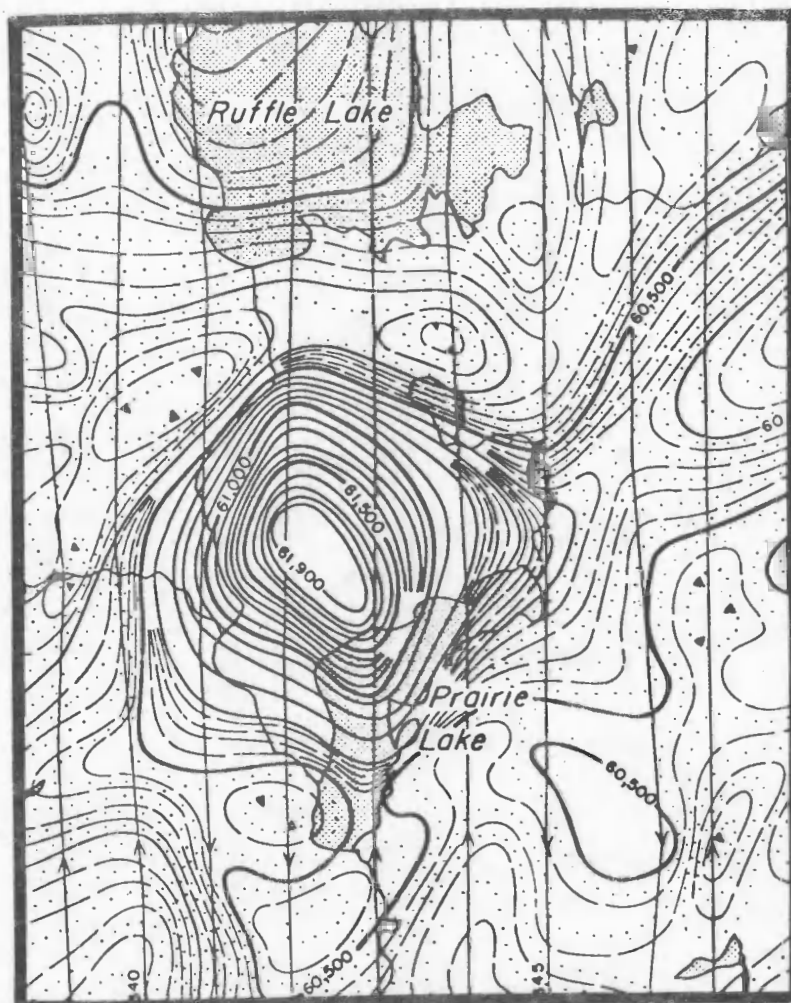


Figure 17: Prominent aeromagnetic anomaly that outlines the Prairie Lake Carbonatite (ODM-GSC, 1963e).

Showing" the company outlined 109,024 tons grading 0.12 % U_3O_8 in a deeply weathered ferruginous dolomite. This was expanded by Nuinsco Resources Limited in 1975 to 200,000 tons grading 1.8 pounds U_3O_8 and 5.0 pounds Nb_2O_5 per ton in a zone 300 feet long and 275 feet deep. Watkinson(1976), using the microprobe, identified the uranium-bearing mineral as pyrochlore which displayed increasing uranium contents from core to rim. Uranium may comprise up to 30 % of pyrochlore (Watkinson, 1976). In 1975 the complex was examined for possible residual apatite deposits by the International Minerals and Chemical Corporation(Canada) Ltd. and in 1983 New Insko Mines Ltd. tested the carbonatite complex for its wollastonite potential. Sage(1987) provided additional information regarding these exploration efforts.

Walk approximately 400 m up the skid road.

Stop 2. Medium- to coarse-grained inequigranular ijolite is present as weathered lumps and outcrop. This exposure is typical of the coarser-grained phases of the Prairie Lake Carbonatite. In addition to the pyroxene and nepheline, calcite, magnetite and garnet are present in this outcrop and titanite has been observed in similar rocks within the complex. The outcrop is deeply weathered with secondary carbonate along joint surfaces. This exposure is much larger than most exposures within the intrusion.

Return to the vehicles. You may note the presence of phlogopite in the skid road as you return to the vehicles.

Drive back approximately 0.1 km where the road splits and take the right hand fork. Drive along this partially overgrown skid trail up the hill for approximately 0.8 km and park.

Stop 3 Wollastonite ijolite. This site displays deeply weathered wollastonite ijolite which has been torn up during logging operations. The material is pegmatitic and prismatic crystals of wollastonite up to 15 cm or more have been found here. The wollastonite will sometimes appear as veins cutting the ijolite. The wollastonite crystals are perpendicular to the vein trend.

Return or drive back 0.35 km. A flagged trail leads off to the left into the bush for approximately 100 m.

Stop 4 Orbicular ijolite(optional). The orbicular ijolite never occurred in solid outcrop projecting above ground level but consisted of large frost-heaved blocks. Collectors

have dug a pit at the discovery site and over the years large quantities have been removed. The site was overgrown and the pit filled in the fall of 1994 and it will have to be exposed with heavy equipment for additional sampling. A visit to the site will be depend on whether arrangements can be made to expose deeper material.

This material is one of the finest examples of orbicular texture known and this is the only known occurrence of this texture in rocks of this composition(Sage, 1987). The orbicules consist of a multitude of concentric bands consisting of varying proportions of aegirine-augite, nepheline and melanite garnet. Apatite and biotite are also present as well as a white alteration mineral, possibly former melilite(Sage, 1987). The orbicules are up to 3 cm in diameter and the largest have a medium-grained equigranular ijolite core. It is assumed that all orbicules will have this equigranular core if they have been cut through the centre. Several large blocks of the material recovered in initial sampling suggest that the orbicules may occur along distinct bands within equigranular ijolite.

Return to vehicles and return to the main road (Jackpine Road).

Turn right on the main road and go 15.2 km to the junction with the Dead Horse Creek access road.

Turn right and go north 10.1 km and park at the side of the road. A small lake will lie to the west and a steep hill occurs to the east across a flat area used as a storage area for logs.

Walk approximately 100 metres to the east and along the skid road up the steep hill. This hill is the west side of the Prairie Lake Carbonatite. Various cobbles and boulders of sovite and silicocarbonatite have been turn up by logging equipment in the skid road. In deep cuts into the deeply weathered carbonatite-rich soil relict primary banding may sometimes be observed.

At the top of the hill a partly overgrown skid road leads into the interior of the carbonatite and a former campsite near Centre Lake. This trail is not a difficult walk and is approximately 1.2 km in length. Due to the length of the walk and time involved these sites will not be visited unless access can be gained to all sites described for Stops 5 thru 7.

Optional stops 5, 6, and 7

Stop 5 Sovite and banded sovite. This is the largest natural outcrop on the Prairie Lake Carbonatite and occurs in the stream bed of the small stream connecting Centre and Anomaly Lakes. The outcrop is weathered and thickly mantled with moss. The stream bed contains abundant rhombs of carbonate as well as grains of phlogopite and magnetite. This mineral assemblage is typical of sediment found in streams flowing through a deeply weathered carbonatite.

Stop 6 Frost-heaved sovite. Some specimens may contain some ijolitic material as elongated clots or irregular masses. This frost-heaved material occurs on the south shore of Centre Lake but the exposure was flooded out in the fall of 1994.

Stop 7 Trench in deeply weathered, radioactive, ferruginous dolomite of "Jim's Showing". This trench was caved and poorly exposed during mapping in the 1970's and could not be relocated in the fall of 1994. This trench contains material typical of weathered ferruginous dolomite that hosts U-Nb minerals.

Return to the vehicles and drive back down the Dead Horse Creek Access Road toward Highway 17. At approximately 24.9 km park in the open area on the east side of road on the north side of the bridge over Dead Horse creek. At the far end of this partially overgrown parking area is a trail leading to the U-Y showing at Dead Horse West.

DEAD HORSE CREEK DIATREME

The Dead Horse Creek Diatreme is the largest (1600 x 400 m) of several diatreme structures occurring east of the Port Coldwell complex (Sage, 1982, 1991) (Figure 18). The complex consists of a broad spectrum of heterolithic breccias that have undergone varying degrees of alteration and are variably radioactive (Sage, 1982). The diatreme was discovered by Mr. Gordon Yule while completing a field assignment at Lakehead University in 1976 (Mitchell and Platt, 1977; Gordon Yule, personal communication, 1995). Since the diatreme was radioactive prospectors working for Gulf Minerals Canada Ltd. investigated the diatreme structure for its uranium content. The complex occurs within the thermal aureole of contact metamorphism of the Port Coldwell Complex (Walker, 1967) and rare granitoid fragments found in the eastern part of the breccia may be from the Port Coldwell Complex (Sage, 1982). Preliminary unpublished U-Pb isotopic ages on zircons from the Dead Horse Creek West subcomplex have given values of 1128.7 ± 6 Ma (1.82% discordant) and 1112.7 ± 4 Ma (-2.49% discordant) (Krogh and Wilkinson, 1995, personal communication). These preliminary ages suggests the possibility that the Dead Horse Creek diatreme is slightly older than the Port Coldwell Complex dated at 1108 ± 1 Ma by Heaman and Machado (1992). Until the isotopic investigations are completed the age relationship between the diatreme and the alkalic rock complex remain somewhat speculative since critical outcroppings needed to establish a relationship are lacking. The age dating does indicate that the mineralization within the Dead Horse Creek diatreme is Keweenawan in age and related to alkalic rock magmatic events of the Midcontinent Rift. Failure to correlate lithologies across Dead Horse Creek have prompted Walker (1967) and Sage (1982) to propose that a fault may occur along the trend of the creek. The minor subcomplexes generally have a northwest-trending long axis suggesting that they may occupy crosscutting structures (Sage, 1982).

The diatreme lies at the northern end of the Big Bay-Ashburton Bay Fault (Sage, 1978) or Thiel Fault (Klasner et al. 1982) within the Trans Superior Tectonic Zone. The diatreme lies within north, to east of north, trending regional linears, some of which are likely to be faults.

On the basis of the anomalous concentrations of K, Nb, Zr, and LREE elements, Sage (1982) interpreted the Dead Horse Creek Diatreme to be the high level expression of a carbonatite, possibly coeval with the emplacement of the Prairie Lake Carbonatite. Smyk et al. (1993) came to a similar conclusion and proposed that the carbonatite-emplacement event was followed by a later mineralizing event

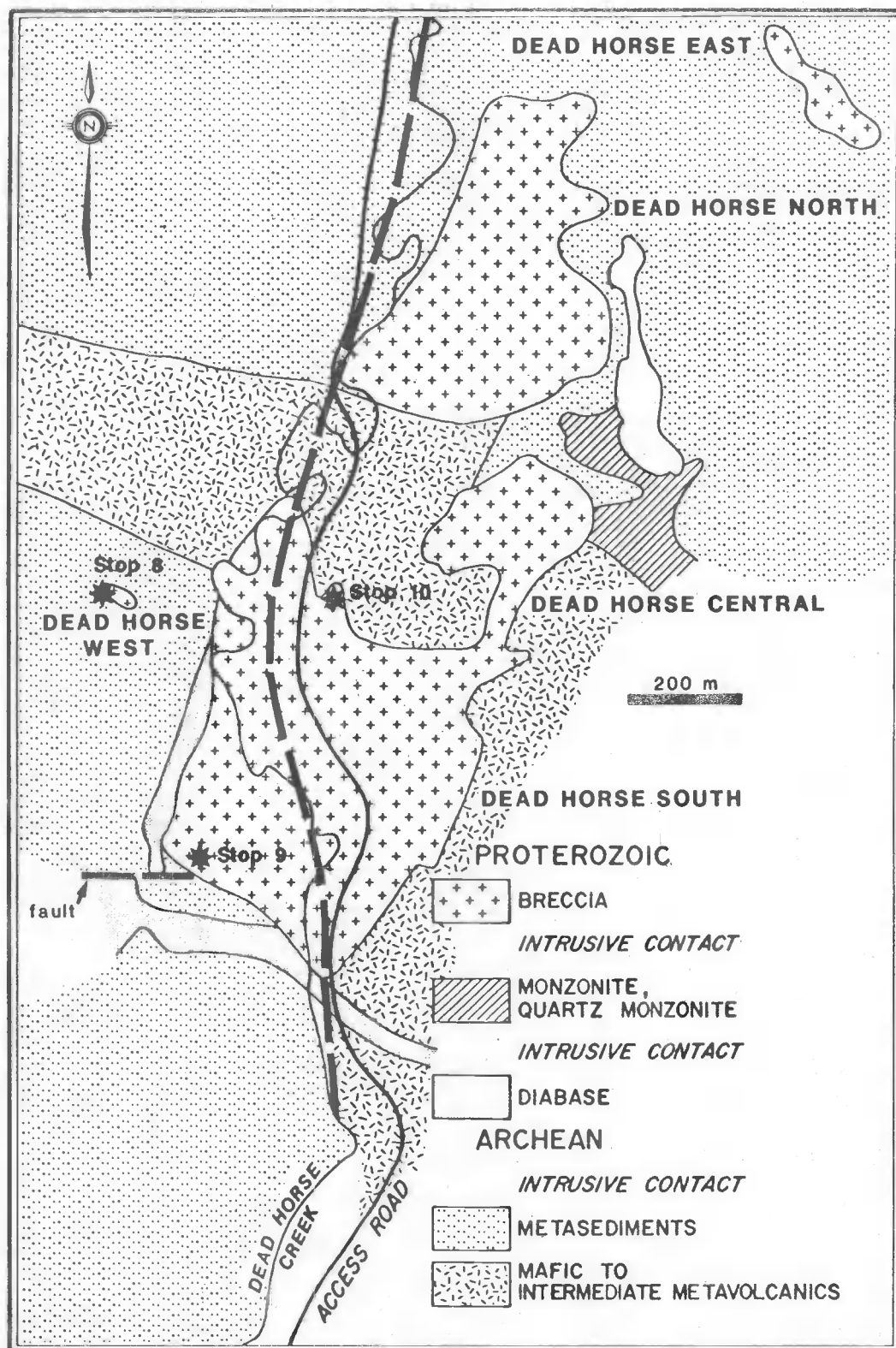


Figure 18: Geological sketch map of the Dead Horse Creek Diatreme with field trip sites indicated(modified from Sage, 1982).

of an agpaitic nature. This suggestion is based on the observation that HREE(Gd-Lu) are concentrated in the mineralized shear cutting the Dead Horse Creek West diatreme breccia(Smyk et al., 1993). This HREE-enrichment is more characteristic of agpaitic syenites and peralkaline granites than carbonatites that are typically LREE-enriched. Smyk et al. (1993) correlated this late-stage HREE mineralization with the possible emplacement of the Port Coldwell complex.

Economic Geology

The complex was prospected for its uranium content by Gulf Minerals Canada Ltd. who found the best mineralized area at the Dead Horse Creek West subcomplex. The company completed mapping, sampling, diamond drilling and radiometric surveys over the complex in 1977-1978. Unocal Canada Ltd. examined the Dead Horse Creek Diatreme in the late 1980's principally for its Y content. The diatreme has very high concentrations of Zr and Be which have not been examined for their economic potential.

Take the trail SW for approximately 750 m from the parking area. This trail is up and down and numerous wet and rocky areas exist.

Stop 8 Unocal Y prospect - Dead Horse Creek West. The U-Y-Be-Zr mineralization found within the Dead Horse Creek Diatreme is most highly concentrated in the Dead Horse West subcomplex breccia and is associated with shearing trending 290 with vertical dip. The mineralization is best exposed at the west end of the stripped area in a narrow zone approximately 15 cm wide. The richest mineralization occurs as very fine-grained, chocolate-brown, vitreous material. Silicification accompanied the mineralization. Highest assays occur over a width of 1.5 m and length of 82 m(Smyk et al., 1993). The main minerals are calcium zirconosilicate with subordinate zircon, uraninite, thorite, monazite-(Ce) and xenotime-(Y)(Smyk et al., 1993). Phenakite(Be_2SiO_4) is the Be-bearing mineral (Sage, 1982, Smyk et al., 1993). Smyk et al.(1993) reported that xenotime occurs as inclusions in niobian rutile and that the monazite is associated with calcite, K-feldspar and riebeckite. Assays range up to 11.6 % Zr, 0.6 % Be, 2.5 % Th, 250 ppm Sc, 1850 ppm Y, 300 ppm Nb, 903 to 1004 ppm HREE(sum of Gd to Lu) and 4600 ppm U(Smyk et al. 1993).

In the central stripped area of Dead Horse Creek West a heterolithic breccia is well exposed. This breccia contains subrounded to subangular clasts of white to pink quartzite

rocks of the Sibley Group, the nearest outcrops of which now exist 60 km to the west on islands in Lake Superior near Rosspoint. The clasts are up to 0.3 m in diameter and have been exposed by stripping (Sage, 1982). At the south end of the stripped area a 10 to 15 cm wide quartz vein trending 115 and dipping south contains phenacite (Be_2SiO_4) (not visible to the naked eye).

At the eastern end of the stripped area exposures of heterolithic breccia and a lamprophyre (carbonatite) dike are present.

Follow a flagged trail from the partially stripped area below the main stripping for a distance of approximately 550 m to the south.

Stop 9: Cliff face of heterolithic breccia. This is one of the larger exposures of breccia close to the western contact of the Dead Horse Creek South subcomplex. The cliff is approximately 10 m high and 100 m long and gives a strong response on a scintillometer, particularly on the K channel. The clast-supported, angular to subangular breccia is red to brick red with clasts commonly displaying concave centres and raised rims suggesting silicification of the margins. Clasts up to 1 m occur but most are less than 0.3 m. The matrix consists of carbonate, biotite, amphibole and quartz and because of the high carbonate content weathers recessively, leaving the clasts in relief. Exposed within the face of the cliff is a carbonate-rich lamprophyre dike 28 cm wide, trending 250 and dipping 75 south. This dike sharply cross cuts the breccia and is visually estimated to contain 10 to 15 % opaque minerals, 50 to 70 percent carbonate and 30 to 40 % biotite (Sage, 1982).

Return to the vehicles and continue south to 25.7 km and park on the east side of the road. This is on a curve so be careful.

Walk a few metres east of the road to the outcrop of heterolithic breccia exposed on the side of the hill.

Stop 10: Scapolite-replaced heterolithic breccia - Discovery outcrop. This outcrop is the first outcrop on the Dead Horse Creek Diatreme examined by prospectors working for Gulf Minerals Canada Ltd searching for the cause of radioactivity in the area. The breccia weathers grey, is clast-supported, and has a "ragged jagged" surface (Sage, 1982). The fragments are angular to rounded and generally less than 0.3 m in

maximum dimension. The clasts are extensively altered along the margins and scapolite in fibrous crystals up to 5 cm may replace parts of some clasts. On weathered surface the scapolite is easily recognizable by its white fibrous habit; however, on fresh surface it is extremely difficult to recognize in hand sample. In thin section the matrix consists of quartz, carbonate, amphibole, opaques and scapolite and the clasts show replacement by carbonate, amphibole and scapolite(Sage, 1982).

Return to the vehicles and continue south to Highway 17 at 28.6 km.

Time permitting we will go to the optional stop, the McKellar Creek diatreme. From the Junction of the Dead Horse Creek access road turn right(west) 2.2 km to McKellar Creek. The trail to the McKellar Creek Diatreme is on the west side of McKellar Creek and one must park on the shoulder of Highway 17 so be careful. There is a skid road going south on the east side of McKellar Creek which may also be used as a parking site.

Walk approximately 430 metres north along the skid road and the two outcrops of breccia representing this diatreme occur on the east side of the trail.

Stop 11: McKellar Creek diatreme. This diatreme was located by Walker(1967) who interpreted it to be a possible Animikie conglomerate. The fact that it was radioactive and most likely intrusive into the enclosing rocks was recognized much later(Sage, 1982). The McKellar Creek Diatreme is a clast supported heterolithic breccia similar to that exposed at stop 9 on the Dead Horse Creek South subcomplex. The clasts are red from alteration, angular to subrounded, and generally 0.3 m or less in maximum dimension. Sibley Group clasts are abundant and in thin section the quartz displays curved deformation lamellae. In addition to Sibley Group clasts and clasts from the surrounding Archean rocks, Rove Formation clasts have also been identified by thin section examination(Sage, 1982). The space between the clasts is carbonate-rich and generally weathers low leaving the clasts in relief. Behind the outcrops exposed by the road, a 2 x 3 m clast of rounded Sibley Group rock may be examined. On the basis of outcrop and soil composition the McKellar Creek Diatreme is estimated to be 240 x 60 m in size and on the basis of similar alteration styles, interpreted to be coeval with the Dead Horse Creek Diatreme(Sage, 1982). At present, the closest outcroppings of Sibley Group rocks are 60 km west, on islands in Lake Superior near Rossport.

DAY 2

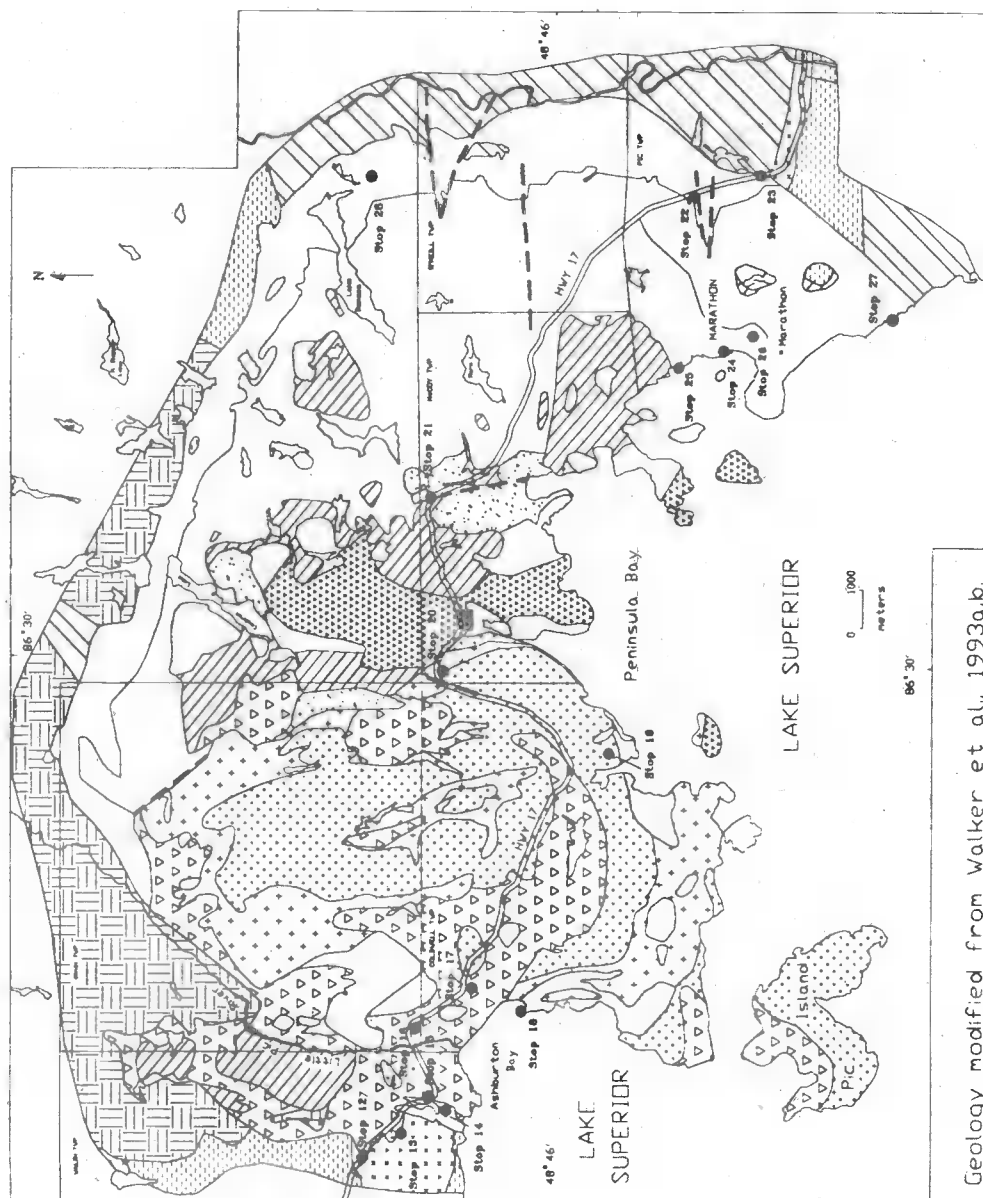
The second day is devoted to the Port Coldwell Alkalic Rock Complex and will consist of a west to east traverse through the complex along Highway 17. The tour starts at the Dead Horse Creek access road and distances are given from that point. The Port Coldwell Alkalic Rock Complex has been the subject of many field trips in the past and the reader may wish to refer to these guidebooks for additional information (Puskas, 1970; Mitchell and Platt, 1977, 1982, 1994). Several stops described by Mitchell and Platt, (1994) are revisited and the reader should refer to their guidebook for supplementary detailed mineral chemistry.

General Geology

The Port Coldwell Complex occurs within the Trans Superior Tectonic Zone (Klasner et al. 1982) where the Big Bay-Ashburton Bay (Sage, 1978) or Thiel Fault (Klasner et al. (1982) intersect Archean rocks of the Superior province on the north shore of Lake Superior. This fault may be extrapolated 140 km north, as far as Chipman Lake using airphotographs, LANDSAT and local geologic mapping (Coates, 1970; Sage, 1985). The Port Coldwell Complex occurs on the northern flank of the Lake Superior Basin at or close to the Proterozoic-Archean contact (Sage, 1991). The Slate Islands, located approximately 10 km southwest of the Port Coldwell Complex, have Keweenaw volcanic rocks of the Osler Group exposed on the shoreline and as isolated blocks in diatreme structures (Sage, 1991). These islands occur very close to the intersection of the Michipicoten Island Fault and the Big Bay-Ashburton Bay Fault or Thiel Fault. The Michipicoten Island Fault appears to closely follow the Proterozoic-Archean contact. The Port Coldwell Complex has been mapped by Tuominen (1967), Currie (1980) and more recently by Walker et al. (1993a,b). The Port Coldwell Complex is one of the largest alkalic rock complexes in the world having a diameter of approximately 25 km and a surface area of approximately 490 km² (Figure 19). Milne (1967) and Muir (1982) completed a limited amount of mapping along the eastern margin of the intrusion and Walker (1967) completed a similar area of mapping along the western margin. Walker (1967) defined a thermal aureole of contact metamorphism of approximately 1.6 km width along the western margin which would envelope the Dead Horse Creek Diatreme structure. Aubut (1977) determined that this thermal aureole attained pyroxene hornfels facies where in direct contact with the intruding gabbro.

The Port Coldwell Complex consists of 3 ring complexes (Mitchell and Platt, 1978; Currie, 1980) superimposed on each other and younging systematically to

Geological sketch map of the Port
Coldwell Alkalic Rock Complex.



Geology modified from Walker et al., 1993a,b.

Figure 19:

LEGEND

PRECAMBRIAN COLDWELL ALKALIC COMPLEX

- Amphibole Quartz Syenite
- Amphibole Natrolite-Nepheline Syenite
- Nepheline Syenite
- Amphibole Syenite
- Iron-Rich Augite Syenite
- Recrystallized Amphibole Quartz Syenite
- Alkaline Gabbro
- Gabbro
- Mafic Volcanic, Subvolcanic and Hypabyssal Intrusive Rocks

ARCHEAN

FELSIC TO INTERMEDIATE PLUTONIC ROCKS

- Massive to Foliated Granitoid Rocks
- Metasedimentary Rocks
- Intermediate to Felsic Metavolcanic Rocks
- Mafic Metavolcanic Rocks

the southwest along the trend of the Trans Superior Tectonic Zone (Figure 19). The oldest ring complex is centered north of Craddock Cove, the next north of the town of Port Coldwell and the youngest is centered on Pic Island (Currie, 1980). Mitchell and Platt (1994) classified these centres as follows: Centre 1 - Saturated alkaline rocks with oversaturated residue; Centre 2 - Miassicitic alkaline rocks with undersaturated residue and Centre 3 - Alkaline rocks with oversaturated residue.

The oldest rocks of Centre 1 consist dominantly of an olivine gabbro outer ring and a ferroaugite centre. The gabbros have been studied as a consequence of their sulfide content and are described below under economic geology. Shaw (1994) subdivided the eastern gabbro into four intrusions; Gabbronorite, Two Duck Lake Intrusion, Layered Gabbro and Malpas Lake Intrusion. The most primitive gabbro is the Two Duck Lake Intrusion which is equivalent to a high-Al basalt in composition (Shaw, 1994). The ferroaugite syenites have been classified as laurvikite by Puskas (1967) and have undergone detailed study by Jago (1980), McGill (1980), Mitchell and Platt (1978) and Whittaker (1976). The detailed composition of the amphiboles in centres 1 and 3 have been summarized by Mitchell, (1990 p.146-149). The augite syenite contains minor amounts of fayalite olivine (Fa₈₃₋₉₈) (Whittaker, 1976; Mitchell and Platt, 1978). The ferroaugite syenite was highly evolved at the time of intrusion as revealed by fayalitic olivine, diopsidic-hedenbergite and ferro-hastingsitic hornblende (Mitchell and Platt, 1978). The REE mineralogy of the syenites and associated pegmatites was studied by Nicol (1990).

The centre 2 rocks consist dominantly of an outer ring of alkalic biotite gabbro and a inner core of nepheline syenite. The syenite has been studied by Whittaker (1976), McGill (1980), Clark (1983) and Mitchell and Platt (1982). Mitchell and Platt (1982) interpret the nepheline syenite to be the product of fractional crystallization of a basaltic magma.

Centre 3 syenites have been studied in detail by Jago (1980), Lukosius-Sanders (1988) and Mitchell et al. (1993). The REE and rare-metal minerals associated with centre 3 were examined by Nicol (1990) who documented the concentration of REE and rare metals in the residual liquids of these syenitic magmas. Mitchell et al. have interpreted the centre 3 syenites to be the differentiation products of more than one batch of mantle-derived basalt magma. These syenites now represent the product of multiple intrusion, contamination and brecciation giving rise to complex field and petrologic relationships (Mitchell et al. 1993).

Lamprophyres

Lamprophyres are abundant within the centre 1 rocks, particularly in the western part of the complex. They have been studied by Aubut(1977), Evans(1984), Laderoute(1987), Mitchell et al. (1991) and those external to the complex at McKellar Harbour by Platt and Mitchell(1979, 1982a,) Platt et al. 1983). The lamprophyres have a variety of strikes but the dominant trend is approximately east-west with greatly subordinate NW and NE trends. The lamprophyres are generally less than 1.0 m in width and are interpreted to be dominantly related to centre 2 magmatism(Mitchell et al., 1991). Mitchell and Platt(1994) summarized classification of the lamprophyres and one alkalic basalt dike in their order of emplacement as follows:

1. MAFIC OCELLAR LAMPROPHYRES of the camptonitic variety, with or without carbonate-rich ocelli, are widespread. These rocks contain phenocrysts of olivine, aluminous pyroxene, kaersutite and titanian ferropargasite set in a matrix of magnesian hastingsite, augite, plagioclase, biotite, magnetite, sphene and minor nepheline. The ocelli are commonly concentrated by flow differentiation in the centers of the dikes. In addition to calcite, the ocelli contain scapolite, epidote and fluorite.
2. QUARTZ-BEARING MAFIC LAMPROPHYRES which are camptonites, characterized by phenocrysts of quartz with a hexagonal habit. The phenocrysts are typically resorbed and mantled by clinopyroxene. Carbonate ocelli are commonly present. The principal exposures of these dikes are to be found within the gabbros and ferroaugite syenites located on the lakeshore to the west of the Little Pic River(see Aubut, 1977).
3. SANNAITE-TYPE LAMPROPHYRES characterized by the presence of euhedral phenocrysts of aluminian and chromian diopside. These pyroxenes exhibit complex zonation and mantling due to the effects of magma mixing. The groundmass is composed of ferroan paragasite, aluminian diopside, biotite, albitized plagioclase and epidotized alkali feldspar.
4. MONCHIQUITIC-TYPE LAMPROPHYRES with silicate and carbonate ocelli. These rocks are characterized by the presence of brown titaniferous ferro-paragasite phenocrysts set in a matrix of devitrified glass. The silicate ocelli consist of paragasitic amphibole, alkali feldspar and analcime. The monchiquites are considered to be heteromorphs of the camptonites(Mitchell et al. 1991).

5. FELDSPAR GLOMEROPORPHYRY AND ALKALI BASALT DIKES. These rocks have not been studied extensively but may be representative of contemporaneous basic magmas.

6. ANALCIME TINGUAITE (heronites). Porphyritic salic dikes commonly of an orange or reddish-brown colour. Phenocrysts of analcime, alkali feldspar and nepheline can be found, set in a fine grained matrix of altered alkali feldspars and acmitic pyroxene. The least altered occurrences of this suite are found in the country rocks close to the eastern margin of the complex and along railroad tracks in the Heron Bay region. Modern mineralogical studies of these rocks have not been undertaken.

The lamprophyres are common in centre 1 rocks less common in centre 2 and rare in centre 3 (Mitchell et al., 1991; Mitchell and Platt, 1994). Mitchell et al. (1991) interpreted the lamprophyres to be the product of centre 2 magmatism and their absence within a given lithologic unit may be used as one criterion for separating centre 3 rocks from centres 1 and 2. The Marathon lamprophyre dikes dated as 1.65 ± 0.12 Ga by the Rb-Sr technique (Platt and Mitchell, 1983) and the Slate Island lamprophyre dated as 282 ± 11 Ma to 310 ± 18 Ma by the K-Ar technique (Sage, 1991) have been redated by Heaman (personal communication, 1994) as $1145 \pm 15/-10$ Ma and 1141 ± 9 Ma respectively using the U-Pb technique on perovskite. The Marathon dikes and the Slate Island dike intrude the Archean supracrustal assemblage west of the Port Coldwell Complex and may represent a pre-Port Coldwell Complex alkaline event since the Complex has been dated by the same method at 1108 ± 1 Ma (Heaman and Machado, 1992).

Diatremes

On the west coast of the Coldwell peninsula, 4.0 km south of the Neys Provincial Park headquarters, is a diatreme structure described in detail by Balint (1977). This diatreme is approximately 75 x 240 m and contains rounded to angular clasts of Port Coldwell Complex rocks. The long axis of the intrusion strikes north of east and it appears to have a vertical dip (Sage, 1982). The rounding of the clasts suggests turbulent flow in a high velocity gas-solid system (Balint, 1977, Sage, 1982). Sage (1982) considered the diatreme to be a high-level feature similar to those found in porphyry copper systems exposed at high structural levels. The origin of this diatreme is not comparable to the Dead Horse Creek and McKellar Creek diatremes on the east flank of the Port Coldwell Complex which are likely the high level expression of mantle-derived carbonatite magmatism (Sage, 1982).

Hornfelsed Cap Rocks

In the Wolf Camp Lake area and at a number of sites on the north flank of the Port Coldwell Complex subhorizontal massive to amygdaloidal flow rocks are exposed (Walker 1993a,b). These volcanic remnants have been preserved by ring fracturing and down-dropping of the former volcanic edifice above the intrusion (Sage, 1986). The basaltic rocks vary from fresh andesine-oligoclase basalt, hornfelsed basalt to metasomatized basalt in flows up to 5 m thick (Mitchell and Platt, 1994). Nicol (1990) has studied the flows and considers them to be of tholeiitic lineage. The hornfelsed flows contain some orthopyroxene and the amygdule fillings were metamorphosed to a greenish calc-silicate mineral assemblage.

Xenoliths

In addition to the study of xenoliths by Nicol (1990), Mitchell and Platt (1979) examined feldspar porphyry xenoliths in nepheline syenite displaying nepheline-plagioclase intergrowths and determined that the nepheline has likely resulted from a Na-K cation exchange. This metasomatism caused desilication of the andesine phenocrysts but did not form nepheline in the groundmass (Mitchell and Platt, 1979). The metasomatism developed alkali feldspars of a wide range of compositions (Mitchell and Platt, 1979).

Isotope Studies

Excluding isotopic studies directed toward age determinations, Heaman and Machado (1992) examined Sr, Nd and Pb isotopes and determined that Midcontinent Rift magmas have a uniform Nd isotopic composition. Heaman and Machado (1992) compared 1.1 Ga carbonatites with Port Coldwell Complex data and found that carbonatites have lower Rb/Sr and higher U/Pb and Sm/Nd ratios. This means that there were two isotopic reservoirs present at 1.1 Ga; an enriched plume-component represented by most Midcontinent Rift magmatism and a carbonatitic component (Heaman and Machado, 1992). Heaman and Machado (1992) interpreted the presence of nonradiogenic Pb compositions and negative Nd values in silica-saturated Port Coldwell magmas as documenting the interaction of plume-derived mantle melts with low U/Pb granulite-facies lower crust.

Geochronology

Currie (1980) provided some K-Ar isotopic ages and Rb/Sr isotopic ages have been provided by Platt and Mitchell (1982b) and Bell and Blenkinsop (1980). The means of determining the Rb/Sr isochron of Platt and Mitchell (1982b)

was discussed by Blenkinsop and Bell (1983) and Platt and Mitchell(1984). Uranium-lead isotopic ages were first undertaken by Turek et al. (1985) and the results discussed by Thorpe(1986) and Turek et al(1986). Heaman and Machado(1992) reported a U-Pb isotopic age of 1108 $\pm$ 1 Ma which is now accepted as the most accurate age. Heaman and Machado(1992) could not distinguish the three intrusive centres on the basis of isotopic ages. The distinction between these intrusive centres rests on field observations.

Local Structure

The Port Coldwell Complex occurs at the Proterozoic-Archean contact just north of the intersection of the Michipicoten Island Fault and the Big Bay-Ashburton Bay Fault(Sage, 1978) or Thiel Fault(Klasner et al., 1982). The Michipicoten Island Fault parallels and may mark the limits of the Proterozoic basin. The Port Coldwell Complex lies within the Trans Superior Tectonic Zone(Klasner et al., 1982). Within the Port Coldwell Complex mafic banding, trachytoidal textures and xenolith preferred elongation may locally be used to outline the configuration of parts of the various centres of intrusion. These planar features have been used by Currie(1980) to speculate on the centres of emplacement for the three ring complexes that make up the Port Coldwell Complex. The three intrusive centres young in a southwest direction toward the Slate Islands(Sage, 1991) and along the Big Bay-Ashburton Bay Fault trend. The successive emplacement of superimposed ring complexes accompanied by repeated faulting and stoping, multiple intrusion, metamorphism and metasomatic activity has created very complex petrographic, petrologic and field relationships. The presence of abundant xenoliths in some lithologies and large roof pendants indicate that the complex is barely unroofed and is thus exposed at a very high structural level.

Lilley(1964) proposed that the Port Coldwell Complex is a funnel-shaped intrusion and Puskas(1967) has proposed that it is a lopolith. Currie(1980) considers the complex to consist of an interlocking group of cone sheets and ring dikes. The arcuate outer gabbro unit of centre 1 may be interpreted as a ring dike as may the alkaline biotite gabbro of centre 2. Arcuate faults and large down-dropped roof pendants are consistent with ring fracturing and subsidence, characteristic of caldera emplacement(Sage, 1986, Mitchell and Platt, 1982b, 1994, Walker, 1993)(Figures 20, 21). Each intrusive centre has likely undergone a similar series of intrusive events. Mitchell and Platt(1994) consider intrusive centres 2 and 3 to be exposed at a higher level than centre one due to the abundance of xenoliths; however, the large xenoliths or roof pendants near Wolf Camp

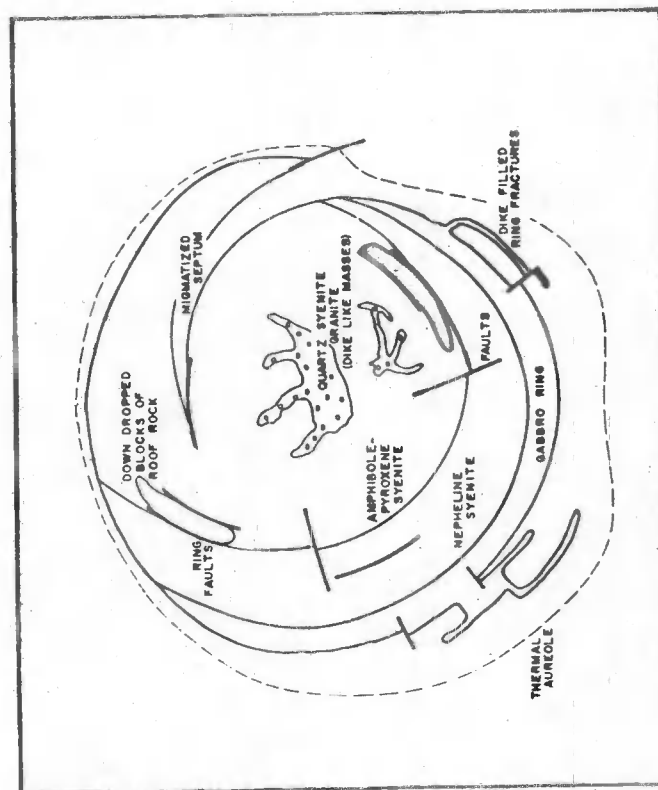


Figure 20: Idealized map of rock type distribution in a simple ring complex (from Sage, 1986).

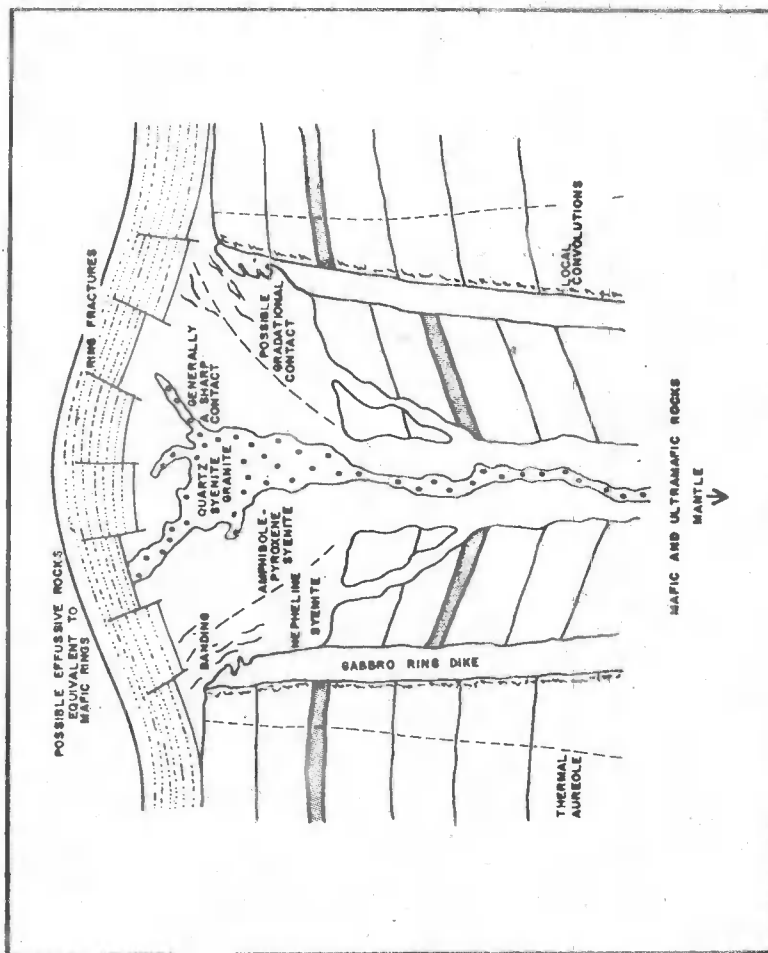


Figure 21: Idealized vertical section displaying rock type distribution in a simple alkalic ring complex (from Sage, 1986).

Lake and in the northern parts of the complex are largely centered on centre 1. The difference in the level of structural exposure between the three centres is not likely to be very great. Sage(1986) suggested that the southern part of the Port Coldwell Complex may be exposed at a slightly deeper level than that of the north due to the abundance of xenoliths and rook pendants in the northern part of the intrusion.

Geophysics and LANDSAT Imagery

The Port Coldwell Complex is covered by LANDSAT imagery(Canada Centre for Remote Sensing, 1987) and the large Port Coldwell Complex is easily recognized. Local faults and linears are easy to trace and often can be extended beyond the limits of the complex.

Meru(1965) studied the angle of emergent seismic P waves at Marathon, Ontario and interpreted his data to indicate that the Moho below Marathon dips southwest under the Lake Superior Basin at 4 degrees. He also concluded that the crust was much thicker to the southwest of Marathon than to the northeast(Meru, 1965).

Lilley(1964) examined the aeromagnetic expression of the Port Coldwell Complex and observed intense negative anomalies over the eastern gabbros of centre (Figure 22). The emplacement of the Port Coldwell Complex at 1108 ± 1 ma(Heaman and Machado, 1992), therefore, took place just prior to a magnetic reversal that occurred between 1096.2 ± 1.8 Ma and 1097.6 Ma(Davis and Paces, 1990). The age of reversal may closely approximate the age of reversal for the Osler Group(Davis and Paces, 1990, Paces and Miller, 1993). Isomagnetic contouring of the aeromagnetic data clearly distinguishes the Port Coldwell Complex from the surrounding Archean rocks and reveals a prominent circular pattern(Figure 22).

Currie(1980) cites gravity data as indicating that the Port Coldwell Complex has a mafic component at depth(Figure 23). Since mostly light syenitic rocks are exposed at surface postulation of mafic rocks beneath the syenites is required to explain the positive gravity anomaly. Mitchell et al. (1983) modeled the gravity data and suggested that the felsic rocks represent a 3 to 5 km-thick layer over a differentiated basic intrusion consisting of 3 to 5 km of gabbro underlain by peridotite or pyroxenite. The presence of large volumes of mafic rocks at depth supports the concept that some of the Port Coldwell Complex felsic rocks are differentiates (Figure 23). During Keweenawan rifting upper mantle-derived mafic magmas migrated from the southwest to the northeast and came to rest at Port Coldwell(Mitchell et al., 1983).

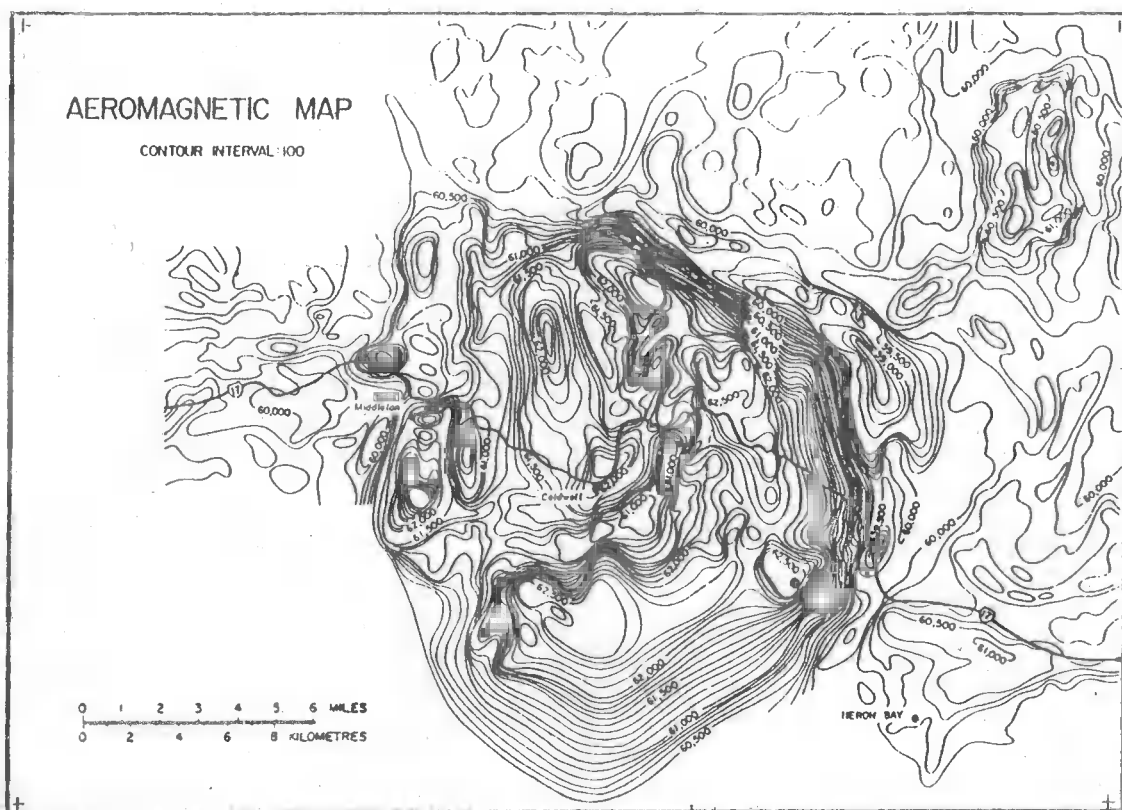


Figure 22: Aeromagnetic map of the Port Coldwell Alkalic Rock Complex (ODM-GSC, 1963a,b,c,d).

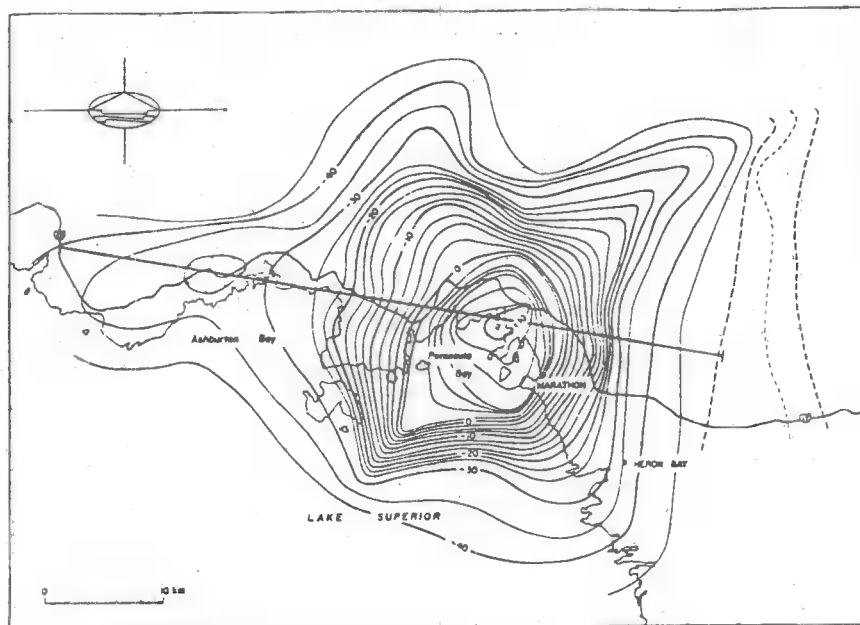


Figure 23: Bouguer anomaly map of the Port Coldwell Alkalic Rock Complex. The solid straight line is the location of a gravity profile (from Mitchell et al., 1983).

The Port Coldwell Complex and surrounding Archean terrane has been covered by a reconnaissance airborne gamma-ray spectrometer survey (Graham and Bonham-Carter, 1993). The concentration of radioelements in the Port Coldwell Complex is high in contrast to the enclosing Archean supracrustal rocks. Graham and Bonham-Carter (1993) indicated that in general each lithologic unit may be distinguished on the basis of its gross radiometric signature and that the three intrusive centres display minor correlatable differences. The Port Coldwell Complex is also covered by side-scanning radar imagery completed by the Canada Centre for Remote Sensing and these data were merged by Graham and Bonham-Carter (1993) in their study of the radiometric data. The radar imagery reflects the rugged topography of the Port Coldwell Complex area.

Economic Geology

The Port Coldwell Complex had several small quarries developed for building stone in the ferroaugite syenite close to the CPR tracks and the town of Marathon in the early 1930's (Puskas, 1967). These efforts failed to be commercial and the efforts were soon abandoned. The Port Coldwell Complex has repeatedly been evaluated as a potential source for building stone due to the pronounced blue to gold schiller texture of the feldspars in the ferroaugite syenite. In the late 1930's efforts were made to evaluate the titaniferous magnetite showings in the peripheral gabbros (Puskas, 1967). Other than surface trenching and pitting little work was completed and no tonnage was outlined. The titaniferous magnetite at Stop 15 occurs in the area where most efforts were expended in search of titaniferous magnetite deposits.

In the 1950's porphyritic syenite dikes were prospected for their Nb and Zr content. The principal areas of investigation were Port Munro near the former siding of Angler on the CPR and north of west from the west end of Craddock Lake within the interior. None of these efforts proved sufficient grade or tonnage to be of economic interest. One of these dikes was tested for U and Th without success.

In the 1960's the nepheline syenites on the Neys Peninsula and Pic Island were evaluated as a possible source of ceramic grade nepheline. This work proved unsuccessful because the impurities could not be removed from the nepheline. The poikilitic nepheline commonly contains tiny inclusions of ferromagnesian minerals which cannot be removed in milling, thus the desired ceramic grade purity could not be obtained.

The Port Coldwell Complex offers the best potential as a source for Cu-Ni-PGE. Exploration of the sulphide mineral assemblages in the peripheral gabbros of centre 1 on both the western and eastern flanks of the intrusion started in the 1950's and peaked in the 1960's. Most of this work was done in the eastern gabbro north of Marathon by Anaconda American Brass Ltd. on a deposit now known as the Two Duck Lake Occurrence. Watkinson and Ohnenstetter (1992) indicated that the Two Duck Lake Occurrence occurs in a separate gabbroic intrusion within the outer ring of centre 1 gabbro. The mineralization occurs in coarse gabbro and pegmatite and is interpreted to be the product of interaction of a magmatic mineral assemblage with fluids of mixed magmatic origin and fluid generated by the breakdown of abundant xenoliths, principally felsic metavolcanic rocks (Watkinson and Ohnenstetter, 1992). Good and Crocket (1994b) have proposed that the Two Duck Lake Occurrence was intruded as a plagioclase-rich crystal mush which had fractionated at depth and contained sulphide droplets. The sulphide mineral assemblage precipitated after rapid solidification of the crystal mush with little Cu or PGE migration (Good and Crocket, 1994b). Good and Crocket (1994b) suggested that the origin of the mineralization is analogous to that of the Duluth Complex. Shaw (1994) has interpreted the Two Duck Lake Intrusion to be the most primitive unit of 4 subunits within the eastern gabbro and similar to high-Al basalt. This magma underwent fractionation in an open system with removal and recharge in 3 - 10 cycles in an expanding magma chamber (Shaw, 1994). The reader should refer to Shaw (1994) for details. In the 1980's interest in the Two Duck Lake deposit was revived by Fleck Resources Ltd. who have outlined 37,000,000 tons grading 0.31 % Cu, 0.04 % Ni, 251,000 oz Pt, 1,001,000 oz Pd, 84,000 oz Au and 43,000 oz Rh (Canadian Mines Handbook, 1993-1994, p. 145) (See Stops 25, 27). Platinum-group-element mineralization is also present in association with a large block of gabbro within the north central part of the Port Coldwell Complex. This is known as the Geordie Lake Occurrence and grade and tonnage have not been determined. The Geordie Lake Occurrence has been interpreted as a late-stage hydrothermal product of the final crystallization of a plagioclase-rich intrusion of gabbroic magma (Good and Crocket, 1994a). The dominant sulphides are chalcopyrite, pyrrhotite, pyrite, cubanite and rare pentlandite. Table 2 lists other minerals found in studies completed over the years in these mineralized rocks. The reader should refer to the references for details and to Walker (1993c) for another brief discussion of the economic geology.

Table 2: Mineralogy of the mineralized gabbros of the Port Coldwell Complex

Mineral	Reference
Altaite(PbTe)	2
Atokite(Pd ₃ Sn)	1,2
Bornite(Cu ₅ FeS ₄)	5
Electrum(Au,Ag)	2
Guanglinite(Pd ₃ As)	1,2
Galena(PbS)	2
Hessite(Ag ₂ Te)	4
Hollingworthite(RhAsS)	1,2
Irasite(IrAsS)	1,2
Kotulskite(PdTe)	1,2
Bi and Pb-bearing Kotulskite	
Mackinawite((Fe,Ni) ₉ S ₈)	5
Majakite(NiPdAs)	2
Merenskyite, solid solutions, (PbTe ₂)	1,2
Mertieite-II(Pd ₈ Sb ₃)	1,2
Palladoarsenide(Pd ₂ As)	1,2
Ag-Pentlandite(Fe,Ni) ₉ S ₈ with Ag	2
Platarsite(PtAsS)	1,2
Rhenium sulphide	4
Sopcheite, solid solutions, (Pd ₂ Ag ₄ Te ₄)	1,2
Sperrylite(PtAs ₂)	1,2
Telargpalite((Pd,Ag) ₃ Te(?))	2
Troilite(FeS)	5
Unnamed (Pd ₅ As ₂), Pd-bearing nickeline	1,2
Unnamed Pd _{1.6} NiAs _{1.5}	3
Zvyaginskite(Pd ₃ Pb)	1,2

1) Ohnenstetter et al., 1991; 2) Watkinson and Ohnenstetter, 1992; 3) Mulja and Mitchell(1990); Mitchell et al., 1989; 5) Lum, 1973

Field Trip Log

At 2.0 km from the Dead Horse Creek access road

Stop 12: Molybdenite along fractures in hornfelsed wacke. Molybdenite occurs throughout the Port Coldwell Complex as tiny isolated grains and rosettes along joint planes. No concentration of molybdenite has been found which warrants further testing. The outcrop is locally cut by gabbro and syenite. At this site the molybdenite occurs as tiny grains and rosettes up to 2 mm along joint planes which may display some alteration to yellow ferrimolybdite. The syenite dikes contains abundant xenoliths of both wall rock and phases of the complex. East of the exposure, the road cut reveals various phases of syenite with abundant xenoliths. The general trend of the bedding in the wacke is 050 with a

steep southerly dip, but extreme variation in bedding attitudes occur.

At 2.9 km

Stop 13: Optional: Rheomorphic breccia. This stop requires at least a 20 minute walk over a skid trail that is overgrown and often wet. Park the vehicles on the north side of the curve in the highway. Be careful and be sure that you can get on and off the highway safely and that the shoulder of the road is not soft. The skid trail was made for power line construction and leads off into the bush from the top of the outcrop on the south side of the highway.

At approximately 795 m the power line tower to the east of the trail is on a syenite outcrop containing very fine-grained gabbro xenoliths with concentric rings of alteration. The reaction rings are 0.5 to 1.0 cm wide and weather high compared to the core of the xenolith. The xenoliths are in coarse-grained syenite cut by a fine-grained syenitic aplite up to 20 cm wide that trends 165. Between this tower and Highway 17 to the east, the outcrops are dominantly layered gabbro. The very fine-grained gabbro is the oldest gabbroic phase and often has a knobby weathered surface which may be observed in outcrops on the trail beneath the power line.

Continue walking to approximately 1270 m . The skid trail takes a wide loop to the south and comes out at the base of a tower on the power line east of the earlier stop. The exposure beneath the tower and a few metres to the east is one of the best examples of rheomorphic breccia observable on the Port Coldwell Complex. Angular to rounded clasts of metasedimentary rock occur in a fine grained remobilized anatectic matrix. Some clasts are up to 0.3 m and some are rusty. The clasts have contorted banding in a matrix displaying a fluxion structure suggesting the fragments were ductile in a flowing medium. The xenoliths are randomly oriented and angular to rounded fragments of vein quartz are locally present. From the exposure beneath and just east of the tower walk west along the skid road beneath the power line and note the rapid change from rheomorphic breccia into less and less deformed metasedimentary rocks.

At 60 to 70 m convoluted bedding as well as segmented quartz veins are easily recognized. There are good examples of boudinaged quartz veins at this site directly below the power line. At approximately 100 m good compositional layering (bedding) displays much less convolution but the quartz veins are still boudinaged. The bedding is trending approximately 295 with a 75 south dip and there is open kinking of the bedding. The kink bands trend 345 and dip

vertically. The plunge of the fold axis in the kink band is approximately 35 NW.

At approximately 165 m the bedded metasedimentary rocks are much less contorted and bedding trending 300 and dipping 70 north is easy to recognize. Some open folding in the metasedimentary rocks and some minor faulting trending 345 dipping 75 north offset the bedding by as much as 20 - 25 cm. The fault trend is parallel to the trend of the kink bands.

To observe even less deformed rocks continue west. After completion of your observations return to the vehicles.

At 3.2 km

Stop 14: Mineralized western gabbro. Park at the entrance to the trail to Middleton siding on the south side of the road. The trail is a good walking trail which leads to the CPR tracks.

At approximately 925 m an old trench in rusty, medium-grained gabbro occurs on the west side of the trail. The gabbro is massive and locally contains 0.5 to 1.0 % chalcopyrite and pyrrhotite. This exposure is typical of mineralized western gabbro and differs from the mineralized eastern gabbro in the lack of well developed pegmatitic textures. This trench was probably made in the late 1950's. Wilkinson(1983) reported that the olivine from this site is Fo 49 to 51 and that plagioclase is An 47 to 52. The western gabbros are less evolved than the eastern gabbros which have olivine compositions of Fo 43 to 46 and plagioclase compositions of An 51 to 57 (Wilkinson, 1983).

At 3.3 km

Remain parked at stop 14 and walk east along Highway 17 to the large outcrop on the north side of the road. This is stop 10 of Mitchell and Platt(1994)

Stop 15: This site is noted for its spectacular "hydronephelinite"(natrolite) syenite pegmatite enclosing medium grained gabbro xenoliths. The outcrop is cut by a late lamprophyre dike. The medium-grained angular gabbro xenoliths up to 1.0 m in size have a wide 1 to 2 cm dark reaction rim with the enclosing syenite pegmatite. The natrolite syenite pegmatites contain patches of red to orange-red natrolite up to 10 - 15 cm, perthitic feldspars up to 30 cm and dark green to black amphibole up to 20 to 25 cm. Mitchell and Platt(1994) reported the presence of pleochroic clinopyroxene, zircon, titanite and biotite.

The fine-grained lamprophyre that cuts the natrolite pegmatite trends 040, dips 55 north and is approximately 40 cm wide. Mitchell and Platt(1994) indicated that most lamprophyres in this area may be classified as camptonites.

Continue east along the face of the outcrop from the natrolite pegmatites. The rock is a fine- to medium-grained nepheline syenite with nepheline indicated by the chalky weathering. Occasionally a grain of reddish orange "hydronephelinite"(natrolite) is recognizable.

Continue east along the face of the outcrop to almost the end of the exposure to examine various syenite and syenite pegmatite units. Near the end of the outcrop medium-grained titaniferous magnetite with minor clinopyroxene, plagioclase and apatite is exposed in the road cut. This material has been prospected in the past as a source for titaniferous magnetite. The best outcropping is approximately 150 m east of the lamprophyre dike that cuts the natrolite pegmatite.

The various rock types exposed in this road cut have been assigned to various intrusive centres by Mitchell and Platt(1994). The gabbro, and ferroaugite syenites are related to centre 1 activity, amphibole syenites to centre 3 and the natrolite pegmatites to centre 2(Mitchell and Platt, 1994). McGill(1980) has examined the natrolite pegmatites and titaniferous magnetite accumulations in the gabbro.

Return to vehicles

At 5.5 km

Stop 16: Intrusive breccia on the east side of the Little Pic River. Park at the large turn out at the east end of the bridge and walk along the highway to the first outcrop on the south side of the road.

Angular blocks of fine- to medium-grained equigranular oligoclase gabbro occur in a matrix of medium-grained pink quartz syenite. The gabbro appears to be shattered and veined by the syenite and some of the fragments may visually be refitted as one would a jig saw puzzle. The quartz syenite filling in around the fragments locally contains miarolitic cavities up to several centimetres in width which may contain euhedral quartz and feldspar crystals. White calcite may partially fill some of these cavities. This oligoclase gabbro occurs only in this area of the Little Pic River and up to the Little Pic River lookout(Stop 17) where Mitchell and Platt(1994) refer to it as an oligoclase basalt. Volcanic textures have not been observed in this unit. The Little Pic River occupies a strong lineament which may be traced on airphotos to the west flank of the Killala Lake Complex and north. The lineament is likely a fault zone

but offset cannot be determined at this location. The Little Pic River lineament can be interpreted to be the onshore extension of a deep water-filled linear located southwest of the Slate Islands (Sage 1991) which probably represents the onshore expression of faulting related to the Thiel (Klasner et al. 1982) or Big Bay-Ashburton Bay Fault (Sage, 1978). Parts of the breccia zone have been studied by Laws (1983).

At the extreme western end of the intrusive breccia, weakly trachytoidal ferroaugite syenite occurs in outcrop. The spectacular cliff faces observed on the west side of the Little Pic River consist of ferroaugite syenite.

At 6.3 km

Stop 17: Little Pic River turn out. This is a frequent stop for visitors to the Port Coldwell Complex as it provides safe parking of the vehicles. Be careful and watch for vehicles while you examine exposures along Highway 17. This is stop 17 of Mitchell and Platt (1994).

Walk from the parking area to Highway 17 and examine the intrusive breccias and xenolith-rich syenite. Oligoclase gabbro veined with quartz syenite occurs along the south side of the road and various syenitic rocks with xenoliths occur on the north side. The syenites are pyroxene-amphibole syenites containing xenoliths of alkali gabbro, alkali diorite and various syenites. The syenitic xenoliths vary from coarse-grained equigranular blocks to porphyritic phases. The ratio of alkali gabbro to diorite increases to the east toward the entrance to the lookout. The alkali gabbro to diorite is cut by grey nepheline syenite with some reddish orange "hydronephelinite" toward the entrance to the lookout. The amphibole tends to be somewhat acicular in the nepheline-rich phases and locally they have somewhat gradational contacts with the alkali gabbro to diorite. Pegmatitic dikes of quartz syenite outcrop opposite the entrance to the lookout.

East of the turn off to the lookout the medium- to coarse-grained amphibole-pyroxene syenite is inequigranular seriate and the number of xenoliths has greatly diminished.

The syenites at this location have been related to centre 3 intrusive activity and Lukosius-Sanders (1988) and Mitchell et al. (1993) described these syenites in detail. In the order of emplacement Mitchell and Platt (1994) list the syenites as follows: magnesio-hornblende syenite, contaminated ferro-edenite syenite, ferro-edenite syenite and quartz syenite. Sannaite lamprophyre and ocellar camptonitic lamprophyre are reported to occur in the area (Mitchell and Platt, 1994). Lukosius-Sanders (1988)

classified the syenites as miaskitic, metaluminous enriched in U, Th, REE and Zr due to the presence of zircon, chevkinite $\{(Ca, Ce, Th)_4(Fe, Mg)_2(Ti, Fe)_3Si_4O_{22}\}$, REE-carbonates, Nb-rutile and aeschylite $\{(Y, Ca, Fe, Th)(Ti, Nb)_2(O, OH)_6\}$. The centre 3 syenites have affinities to A-type granites and have been interpreted to be the result of fractional crystallization of mantle-derived basaltic magma (Lukosius-Sanders, 1988; Mitchell et al. 1993).

At 9.8 km

Stop 18(Optional): Turn off to Neys Provincial Park. This will be the site of an optional stop. Access to this area may not be possible in the early spring since the park is closed to visitors. This stop has been described as stop 15 by Mitchell and Platt(1994).

Drive approximately 2.8 km along park access road and park at Neys Park headquarters. Walk east and south along the beach for approximately 400 m to the first large outcrop on the shore of Lake Superior.

SAMPLE COLLECTING AT THIS SITE IS PROHIBITED

Medium- to coarse-grained biotite gabbro of centre 2 outcrops on the eastern part of the exposure and nepheline syenites of centre 2 are exposed on the western part of the outcrop. The texture and composition of the nepheline syenites are heterogeneous and possibly represent magma mixing and hybridization, process not yet studied in detail within the Port Coldwell Complex. One may continue walking south along the coast for several 100 m and observe nepheline syenites with well developed layering and structures reminiscent of "soft-sediment" deformation. The layered syenites are divided into large blocks by a later intrusion of nepheline syenite. The earlier syenites were subjected to current action and "sedimentation" and then subsequently forcibly intruded by younger syenites of similar composition. The mafic mineral is hastingsite (Mitchell and Platt, 1994). The general trend of the banding is NNW with a highly variable dip.

Return to Highway 17 and continue east

Stop 19(Optional):

At 14.6 km turn off to former townsite of Port Coldwell. Distance remains from the Dead Horse Creek access road. Drive approximately 0.2 km and park near the CPR tracks.

THIS IS A VERY DANGEROUS LOCATION. THE SHAPE OF THE HILLS MODIFY THE SOUND OF APPROACHING TRAINS AND ONE MAY NOT HEAR A TRAIN UNTIL IT IS EXTREMELY CLOSE. BE ALERT AT ALL TIMES.

Amphibole-nepheline syenites of centre 2. Walk east for approximately 700 m along the CPR tracks to the high cliff faces along the west side of the railroad opposite Echo Lake. These cliffs are composed of nepheline syenite and there is abundant loose rock above so be careful. The nepheline syenite is generally a massive equigranular rock with local reddish orange spots of "hydronephelinite". The rock surface is highly stained and fresh surfaces may only be obtained from breaking open pieces that have fallen from the cliff.

Continuing beyond this cliff face one may examine Stop 8 of Mitchell and Platt(1994). This stop is on the shore of Redsucker Cove 1.6 km from Port Coldwell village. At a concrete embankment leave the tracks and go down to the lake shore to examine diorite, nepheline syenite veins and a analcite trachyte dike(Mitchell and Platt, 1994). To the east various varieties of nepheline syenite may be observed, and to the west, breccias of gabbro, diorite and nepheline syenite may be examined. A number of camptonitic lamprophyres and one sannaitic lamprophyre occur at this site(Mitchell and Platt, 1994). For details of this stop refer to Mitchell and Platt, (1994).

Return to the vehicles and continue east along Highway 17

At 16.1 km highway 17 crosses Mink Creek. There is a waterfall in syenite approximately 100 m south of the highway. You may park on the small logging access road leading north from Highway 17 west of Mink Creek and take a short trail south of the highway to the falls.

At 19.2 km

Stop 20: Biotite gabbro intruded by various types of nepheline syenite.

This stop is at a broad curve in highway 17 and one should be very CAREFUL OF VEHICLES.

Starting at approximately 18.8 km outcrops on the east side of the highway of grey to buff pyroxene amphibole syenite contain orange fluorescing hackmanite, a variety of sodalite. This mineral can only readily be seen with a UV lamp. The syenite contains numerous mafic xenoliths up to 25-30 cm in maximum length. The xenoliths are subangular to subrounded and the mafic minerals within the syenite tend to

occur in clots. The larger xenoliths may have feldspar phenocrysts or porphyroblasts up to 1.0 cm. The phenocrysts or porphyroblasts display a seriate size distribution and comprise up to 5 % of the rock.

At the centre of the curve at 19.2 km an alkalic biotite gabbro is exposed on the north side of the highway. The medium- to coarse-grained gabbro is extensively cut by medium- to coarse-grained syenite and some of the nepheline has been altered to reddish orange "hydronephelinite". There may be two ages of alkalic gabbro with the older coarser grained gabbro displaying dark selvages up to 7 or 8 cm wide and an irregular shape suggesting that it behaved in a more or less ductile manner. The alkalic gabbros have a clotty mafic mineral assemblage. The surface of the coarser grained biotite gabbro is pitted from the weathering of mafic clots. Dikes of nepheline syenite pegmatite occur on the north side of the highway at the inflection in the curve. At this site two ages of nepheline syenite pegmatite of essentially identical composition cut each other. The trends of these dikes are approximately 340 dipping 60 south and 090 dipping 50 north. The dike trending 090 cuts the dike trending 340 and both are on the order of 20 to 30 cm in width. The dikes have been sketched by Puskas(1970). Both dikes are zoned from an amphibole-rich margin to a feldspar-natrolite(hydronephelinite)-rich core. The central parts of these dikes are commonly relatively rich in reddish orange "hydronephelinite". West from the two dikes toward the hackmanite-bearing outcrop, coarse-grained alkalic biotite gabbro is intruded by medium to coarse grained pyroxene, amphibole syenite with traces of nepheline. Both of these rock types are in turn cut by coarse-grained nepheline syenite dikes. Brecciation is so intensive that the outcrop is an igneous breccia. West toward the hackmanite-bearing outcrop the syenite is pink to grey, fine- to medium-grained, inequigranular seriate with clotty assemblages of mafic minerals.

On the south side of the highway coarse-grained alkalic gabbro appears to be cut by medium-grained alkalic gabbro which is in turn intruded by mottled pink to grey, inequigranular seriate amphibole syenite. There is a slight coarsening in texture next to the medium-grained gabbro and a dike of similar material projects from the contact with the medium-grained gabbro through both phases of alkalic gabbro.

At 24.4 km

Stop 21: Hornfelsed roof pendants, Wolf Camp Lake

This is stop 7 of Mitchell and Platt(1994). Hornfelsed rocks overlying the Port Coldwell Complex were recognized early in the mapping of the intrusion by Tuominen(1967) and Puskas(1970) and are likely representative of a former volcanic edifice(Sage, 1986). Mitchell and Platt(1994) and Nicol(1990) have studied these rocks in detail and consider these basalts to be contemporaneous with the Port Coldwell Complex and to represent a tholeiitic lineage. The basalts are fresh, metasomatized and hornfelsed andesine-oligoclase basalt flows estimated to have a thickness of up to 5 m (Nicol, 1990, Mitchell and Platt, 1994).

At this stop amygdaloidal basalt with calc-silicate amygdule fillings is well exposed. The ameboid amygdules are unlike amygdules found in Archean metavolcanic rocks of the region and more typical of those found in Keweenawan flows. The greenish calc-silicate amygdule fillings are interpreted to represent the metamorphic product of former calcite, quartz etc. fillings of the amygdules. Exposures of both massive and amygdular flows may be found on the hills and cliffs southwest of Wolf Camp Lake and for a short distance along Highway 17 east of this stop. Hornfelsed mafic flows are also present in the northern part of the complex (Tuominen,1967) where they have been observed to overlie syenitic rocks of the Port Coldwell Complex. These flows are thought to be the remnants of a once former volcanic edifice that have been preserved due to ring fracturing and caldera subsidence.

At 33.9 km Turn off to the town of Marathon

At 35.6 km

Stop 22: Layered gabbro. This is stop 4 of Mitchell and Platt(1994). The gabbros of the eastern contact area are well exposed at this site and considered to represent centre 1 intrusive activity. This gabbro is part of an outer ring that encompasses the eastern and most of the northern parts of the Port Coldwell Complex separating the external Archean rocks from the internal syenitic rocks of the intrusion. These gabbros are reversely magnetized and form prominent lows on aeromagnetic maps(Lilley, 1964). This stop is close to the ferrosyenite-gabbro contact and there is a lack of intermediate compositions between these two rock types implying two distinctly different magmas. Relationships between these two magma types are unknown. The ferroaugite syenite is best observed at the western end of the outcrop where pegmatitic phases contain large miarolitic cavities up to 2 m lined with euhedral feldspar, quartz and amphibole. McLaughlin, (1990) reports the presence of fluorocarbonates(bastnaesite, parisite, synchysite) Nb-

rutile, Nb-bearing ilmenite, columbite, pyrochlore and zircon in these pegmatites.

The best exposure for viewing the layered gabbro is in the east central part of the road cut at the top of the outcrop. Small pegmatitic dikes of quartz syenite cut the outcrop. The rhythmic layering is due to variation in the plagioclase-mafic mineral content and individual layers of plagioclase-rich rock are 0.5 to 15 cm thick and mafic mineral-rich layers are 0.2 to 2.0 cm thick. The banding strikes roughly parallel to the road and dips approximately 45 west. Mitchell and Platt(1994) believe this rhythmic-layered gabbro is a rafted block. The trend of banding in this area is somewhat erratic, dips are consistently west but variable. Outcrops west of this stop on the north side of the highway display a less distinct, wispy, streaky banding due to slight variations in mafic-felsic mineral contents. The reader may wish to refer to Shaw(1994) for a detailed petrological study of the eastern gabbro.

At 37.3 km

Stop 23: Eastern gabbro in contact with rheomorphic breccia at the eastern contact.

This is a deeply weathered outcrop of sulphide-bearing gabbro at the contact with Archean wall rocks. The gabbro contains abundant xenoliths of the wall rock and very fine grained gabbro. The gabbro is cut by irregular dikes of quartz syenite pegmatite. The rheomorphic breccia is identical to that observed at the western contact(Stop 13) where the exposure is better. The rheomorphic breccia is characterized by convoluted flow lines and xenoliths in random orientation, some of which appear to have been ductile and partially assimilated. Some xenoliths retain a somewhat angular outline and some display a boudinage structure as they underwent further fragmentation. The contact between the gabbro and wall rocks may be observed but due to its irregular trend its attitude is uncertain. The sulphide-bearing gabbro has been prospected north and south of highway 17 at this point. The sulphides are principally chalcopyrite and pyrrhotite.

Return to the vehicle and return to the intersection of the road to Marathon and Highway 17. From the intersection of the Marathon road with Highway 17 take the following route:

Go 5.3 km from the turn off, cross the CPR tracks and turn right toward Lake Superior.

Go to 5.7 km and pass the hotel on the right, Moose Lodge on the left and turn right on Winston St.

Go to 6.0 km crossing the CPR tracks and continuing to the right.

Go to 6.7 km where the road splits. Take the left hand fork along the shore of Lake Superior.

At 7.2 Km

Stop 24: Ferroaugite syenite. This is a small quarry in ferroaugite syenite on the shore of Lake Superior where coarse grained equigranular specimens of the syenite may be collected. This quarry was probably made in the 1930's and is typical of centre 1 ferroaugite syenite referred to as laurvikite by Puskas(1967).

At 7.8 km

Stop 25: Layered ferroaugite syenite.

Park at the end of the road which is a boat landing and walk north along the shore line for approximately 680 m. There is a partial trail for most of the distance. The broad expanses of ferroaugite syenite exposed along the shoreline display excellent glacial grooving and polishing. Be careful where seepage from the surrounding bush onto the rocks has permitted green algae to grow. This green algae is very slippery. Also be careful after walking through a sandy area and then stepping onto outcrop. Sand on the sole of the boot may act like ball bearings on the glacially polished outcrop surface and you may slip.

At approximately 340 m along this traverse near the bush-outcrop line of demarcation a fine-grained inequigranular porphyritic seriate syenite dike 1.5 m wide cuts the ferroaugite syenite. The dike trends 150 and has a vertical dip. The dike locally contains 10 to 15 % feldspar phenocrysts up to 1.0 cm in long dimension. Similar dikes in the region have been prospected for their niobium(pyrochlore) and zirconium(zircon)content.

Continue north to the banded ferroaugite syenite. The banding is wispy, discontinuous and may bifurcate. The general trend is 070 with a 60 north dip. The banding is due to varying mafic mineral content. The glacial grooving has a general trend of 240 degrees. This is the second best location to study mafic banding in the pyroxene syenites. The best site is east of Marathon on the shore of Lake Superior and involves a long walk(Stop 27).

Return to the core shack of Fleck Resources Ltd. Marathon. This is a blue-grey building on the west side of the road into Marathon.

Stop 26: Mr. J. McGoran, President, Fleck Resources Ltd. will place on display drill core from the Two Duck Lake Gabbro. The PGE-rich sulphides are cubanite, pyrrhotite and chalcopyrite interstitial to the rock forming minerals of the gabbro. The pegmatitic phases of the gabbro are the most sulphide-rich.

End of formal tour.

Optional Stop 27: For the connoisseur of igneous petrology or of the geology of the Port Coldwell Complex, the better part of a day should be spent walking along the shore of Lake Superior east of Marathon to observe analcite tinguaitite dikes cutting the Archean wall rocks, excellent exposures of banded ferroaugite syenite and fine-grained hybrid gabbro. These glacially polished, wave-washed outcrops are excellent to study. Good weather and a reasonably calm Lake Superior are highly desirable. The trail along this shore line starts at the end of Howe St. in Marathon and these outcrops have been described in some detail by Mitchell and Platt(1994) as stop 6.

Optional Stop 28: For the economic geologist a visit to the Cu-Ni-PGE deposits at Two Duck Lake is advisable. This is private property and requires the permission of Fleck Resources Ltd. to visit. The road to these deposits begins opposite the junction of highway 17 and the road to Marathon. The distance to the Cu-Ni-PGE deposits is in excess of 12 km and the road is often impassable even with a 4-wheel drive vehicle.

At this site a sill-like gabbro body referred to as the Two Duck Lake Gabbro intrudes earlier eastern gabbros of the centre 1 intrusive centre. This gabbro is medium- to coarse-grained and contains pegmatitic phases. The Cu and PGE enrichment is considered to be of hydrothermal origin(Watkinson and Ohnenstetter(1992) and is represented by platinum-group minerals, chalcopyrite, and cubanite. The Canadian Mines Handbook(1993-1994, p. 145) lists the reserves as 37,000,000 tons grading 0.31 % Cu, 0.04 % Ni, 251,000 oz Pt, 1,001,000 oz Pd, 84,000 oz Au and 43,000 oz Rh. One should refer to Ohnenstetter et al. (1991), Watkinson and Ohnenstetter(1992), Good and Crocket(1994b) and Shaw(1994) for details on the mineralization and petrology of eastern gabbro.

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