Rb-Sr whole rock and U-Pb zircon geochronology of the Michipicoten Greenstone Belt, Wawa area, northwestern Ontario.

Patrick E. Smith

University of Windsor

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LA THÈSE A ÉTÉ MICROFILMÉE TELLE QUE NOUS L’AVONS RÉCEVE
Rb-Sr WHOLE ROCK AND
U-Pb ZIRCON GEOCHRONOLOGY
OF THE MICHIPICOTEN GREENSTONE BELT,
WAWA AREA, NORTHWESTERN ONTARIO

by

PATRICK E. SMITH

A thesis
Submitted to the Faculty of Graduate Studies through
the Department of Geology in Partial Fulfillment
of the Requirements for the Degree of
Master of Science at the
University of Windsor

Windsor, Ontario
1981
ABSTRACT

The Archean crust in the Wawa area of Ontario consists of complexly folded and faulted metavolcanic and metasedimentary rocks which constitute the Michipicoten greenstone belt. These supracrustal rocks are intruded by granitic stocks and embayed by granitic rocks of batholithic dimensions. The metavolcanic rocks are subdivided into three sequences, lower, middle and upper, and are stratigraphically separated by intervening metasedimentary rocks. Published K-Ar ages range from 800 to 2505 Ma and have little stratigraphic significance. This study reports Rb-Sr and U-Pb ages for acid volcanic rocks from the upper and lower volcanic units and for granitic rocks from within and outside the greenstone belt. The apparent Rb-Sr ages are 2530 ± 90 Ma and 2680 ± 490 Ma for the acid volcanic rocks and 2550 ± 175 and 2560 ± 270 Ma for the granitic rocks, and are interpreted to reflect the Kenoran orogeny. The uncertainties in the Rb-Sr ages are large, and the scatter of the data is attributed to a younger thermal event at 2285 ± 70 Ma. In contrast, the zircon data yield low uncertainties in ages and are stratigraphically meaningful. Acid volcanic rocks from the lower sequence have U-Pb ages of 2744 ± 10 Ma and 2749 ± 2 Ma. An acid volcanic rock from the upper unit has an age of 2696 ± 2 Ma thus bracketing a minimum time span of 53 Ma for the evolution of the belt. A granitic stock within the belt has an age of 2737 ± 6 Ma and is considered coeval and cogenetic with the acid volcanics of the lower sequence. The Hawk Lake trondhjemite, which borders the greenstone belt on the east also appears coeval with the lower volcanics, having an age of 2747 ± 6 Ma, but contains evidence of an earlier sialic rock exceeding 2812 ± 3 Ma in age.
Zircon ages obtained can be correlated with other greenstone belts of the Superior province and when all volcanic ages are considered certain systematic trends are evident. Greenstone belts get progressively younger to the east. Also, the duration of volcanism becomes progressively shorter eastwards. These apparent trends are suggestive of plate tectonic mechanisms involved in the eastward accretion of new crustal material onto the early Superior protocontinent.
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Finally I thank my fiancee Bosiljka for her assistance in typing this thesis and my brother James for his help with the field work. This study has been financed by Natural Science and Engineering Research Council of Canada grants A8299 to Dr. Turek.
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INTRODUCTION

Greenstone belts are common to Precambrian rocks all over the world and they are of particular geochronological interest for numerous reasons. These areas host rich mineral deposits (e.g. iron ore, gold, massive sulfides) which bear uncertain relationships to the stratigraphy of the areas. Thus a more detailed documentation of the stratigraphy of these areas enables a better understanding of the relationships between the rock units and the mineral deposits. Geochronological mapping is an invaluable tool in unravelling the stratigraphy, especially in Archean terrain, where folding and faulting combined with complex facies relationships hinders geologic mapping.

The data arising from this study, and others similar to it, lead to a more precise geochronological definition of the stages involved in Archean crustal development - more specifically, the development of a greenstone belt and associated granitic terrain. It is hoped that this data, when combined with structural and geochemical information, will help define a clearer picture of processes operative in the Archean, and elucidate similarities and differences in geotectonic mechanisms operating today.

There is great need for a unification of the Precambrian time scale - both on a local and a more regional scale - based on absolute chronostratigraphic units. K-Ar ages, because they tend to reflect metamorphic events, are inadequate for this purpose. However, ages obtained by the Rb-Sr and U-Pb methods provide a superior basis for the subdivision of Precambrian time based on primary stages in the earth's evolution. The resulting ages can be regarded as
chronostratigraphic units. This study may be considered part of the basis for this chronology.

The Rb-Sr method has been firmly established for use in the Precambrian, not only in deducing age relationships but also (via the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio) in providing evidence for magmatic provenance, and in tracing the evolution of the earth's crust and mantle. The importances of zircon U-Pb dating to deduce Archean stratigraphy has recently been demonstrated (Pye, 1980). This is possible because of the greatly enhanced precision of ages obtained by way of recent advances in zircon methodology (Krogh, 1973) and refinements in mass spectrometry. In favourable cases zircons, from the Archean, have been shown to be analytically precise to $\pm 1$ Ma (Nunes and Thurston, 1980).

This thesis reports Rb-Sr whole rock and U-Pb zircon ages for the Michipicoten greenstone belt, a sequence of complexly folded, Archean, metavolcanic and metasedimentary rocks located in the Wawa subprovince of the Superior structural province. The belt is located in the district of Algoma, about 200 km north of Sault Ste. Marie and occupies an area of approximately 1500 km$^2$ (Fig. 1). The study area is centred around the vicinity of Wawa and the units sampled include volcanic and granitic rocks within the belt and granitic rocks from outside the belt.

The study was undertaken:

1. To obtain meaningful and precise ages for the metavolcanic rocks of the belt and thus document geochronologically the evolution of volcanism in the area.
Fig. 1. Geological sketch map of the southern portion of the Michipicoten greenstone belt showing sample locations. (Based on Goodwin, 1962, O.D.M. compilation map, 1971, and Sage 1979, 1980.)
2. To investigate the relationships between the volcanic rocks and granitic rocks within and outside the belt.

3. To use the obtained chronostratigraphic markers to correlate time stratigraphically magmatic episodes in the belt with other similar belts within the Superior province.

4. To compare the ages obtained with detailed geologic mapping of the area.

The following is a brief outline of the contents of this thesis. The Geologic Setting gives a description of the regional geology with emphasis on the Michipicoten belt in the context of the Superior province in Archean time. This is followed by a detailed description of the geology of the Michipicoten belt based predominantly on the work of Goodwin (1962, 1966), Attoh (1980) and Sage (1979, 1980) who is currently mapping the area.

The Theoretical Background deals with the principles of the Rb-Sr and U-Pb absolute dating techniques. The relevant age equations, as used, are included.

In the Analytical Procedure the treatment and preparation of raw rock samples for age work is described. Also a brief outline of the x-ray and mass spectrometer procedure used is given.

The Discussion of Errors examines the possible uncertainties incurred in the acquisition of data and subsequent data reduction. In addition a short description of the two-error least-squares linear regression technique as applied to Rb-Sr isochrons and U-Pb concordia diagrams is given, as well as a statistical method for distinguishing between ages.
Following this are the Analytical Results, along with interpretations of the obtained ages, and the Discussion of age relationships within the Michipicoten belt and between other belts. Following this, the major Conclusions drawn from this thesis are summarized in point form.

Appendix I gives the petrography and location of the samples used for age work. Appendix II outlines the procedure for zircon separation. Appendix III gives the calibration graphs for the ion exchange of Rb and Sr. Appendix IV gives Na$_2$O, K$_2$O data for the samples which was part of a project to screen nonrepresentative samples used for Rb-Sr whole rock work.
Previous Geochronological Work

Nine K-Ar whole rock and mineral ages have previously been reported for the Michipicoten belt (Leech et al., 1963; Lowden et al., 1963; Wanless et al., 1965; Wanless et al., 1966; Wanless et al., 1974). The ages are plotted on a histogram in Fig. 2, and range from 800 to 2505 Ma. In addition, these ages combined with other ages from the vicinity (i.e. outside the greenstone belt) are shown on another histogram in Fig. 2. The apparent lack of age maxima of these histograms is a reflection of the fact that K-Ar systems can readily be updated, totally or partially by younger events. The rocks of the area have undergone at least a green-schist metamorphism and temperatures of (ca 300° C) are sufficient, in most cases, to disturb K-Ar ages. Thus, many of these ages are considered minimum ages and have limited meaning. Also, rocks can soak up Ar as the result of metamorphism and thus yield erroneously old ages. Several of these ages are those of diabase dikes intruding local country rock and may be related to the Matachewan (ca 2500 Ma), Nippissing or Abitibi dike swarms (ca 2200 Ma), as well as to the younger Keweenawan volcanic activity in the Lake Superior region (ca 1100-1200 Ma).

In addition to the K-Ar ages, two Rb-Sr whole rock isochron ages are reported for the metavolcanic rocks of the area by Brooks et al. (1969). They are 2650 ± 80 Ma with initial $^{87}$Sr/$^{86}$Sr ratio of 0.7012 ± 0.0003 for the lower volcanics and 2500 ± 100 Ma with initial $^{87}$Sr/$^{86}$Sr ratio of 0.7028 ± for the upper volcanic rocks (recalculated ages using $^{87}$Rb = 1.42 * $10^{-11}$ yr$^{-1}$). The authors conceded that the ages for the upper volcanics have been disturbed by metamorphism, however, considered the initial ratios to be undisturbed. Unfortunately, Brooks et al. (1969) do not report
Fig. 2. Distribution of K-Ar ages for Wawa vicinity of the Superior province, including the Michipicoten area (top), and for the Michipicoten belt (bottom). Symbol "d" indicates age of diabase dikes.
their analytical data. This precludes comparisons with Rb-Sr work done in this study which would otherwise be possible.

In a regional context, relevant to this study, are the Rb-Sr and U-Pb ages reported for the Gamitagama area to the south of the Michipicoten belt reported by Turek et al. (1981) and Krogh and Turek (1981).
GEOLOGIC SETTING

The Lake Superior area is located at the southern portion of a vast, exposed continental nucleus of Precambrian rock, the Canadian Shield. The Superior province is one of nine structural provinces of the Canadian Shield which are recognized on the basis of structural trends and isotopic ages (Stockwell, 1964). The Superior province is itself divided into a number of subprovinces consisting of linear, east-west trending volcanic-granitic zones (greenstone belts) alternating with sedimentary-granitic zones. The Wawa subprovince is one of the former zones and comprises the Gamitagama belt and the Michipicoten belt.

The Precambrian in Canada is subdivided on the basis of four major orogenic events. The boundaries between the eras differ slightly because two different absolute age systems are used to delineate them. Archean time includes geologic time older than the Kenoran orogeny. Recently, the Archean has been subdivided into three parts based on the Uivakian, Wanipigowian and Laurentian orogenies which correspond to ages of 3400, 2900 and 2500 Ma respectively (Stockwell, 1980; Douglas, 1980). Geologic time after the Archean, in the Precambrian, is called Proterozoic and is itself further subdivided based on the Hudsonian, Elsonian and Grenville orogenies. The majority of rocks of the Superior province are Archean, and thus were last affected by regional metamorphism during the Kenoran orogeny 2520 to 2670 Ma ago (Douglas, 1980).

The Michipicoten greenstone belt is representative of typical island arc volcanism in the Superior province during Archean time.
(Goodwin, 1968). Archean greenstones were emplaced on top of a pre-existing basement; however, the age and composition of this basement are controversial. Evidence that is suggestive of a sialic crust is summarized by Baragar and McGlynn (1976, 1978). Glikson (1978, 1979), however, provides evidence of a simatic basement. It is quite possible that this basement was a combination of sialic and simatic compositions (Goodwin, 1977).

Cycles of mafic to felsic volcanism are common to greenstone belts of the Superior province. Initial stages of formation of a greenstone belt began with voluminous outpourings of basic lava onto the sea floor via volcanic vents, creating shield volcanoes which coalesced to form island chains. In time, the composition of the lava changed, by magmatic differentiation, through andesitic and dacitic lavas to rhyolite. The latter lava types were gas charged and viscous, and therefore tended to erupt explosively in the form of breccias and ashes. Thus the island arcs were capped and flanked with breccias, while more distal cover consisted of settled ash forming tuffaceous layers. The end of the cycle was characterized by hot spring and fumerolic activity giving rise to iron formations and carbonate deposition. The cycle then repeated itself with renewed mafic flows burying previously deposited strata. Clastic sedimentary rocks accumulated during relative volcanic quiescence and were derived contemporaneously from adjacent volcanics and intrusives by rapid erosion.

While the island arcs were forming, the rocks underlying them were intruded by granitic rocks in the form of batholiths, stocks and sills. Then the rocks were extensively folded and recrystallized.
during the Kenoran orogeny. Numerous granitic bodies of all sizes were formed and engulfed the volcanics and sediments. Following this regional event the rocks were faulted and intruded by dikes.

**Geology of the Michipicoten Belt**

Preliminary geologic mapping of the Michipicoten belt was related to the discoveries of gold and iron exploration. Early investigators include Gledhill (1927), Moore (1931), and Frohberg (1935). Goodwin (1962, 1966) presented a broad overview of the belt with emphasis on the iron formations in relation to volcanic stratigraphy. A more comprehensive stratigraphic study of localized areas within the belt was undertaken by Attoh (1980). In addition, the Michipicoten belt is presently being mapped in detail by the Ontario Geological Survey (Sage 1979, 1980). The following description of the regional geology is based on Goodwin (1962, 1966), Attoh (1980), and Sage (1979, 1980). The general geologic features discussed are depicted in Fig. 3 (after Goodwin, 1966).

The Michipicoten belt has a roughly east-west regional trend and consists of Keewatin type flows and pyroclastic rocks in association with structurally conformable clastic sediments and banded iron formations. These rocks are bordered by batholithic masses of younger granitic rocks. Volcanic rocks are of the andesite-rhyolite association and exhibit great heterogeniety with respect to lithology, structure and thickness. The clastic metasedimentary rocks are predominantly greywackes, shales and argillites, and are the erosional products of volcanic intrusive rocks.

The belt evolved cyclically, through a series of eruptive, erosional and hydrothermal processes, which produced a complex volcanic-sedimentary assemblage related to a common volcanic origin.
The estimated stratigraphic thickness of the belt is ca 8000 m 
(Attoh, 1980).

Structure

The Michipicoten belt has undergone extensive folding and 
faulting. Rocks of the area have been folded along three east-west 
trending axes, forming a central anticline and flanking 
synclines. The axes are approximately equally spaced; the most 
northerly axis is located near the upper extremity of the belt, and the 
southerly axis is located just north of Wawa Lake. These structures 
are themselves cross folded about northwest trending axes forming a 
central cross-syncline and two flanking cross anticlines. The cross 
folds are inclined or overturned to the southwest causing the east-
west trending folds to plunge steeply west and gently east. As a 
result, all fold axes undulate in both longitudinal and transverse 
sections and hence in plan view the axes are sinuous in trend.

Two sets of faults have offset these folds. A series of north 
to northwest striking transverse faults, with left handed horizontal 
offsets, have divided the area into large parallel blocks. Horizontal 
offsets range up to 5 km in length, the distance decreasing northward. 
Vertical components of movement are possibly greater than the hori-
zontal displacement. Overall fault movement is west side down and to 
the south with respect to the east side.

A second set of faults, found in the southern portion of the 
area, strike about N 60° E and dip steeply to the south. Horizontal 
offsets may be as much as 5 km and the movement is left handed. 
Vertical displacement is not known. Some more obvious offsets mani-
fest themselves as aligned, linear lakes – eg. Wawa Lake, Hawk Lake and Manitowik Lake.

A number of fresh Keweenawan dikes are found intruding along the transverse north-northwest trending faults. This faulting, together with dike intrusion, appears to be related to regional subsidence during the Keweenawan time (Goodwin, 1962).

Supracrustal rocks of the area have undergone regional metamorphism and display greenschist to amphibolite metamorphic facies. In general the macroscopic structures and textures of the rocks are will preserved even though the mineralogy has changed. In some cases rocks have a high content of carbonate.

**Stratigraphy**

Metavolcanic rocks of the belt are subdivided into lower, middle and upper units separated by two intervening metasedimentary horizons – the banded iron formation horizon and the Dore clastic rocks. The distribution of volcanic rocks and metasedimentary rocks within the stratigraphic pile is complicated by the heterogeneity of volcanism and the diversity of sedimentary responses to the volcanism.

Factors contributing to stratigraphic complication may be summarized as follows:

i) Thick hybrid piles were formed by the simultaneous ejection of flows and pyroclastics from the same or nearby vent. ii) Violent, felsic pyroclastic eruptions occurred contemporaneously with more fluid andesitic effusion producing thick felsic units with thin mafic time equivalents. iii) Discharge areas migrated throughout the evolution of the pile producing time rock equivalents ranging from
thin or absent to very thick; composition ranging from felsic to mafic, within small lateral intervals. iv) Where sedimentation coincided with volcanism in depositional basins at some distant vent, volcanic rocks and sedimentary rocks accumulated in the pile at different rates in overlapping environments. v) Where volcanics accumulated rapidly, and in large quantities in comparison to sediments, thick sequences of volcanic units with thin sedimentary time equivalents resulted. vi) In areas of local volcanic quiescence, subaerial exposure of thick volcanic piles resulted in thick sediments with thin volcanic equivalents.

**Lower Volcanics**

The lower volcanic rocks represent the dominant stratigraphic unit and are distributed throughout the belt. They are composed of a lower intermediate to mafic flow section and an upper felsic pyroclastic section.

The intermediate to mafic section ranges in composition from rhyolite to basalt, with predominant andesite, and consists of medium grey-green flows and breccias. Black graphitic interflow shales and tuffaceous layers are also present. Diorite is present in concordant layers within the flows.

A transition zone, 300 to 900 m thick, separates the lower mafic flow section from the upper pyroclastic section. The transition zone is marked by rocks with angular fragments of whitish rhyolite set in a grey-green dacitic matrix - increasing in size and proportion upward.

Rocks from the upper felsic pyroclastic section form large massive coalescing domes, recognized individually as breccia domes. They are
composed of homogeneous rhyolite breccia, hybrid rhyolite-dacite-andesite breccia, and dacite to andesite breccia, and flows. Dikes and sills of diorite and quartz porphyry, contained within the domes, are possible ancient feeder systems for the extrusives. This supposition is supported by the structural conformity of the dikes and sills with the encasing volcanics and the lack of contact metamorphic aureoles.

The size of the fragments in the fragmental acid volcanic rocks ranges from small grains to blocks one half meter in diameter and are typically composed of massive porphyritic or banded rhyolite, with subordinate jasper, quartz and mafics. The matrix of the breccias is typically massive to porphyritic rhyolite having the same composition as the fragments. Vesicles and quartz-carbonate filled amygdules are common features in both matrix and fragments.

Large arcuate quartz grains, embedded in a matrix of dense pale green rhyolite or dacite, are characteristic of some rhyolite flows of the lower volcanics.

The lower volcanics are carbonitized to a large extent. Ankeritic carbonate is found in small veinlets or in irregular masses. Associated with the carbonate is white quartz which is found in vein or in disseminated form.

**Iron Formation**

Banded iron formations are distributed in discontinuous belts throughout the major synclines and cross synclines. In most areas they conformably overlay the lower volcanics. Iron formations of the area are unique in that they contain two mutually gradational facies associated with sedimentary and volcanic rocks, as noted by Goodwin.
(1962). Iron formations associated with sediments are composed of alternating layers of chert, silicius magnetite, and jasper, and characteristically occur in discontinuous bands and lenses. On the other hand, iron formations of volcanic association are composed of banded chert, pyrite, and carbonate members, in descending stratigraphic order. The iron formation of the volcanic association typically overlies felsic volcanic rocks and is itself overlain by intermediate to mafic rocks.

**Middle Volcanics**

Middle volcanic rocks occupy a conformable stratigraphic position overlying banded iron formation and underlying Dore metasedimentary rocks. They are located on the eastern portion of the central east-west trending anticline, and along the eastern flank of the central northwest trending cross-syncline.

The middle volcanics constitute andesitic flows and pyroclastics in bands ranging from 600 to 3000 m in thickness. Ankeritic carbonate is broadly found as disseminated grains, amygdule fillings, and veinlets. Differentiation was limited as is indicated by the mutual extrusion of basaltic flows and thin layers of rhyolite-dacite pyroclastic rocks. The transition from the middle volcanics to the Dore sediments is gradational and represents an overlapping of sedimentary and volcanic environments.

**Dore Sedimentary Rocks**

The Dore sediments are distributed in east-west trending linear belts following the two major synclines located to the north and south of the belt. In addition, narrow sinuous bands trending northwest
are located on the limbs of the central cross-syncline.

The Dore rocks are a series of clastic rocks derived from adjacent extrusive and intrusive rocks. They consist of greywacke, shale, argillite, and conglomerate, with subordinate quartzite. Generally, thick conglomerate-greywacke facies prevail to the west and thin shaly facies are found to the east.

The upper and lower contacts to the Dore sediments are gradual over several hundred meters, implying a very transitional volcanic-sedimentary relationship. Also, thin andesitic flows are intercalated with sediments which indicates limited volcanic activity during Dore time.

The presence of granitic detritus in the Dore group is attributed to the erosion of exposed, shallow granitic apophysies (Goodwin, 1962).

Upper Volcanics

The upper volcanic sequence is located in a northwest trending belt following the axis of the central cross-syncline. The rocks represent a third major cycle of volcanism following the deposition of the Dore sediments. They comprise mafic to intermediate flows and subordinate pyroclastics. Felsic pyroclastics, where present, are characteristically near the base of the group. Diorite and gabbrosills are common and are probably coeval with the extrusive units. As in the other volcanic cycles, carbonitization is widespread.

Granitic Plutons

Granitic rocks, with broad gneissic margins, surround the Michipicoten belt on all sides. In most instances the rocks occupy a conformable stratigraphic position underlying the volcanic-sedimen-
tary rocks; thus, the granitic margins follow the structural distribution of these rocks. The transition from unaltered volcanic-sedimentary rock to granite gneiss grades from dense, well banded, porphyritic schists with quartz veins, aplites and pegmatitic dikes, to gneissic granite, commonly having relict patches of volcanic or sedimentary material. This transition zone may be up to 1 km in width. Generally, the bordering granitic rocks exhibit gneissosity conformable to the primary flow and bedding features in adjacent volcanic-sedimentary rocks. This feature is apparent when the rocks are steeply dipping, as along the north and south borders of the belt, or gently dipping, as along the western contact.

Numerous plutons intrude the lower volcanics to the south of Wawa Lake and are the host rocks for gold quartz mineralization. These rocks exist as relatively small masses and are possibly derived from a common magma. They may also be allied to or part of the Brule Bay batholith, to the south of the Michipicoten belt, described by Coleman (1906).

In this study the granitic rocks that surround and embay the supracrustal rocks are referred to as the external granites, while the granitic rocks that are physically within the supracrustal rocks are referred to as the internal granites.

**External Granite - Hawk Lake Trondhjemite**

A batholithic complex of acid plutonic rocks borders the supracrustal rocks at the east end of the belt at Hawk Junction. Rock types here include a medium grained equigranular foliated trondhjemite, and

* The term trondhjemite as used in this study conforms to the established usage by Ontario Department of Mines (now Ontario Geological Survey) and follows the definition given by Ayres (1972). Using Streckeisen's (1973) classification most of these trondhjemite rocks would be named tonalites.
granodiorite which grade into quartz and quartz-feldspar porphyry near the contact with the supracrustal rocks. Sage (1980) has referred to this terrain as Hawk Lake granitic complex.

Granitic rock, intruding the Michipicoten belt about 20 km south of Hawk Junction, is possibly an extension of, or related to, the Hawk Lake trondhjemite. Here the rocks are massive to weakly schistose trondhjemite which themselves have been intruded by younger granodiorites to quartz monzonites. The supracrustal granite contact is marked by agmatite or mega-agmatite, and appears partly fault controlled (Sage, 1979).

The Internal Granites - Diorite-Granodiorite Stocks

Intrusive into the lower volcanics are two small stocks, both about 5 km² in surface exposure, located just to the south of Wawa Lake. The more northerly of these two stocks has been called Jubilee stock (Fig. 1) by Sage (1981). It is composed of diorite, quartz diorite and granodiorite and intrudes tuffs and flows of the lower volcanic series. The contact of the stock with the supracrustals is abrupt and consists of an intrusive breccia. Numerous xenoliths and supracrustal blocks have been incorporated into the stock. The surrounding volcanics show little evidence of contact metamorphism or assimilation and are interpreted by Sage (1979) as effusive material associated with emplacement of the stock.

The second stock to the south of Wawa Lake and also to the south of the Jubilee stock (Fig. 1, sample 664) is composed of granodiorite and is similar petrologically and structurally to the Jubilee stock.

Diabase Dikes

Diabase dikes of at least two ages are recognized in the Michipicoten area. The older set predate the granitic rocks and generally trend
northwest. They are fine to medium grain rocks, typified by variable widths, ranging from 3 to 45 m. The dikes are green in colour which is a result of advanced alteration of primary augite to uralitic hornblende.

A younger swarm of diabase dikes, also striking northwest, postdate the granitic intrusives and quartz veins. In contrast to the older set, they are characterized by consistent thickness and a fresh black appearance.

Fig. 2 which shows the K-Ar ages for the Michipicoten area indicates which of these samples are derived from dikes. All of these dikes are diabase dikes, however, the trends are not reported. Therefore, their exact identity is unknown.

Carbonatite Complex

The Firesand carbonatite complex is a small oval intrusive body - about 2.5 km in diameter - located about 3 km east of the Jubilee stock (Sage, 1981). This intrusion consists of a ring of sovite and silicocarbonate with a core of ferruginous dolomite. Surrounding wall rock is brecciated and the country rock is fenitized up to 1.6 km. The complex has K-Ar ages of 1008 and 1087 Ma (Gittins, 1967) and 1048 Ma (Wanless, 1970).

Lamprophyre Dikes

Lamprophyre dikes in the Michipicoten belt predominate in the region south of Wawa Lake and strike in a northeasterly direction. They range in width from a few centimeters to over 1.5 km. Olivine may be present as an essential or accessory mineral and is commonly altered to lighter colour carbonates and serpentine, imparting a spotted appearance to these dikes.

It is possible that these Lamprophyre dikes are related to the carbonatite complex. However, such a relation has not been suggested by either Goodwin (1962), Sage (1980) or Attewell (1980).
THEORETICAL BACKGROUND

Theoretical aspects of the Rb-Sr and U-Pb methods are given here as background information. Several books have been written on the subject; the most recent and most comprehensive is by Faure (1977). Some other useful texts are those by York (1973) and Hamilton (1965). The original definitive papers were published as "benchmark" papers by Harper (1973).

Radiometric age dating is based on the principle of radioactive decay. According to the theory, the rate of decay of a parent nuclide, or the number of atoms that disintegrate per unit time (-dN/dt) is proportional to the number of atoms present (N). With the introduction of a proportionality constant or decay constant (λ), the law can be expressed as an equality:

\[
\frac{-dN}{dt} = \lambda N
\]

Integration of this expression leads to the basic equation of describing radioactive decay processes:

\[
N = N_0 e^{-\lambda t}
\]

where N is the number of radioactive parent atoms that remain at any time t of the original number of atoms present \(N_0\) at \(t = 0\). \(N_0\) is a critical unknown and if the assumption is made that there is no loss or gain of parent isotope (N) or daughter isotope (D) then \(N_0 = N + D\). Substitution of this for \(N_0\) in the above equation, leads to the general age equation used in geochronology,
\[ D = D_0 + N(e^{\lambda t} - 1) \]

where \( D_0 \) represents the number of daughter atoms initially present.

**The Rb-Sr Method**

Rb is a dispersed element which occurs in minerals substituting for K. Rb has two isotopes, \(^{85}\text{Rb}\) and \(^{87}\text{Rb}\), with natural abundances of 72.15\% and 27.85\% respectively. \(^{87}\text{Rb}\) is radioactive and decays by \(\beta\) particle emission to \(^{87}\text{Sr}\). The half-life of \(^{87}\text{Rb}\) is \(4.8 \times 10^{10}\) \(\text{y}\). The age dating of Rb rich mineral is complicated by the fact that Sr is present before the growth of radiogenic \(^{87}\text{Sr}\) in the mineral. This initial Sr is referred to as "common Sr" and consists of isotopes \(^{84}\text{Sr}\), \(^{86}\text{Sr}\), \(^{87}\text{Sr}\) and \(^{88}\text{Sr}\) having natural abundances of 6\%, 10\%, 7\% and 83\% respectively. Since common Sr contains \(^{87}\text{Sr}\), it is necessary to know the contribution of this isotope to the total \(^{87}\text{Sr}\). Thus the initial \(^{87}\text{Sr}\) has to be known or calculated. Substitution of the pertinent isotopes into the age equation and dividing by a non-radiogenic reference isotope, \(^{86}\text{Sr}\), the expression for the growth of radiogenic \(^{87}\text{Sr}\) in a system is expressed as follows:

\[
\frac{^{87}\text{Sr}}{^{86}\text{Sr}} = \left( \frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_0 + \frac{^{87}\text{Rb}}{^{86}\text{Sr}} \left( e^{\lambda t} - 1 \right)
\]

where \(\left( \frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_p\), hereafter \(R_p\), is the present day ratio and \(\left( \frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_i\), hereafter \(R_i\), is the initial ratio at \(t = 0\). From this equation it is possible to calculate a model age of a rock by assuming an \(R_i\) and determining \(^{86}\text{Sr}\) and \(^{87}\text{Rb}\) by mass spectrometry or by X-ray fluorescence and by determining the \(R_p\) directly by mass spectrometry.
The age is obtained by rearranging equation 1.

\[
t = \frac{1}{\lambda} \ln \left( \frac{87\text{Sr}}{86\text{Sr}} - \left( \frac{87\text{Sr}}{86\text{Sr}} \right)^{0} \right) + 1
\]

The numerical value of \( t \) obtained will represent the true age of the rock provided the rock has remained a closed system with respect to Rb and Sr and an appropriate choice of \( R_i \) was made.

Equation 1 is the equation of a straight line whose slope is related to the age, and the Y-intercept corresponds to the \( R_i \). Thus, if a suite of cogenetic and coeval rocks are plotted on a graph of \( \frac{87\text{Sr}}{86\text{Sr}} \) vs. \( \frac{87\text{Rb}}{86\text{Sr}} \) a straight line results which is called an isochron (Nicolaysen, 1961). The age of the rock is given by:

\[
t = \frac{1}{\lambda} \ln (1 + \text{slope})
\]

The \( R_i \) is obtained by extrapolating the line to the intercept on the \( R_p \) axis. If coeval and cogenetic samples do not fit an isochron within experimental error, then the scatter may be attributed to loss or gain of Rb or Sr via metamorphism. However, if metamorphism has caused complete mixing of common and radiogenic Sr, sample points will again be colinear but the \( R_i \) will tend to be high. In this situation the indicated age will be the age of metamorphism or anatexis.

The U-Pb Method

U-Pb age dating was done exclusively on zircons in this study. Zircon crystals (ZrSiO₄) contain trace amounts of U and Th. The presence of these elements in zircons is attributed to the isomorphous
substitution of Th$^{4+}$ and U$^{4+}$ for Zr$^{4+}$ because of their similar charge and ionic radius. U has three natural radioactive isotopes, one of which can be totally disregarded as its abundance is negligible; the other two, $^{238}$U and $^{235}$U, breakdown by series decay to $^{206}$Pb and $^{207}$Pb respectively. A third radiogenic lead isotope, $^{208}$Pb, comes from the series decay of $^{232}$Th. Lead has one more isotope, $^{204}$Pb, which is non-radiogenic and is treated as a reference isotope.

When substituted in the general age equation, the equations for the growth of radiogenic Pb in U minerals is as follows:

$$\frac{^{206}\text{Pb}}{^{204}\text{Pb}} = \left( \frac{^{206}\text{Pb}}{^{204}\text{Pb}} \right)_0 + \frac{^{238}\text{U}}{^{204}\text{Pb}} (e^{\lambda_1 t} - 1)$$

$$\frac{^{207}\text{Pb}}{^{204}\text{Pb}} = \left( \frac{^{207}\text{Pb}}{^{204}\text{Pb}} \right)_0 + \frac{^{235}\text{U}}{^{204}\text{Pb}} (e^{\lambda_2 t} - 1)$$

where $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, $^{238}\text{U}/^{204}\text{Pb}$ and $^{235}\text{U}/^{204}\text{Pb}$ are the present day ratios; $(^{206}\text{Pb}/^{204}\text{Pb})_0$ and $(^{207}\text{Pb}/^{204}\text{Pb})_0$ are the initial isotope ratios of Pb that contaminated the mineral at the time of its formation; and $\lambda_1$ and $\lambda_2$ are the decay constants of $^{238}\text{U}$ and $^{235}\text{U}$ respectively. For estimates of the initial Pb isotope concentrations, that is common Pb contamination to be subtracted from the data, it is assumed that the Pb incorporated into the mineral at the time of its formation has an isotopic composition similar to Pb found in Pb minerals of comparable age. The assumed initial ratios are derived from a two stage Pb evolution model formulated by Stacey and Kramers (1975). Thus separate ages can be determined based on two different decay schemes:
\[ t_{206} = \frac{1}{\lambda} \ln \left( \frac{206_{\text{Pb}}}{204_{\text{Pb}}} \right) \left( \frac{206_{\text{Pb}}}{204_{\text{Pb}}} \right)^0 + 1 \]

\[ t_{207} = \frac{1}{\lambda} \ln \left( \frac{207_{\text{Pb}}}{204_{\text{Pb}}} \right) \left( \frac{207_{\text{Pb}}}{204_{\text{Pb}}} \right)^0 + 1 \]

These dates will represent crystallization ages provided the mineral has remained a closed system to U, Pb and the intermediate daughter products of the decay series. In fact, this is rarely the case and most ages reported based on these independent decay schemes are discordant (i.e. too young) and characteristically show \( t_{206} < t_{207} \). However, ages calculated from the ratio of the two Pb isotopes is less sensitive to U or Pb loss. This age is based on the \( \frac{207_{\text{Pb}}}{206_{\text{Pb}}} \) ratio and avoids the necessity of quantitative U or Pb determinations. The equation is obtained by combining equation 1 and 2:

\[ \frac{207_{\text{Pb}}}{204_{\text{Pb}}} \left( \frac{207_{\text{Pb}}}{204_{\text{Pb}}} \right)^0 = \left( \frac{235_{\text{U}}}{238_{\text{U}}} \right) e^{-\frac{\lambda 2^t}{2}} - 1 \]
The $^{235}\text{U}/^{238}\text{U}$ ratio is a constant and can be replaced by 1/137.88. This is a transcendental equation and its solution is determined by carrying out successive approximations of $t$ for a given $^{207}\text{Pb}/^{206}\text{Pb}$ ratio.

Despite the superiority of the $^{207}\text{Pb}/^{206}\text{Pb}$ age calculation over ages calculated with individual decay schemes, the majority of these ages turn out to be discordant. This problem is overcome by the use of the concordia diagram of Wetherill (1956). This diagram features the concordia curve which consists of the locus of all points giving equal ages by both U decay schemes. If a set of discordant cogenetic zircons are plotted on a graph of $^{206}\text{Pb}/^{238}\text{U}$ vs. $^{207}\text{Pb}/^{235}\text{U}$, they will form a linear array, or discordancy line, that intersects the concordia curve in two places. The upper intersection corresponds to the crystallization or primary age of the rock. The meaning of the lower intersection is less well understood and can have several interpretations. This may correspond to a time when discordant zircons experienced an episodic loss of Pb, or gain of U (Wetherill, 1956); this is the most generally accepted view. Tilton (1960) proposed that radiogenic Pb was lost from crystals by continuous diffusion and concluded that where this mechanism was operative the lower intercept was spurious. Goldich and Mudrey (1972) proposed a dilatancy model for discordant zircons whereby uplift and erosion caused loss of radiogenic Pb with water which was previously held tightly within the crystals. Regardless of the mechanism used to explain Pb loss or U gain to generate the lower concordia intercept, the upper concordia intercept remains unquestioned as representing the primary age for cogenetic zircons.
ANALYTICAL PROCEDURE

Sampling and Sample Preparation

Representative, fresh and unaltered samples were collected in the field. For Rb-Sr ages approximately 2 to 5 kg of sample was taken. For zircon U-Pb ages approximately 100 kg of sample was collected. All weathered surfaces were chipped off. The samples were then crushed using a jaw crusher which reduced the sample to approximately 0.5 cm size. This was followed by pulverizing the sample to about -80 Tyler mesh size. For Rb-Sr age determinations, approximately 20 g of sample was subsampled and further crushed using a shatter box to reduce the sample size to about -200 Tyler mesh size. The large 100 kg sample, designated for zircon separation and analysis, after pulverizing it to -80 mesh size, was concentrated to about 1 litre volume of heavy minerals using a Wilfley table. This concentration technique takes approximately 16 hours. The Wilfley table concentrate was then dried and a zircon concentrate was obtained using a combination of magnetic and heavy liquid separations. Final separation involved hand picking under a binocular microscope to yield 99.9% pure zircons. At this stage typical zircon concentrate is about 0.5 g to 1.0 g. It was further sorted by a Franz magnetic separator into fractions of different magnetic susceptibilities, and also it was sieved using nylon screens, into the following Tyler mesh size ranges: 70-100, 100-200, 200-350, and 350-450. This procedure is described in more detail in Appendix 1. No mineral separation was involved in Rb-Sr age determinations, all samples were analysed as total rocks.

X-ray Technique:

A Philips 1410 x-ray spectrometer with a 220 LiF analysing
crystal was used to analyse all samples for their approximate Rb and Sr content. Samples were prepared as pellets by taking 2.5 g of sample, adding 1 ml of polyvinyl alcohol as binder, and 3 g of boric acid as backing, and pressing for 15 seconds at 8 tonnes.

This semiquantitative estimate of Rb and Sr concentration is needed to assess the technical suitability of a sample for Rb-Sr age determination. It is desired to have a spread of Rb/Sr ratios and hence a spread in \(^{87}\text{Sr}/^{86}\text{Sr} \) values in order to define a good isochron age. To estimate an expected present day \(^{87}\text{Sr}/^{86}\text{Sr} \) ratio of a sample, the age and initial \(^{87}\text{Sr}/^{86}\text{Sr} \) is assumed and substituted in the age equation which can be rearranged as follows:

\[
\left( \frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_p = \left( \frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_o + 4.0973 \times 10^{-5} \times t \times \left( \frac{\text{Rb}}{\text{Sr}} \right)_{\text{wt}}
\]

Also, for isotope dilution measurement of Rb and Sr concentration the estimated Rb and Sr contents of the sample governs the quantity of isotopic tracer added to the sample to optimize mass spectrometer ratio measurement.

Samples selected for age determination were then reanalysed accurately in triplicate for Rb and Sr concentrations. The procedure used here is that of Turek et al. (1977); briefly it is as follows. K X Rb and Sr peaks (20 angles) are measured at 37.9 and 35.8, background is measured at 41.4, 37.1, 35.0. Peaks are counted for 100 seconds and backgrounds for 40 seconds. Rb and Sr calculations are made as follows:

1. Background curvature (f) is found using pure quartz as a standard and measuring the intensities at Rb and Sr peak
angles:

\[ f_{Rb} = \frac{P_{q1}}{(B_1 + B_2)} \]

\[ f_{Sr} = \frac{P_{q2}}{(B_2 + B_3)} \]

where \( P_{q1} \) and \( P_{q2} \) are the dead time corrected count rates in counts per second (cps) and \( B_1, B_2 \) and \( B_3 \) are the dead time corrected background count rates (cps).

2. The net \( Rb \) and \( Sr \) counts are obtained using the previously derived factors as per equations:

\[ Rb_{net} = Rb_{cps} - f_{Rb} ((B_1)_{cps} + (B_2)_{cps}) \]

\[ Sr_{net} = Sr_{cps} - f_{Sr} ((B_2)_{cps} + (B_3)_{cps}) \]

3. The reciprocal values of dead time corrected Compton scatter count rates are plotted against the mass absorption of standards. The mass absorption of the unknown sample \((MA_{sa})\) is determined from the slope

\[ (MA)_{sa} = \frac{\left( \frac{1}{cps} \right) cs - b}{m} \]

where \( cs \) is Compton scatter, \( m \) the slope, and \( b \) the Y-intercept.

4. The corrected peak intensities are converted to concentrations using GSP-1 as a standard as per the equations below:

\[ Rb \text{ ppm} = \frac{(Rb \text{ ppm})_{standard} \cdot (MA)_{sa}}{(MA)_{standard}} \]
\[
\text{Sr ppm} = \frac{(\text{Sr ppm})_{\text{standard}} \times (\text{MA})_{\text{sa}}}{(\text{MA})_{\text{standard}}}
\]

The values used for GSP-1 were 252 ppm Rb and 235 ppm Sr.

The Rb/Sr weight ratio obtained by the above procedure can be converted to \(^{87}\text{Rb}/^{86}\text{Sr} \) atomic ratio by the formula:

\[
^{87}\text{Rb}/^{86}\text{Sr} = (\text{Rb/Sr})(0.2832)(9.514 + ^{87}\text{Sr}/^{86}\text{Sr})
\]

**Sample Dissolution**

**Rb-Sr**

Whole rock samples were dissolved in Pt dishes using 48% hydrofluoric acid (HF) and 70% perchloric acid (HClO₄). For an unspiked run (i.e. to determine natural \(^{87}\text{Sr}/^{86}\text{Sr} \) ratio) about 0.5 g of sample is taken (weighing is unnecessary). For a spiked run used to determine concentration of \(^{87}\text{Rb}, ^{86}\text{Sr} \) and the \(^{87}\text{Sr}/^{86}\text{Sr} \) ratio, the sample is weighed out accurately, between 0.25 g to 0.5 g depending on the estimated concentrations of Rb and Sr. To the sample in the Pt dish is added 10 ml HF and 10 ml HClO₄; this is evaporated to dryness. Another 10 ml HF and 10 ml HClO₄ is added, taken to dryness, and baked at full hot plate setting to drive off fluorine. The residue is taken up in 20 ml 2\(\text{N} \) HCl. This is warmed on the hot plate and centrifuged. The supernatant is loaded onto an ion exchange column. The ion exchange column contains about 19 cm³ of cation exchange resin Dowex 400W - X8 200-400 mesh. Prior to sample loading resin is cleaned with 6\(\text{N} \) HCl and conditioned with 2\(\frac{1}{2} \text{ N} \) HCl. The sample is loaded onto the column and allowed to permeate the resin; 50 ml of 2\(\frac{1}{2} \text{ N} \) HCl is added and discarded. An additional 30 ml of 2\(\frac{1}{2} \text{ N} \) HCl is added and collected. This fraction contains most of the Sr from the
sample. The eluted Sr is taken to dryness. As there is an overlap in the elution of Rb and Sr, it was necessary to repeat the ion exchange separation. Moreover, in some samples with high Rb, Rb was coprecipitated with K as a perchlorate by adding 2 drops of concentrated HClO₄ before ion exchange separation.

For 12 samples, Rb and Sr concentrations were determined by isotope dilution as well as by XRF. In this case, a mixed ⁸⁷Rb and ⁸⁴Sr tracer was weighed out in a syringe and added to the sample in the Pt dish. The ion exchange separation was as described above, except that the Rb cut was taken at 30-60 ml prior to the Sr cut at 60-90 ml.

**U-Pb**

Approximately 5 to 10 mg of zircons from a sized magnetic fraction was dissolved in a Teflon pressure bomb using ultrapure redistilled acids; 3 ml of 48% HF and 2 drops of concentrated HNO₃ following the procedure of Krogh (1973). The pressure bomb is placed in an oven at 180°C for 5 days. The dissolved zircon solution was evaporated to dryness and taken up in 3 N HCl. An aliquot was taken, weighed and spiked with a mixed ²³⁵U and ²⁰⁸Pb tracer solution. The remaining solution was designated for unspiked isotopic ratio measurement. Both portions of the sample were evaporated to dryness and taken up in 3 N HCl. They were then loaded on 5 cm³ ion exchange columns containing 1 ml resin. Zr, Th and the REE were eluted with 1 ml 3 N HCl and discarded. Pb was eluted with 1 ml 6 N HCl; U was eluted with 1 ml H₂O. The samples were then evaporated to dryness.

**Leaches**

Four zircon fractions from four rocks were leached with acid. The
purpose of leaching is to remove alteration products and is thus a
beneficiation procedure. However, this procedure is not fully under-
stood; while, it appears to work in some cases, in others it does not
work and the isotopic equilibrium is disturbed. The leaching pro-
cedure used in this study was done by adding 50 ml of 48% HF to
about 50 mg of zircon crystals, and agitating them for 30 minutes in an
ultrasonic cleaner. The undissolved zircon residues were then dissolved
in the routine procedure as described above.

**Mass Spectrometry**

Isotope ratios were measured on a Neir type 23 cm (9") 60° sector
solid source mass spectrometer at the University of Kansas. This
instrument uses a single filament for ionization; the collector is a
Faraday cup. The isotope ratios are obtained by magnetic field
switching. The accelerating voltage used was 4,870 volts for Rb and
Sr, 3,590 volts for U and 4,650 volts for Pb. The peak switching and
data acquisition is done automatically by an Hewlett Packard desk top
computer which is on line with the mass spectrometer. The background
readings are taken for 2 or 3 seconds and peak heights for 7 to 10 sec-
onds. A settle time of 4 or 5 seconds is allowed for magnet histeresis
between all peak and background readings. Typically the ratios were
measured in blocks of 10 until the error distributed about the means of
these blocks was reduced to a minimum. Before running the samples, the
system was pumped down for 3 hours to a vacuum of less than 1.3 × 10⁻⁷ Pa.

**Rb-Sr**

The Rb-Sr samples which were in the chloride form after ion exchange,
were converted to a perchlorate and sulfate respectively. The samples
were then dried. The filament used is outgassed Ta, oxidized to Ta$_2$O$_5$

to form a porous base for consistent ion emission.

There are three possible Rb-Sr runs: The Rb run, the spiked Sr run,
and the unspiked Sr run. For a Rb run by isotope dilution, the $^{85}$Rb/$^{87}$Rb
ratio is measured. Possible interference from Sr is monitored on mass
88. In a spiked run the ratios measured are $^{88}$Sr/$^{86}$Sr, $^{87}$Sr/$^{86}$Sr and
$^{84}$Sr/$^{86}$Sr. Background is counted on masses 86.6 and 87.5. The pre-

cence of any Rb is monitored by mass 85. The $^{88}$Sr/$^{86}$Sr ratio is used
to correct for mass fractionation during a run. The $^{86}$Sr/$^{88}$Sr is calcu-
lated and normalized to 0.1194. This correction is then applied to the
other two ratios. Isotope composition on a spiked Sr run is made as
stated above, except $^{84}$Sr/$^{86}$Sr is not measured. This involves
measuring $^{88}$Sr/$^{86}$Sr ratios; $^{87}$Sr/$^{86}$Sr is normalized to $^{86}$Sr/$^{88}$Sr =
0.1194.

U-Pb

U and Pb samples were dissolved in H$_3$PO$_4$ and loaded on a Re
filament which had previously been coated with silica gel to act as a
binder. Mass fractionation corrections on both spiked and unspiked
samples were adjusted by 0.1% per mass unit – an empirically estab-
lished constant.

For a spiked run, U and Pb ratios are measured on the same fila-
ment. Pb ratios are obtained first, then the filament current is
increased for U ratio measurement. The $^{208}$Rb/$^{206}$Pb is measured
with background readings taken at mass 205. U ionizes as UO$_2^+$, therefore,
the actual masses counted are 254 (i.e. 238 + 16) and 251 (i.e. 235 + 16).
Background counts are measured at mass 249 (i.e. 233 + 16).

For an unspiked run, which is also referred to as an isotopic
composition measurement, $^{208}\text{Pb}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ are obtained first with zeros taken at mass 205. $^{204}\text{Pb}/^{206}\text{Pb}$ is measured with background counts taken on masses 204.5 and 203.5. Since this ratio is very small ($10^{-3}$ to $10^{-4}$), the peaks are counted for 15 seconds.
DISCUSSION OF ERRORS

Errors affecting determined values can be considered to originate from two sources and may be grouped as geologic errors and experimental errors. Geologic errors are difficult to control and are dependant upon the geologic history of the samples. The treatment of geologic error, when recognized, is discussed later. Experimental errors may be random or systematic. Random errors have a normal distribution and affect precision only. Systematic errors are non-normally distributed and affect accuracy.

Random Errors

Splitting Errors

These are errors that occur when a subsample taken is not completely representative of the larger sample. These errors were avoided by the use of conventional splitting techniques when subsampling. These errors can be tested by duplicate sampling. Sampling errors will simply slide a point up or down the isochron or discordia line and do not affect the age determination.

Weighing Errors

Weighing errors arise when the sample and the spike solution are weighed. For unspiked Sr samples weighing is unnecessary. For U-Pb analysis, weights of samples and spike solutions were cross checked and agree to within ± 0.1%. Quantitative error no longer affects the results after the sample and tracer are combined and homogenized since thereafter only isotopic ratios are measured.

In Rb-Sr analysis, error in sample weight is unimportant and cancels out when the concentrations are converted to ratios, but
an error in the tracer solution weight will cause the measured $^{87}\text{Rb}/^{86}\text{Sr}$ ratio to be displaced parallel to the X-axis of the isochron diagram and will cause variations in calculated Rb and Sr concentrations. In U-Pb analysis, weighing errors will cause displacement of a point along the discordancy line and slight variations in measured U and Pb concentrations.

**Contamination Errors**

These errors are the result of contamination of the Rb-Sr or U-Pb analyses with common Sr or common Pb respectively. Blank values for Rb and Sr were not taken as it was thought that they would have little or no effect on the Rb-Sr ages compared to other parameters. Sr contamination simply slides a point along the isochron and does not affect the age. For U-Pb analysis blank values measured during the course of this study were 0.001 $\mu$g $^{206}\text{Pb}$ and 0.001 $\mu$g $^{238}\text{U}$. Values for the isotopic composition of the blank were $^{208}\text{Pb}/^{204}\text{Pb} = 39.0$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.8$ and $^{206}\text{Pb}/^{204}\text{Pb} = 18.6$ which is the modern composition of common Pb. Values used for the initial isotopic composition of common Pb were $^{208}\text{Pb}/^{204}\text{Pb} = 33.37$, $^{207}\text{Pb}/^{204}\text{Pb} = 14.69$ and $^{206}\text{Pb}/^{204}\text{Pb} = 13.64$, which are those defined by the two-stage terrestrial Pb evolution model of Stacey and Kramers (1975).

**Mass Spectrometry Errors**

The ratios measured on the mass spectrometer deviate slightly from their actual values because of mass fractionation and mass spectrometer discrimination. During a Sr run, fractionation can be corrected for by normalizing the sample $^{86}\text{Sr}/^{88}\text{Sr}$ to 0.1194, the
invariant natural ratio of these isotopes. Rb has only two isotopes and therefore cannot be normalized to correct for fractionation. A Pb fractionation error of 0.1% per atomic mass unit was used to correct Pb data. This is based on repeated runs of the NBS Pb standard.

The precision of the mass spectrometer is evaluated by the periodic measurement of standard samples. This is also a measurement of the accuracy of the calibration of the mass spectrometer. During the course of this study replicate laboratory analysis of the NBS Sr standard averaged 0.7101 ± 0.0001 for the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio. One analysis of the NBS radiogenic Pb standard gave values of 0.00037 ± 0.00001, 0.07117 ± 0.00005 and 0.01361 ± 0.00002 (1σ) for the $^{204}\text{Pb}/^{206}\text{Pb}$, $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{208}\text{Pb}/^{206}\text{Pb}$ ratios respectively. Values obtained for both these standards are in agreement with their recommended values.

**Systematic Errors**

Systematic errors affecting the data in this study are the result of the uncertainty of the decay constants (λ) of radioactive isotopes, the uncertainty in the calibration of standards, and the uncertainty in the isotopic compositions of non-radiogenic Sr and Pb.

The uncertainty in the $^{87}\text{Rb}$ decay constant arises from the difficulty in counting the low energy $\beta$ particles. The variation may be as large as ±5% but will not affect age relationships between different laboratories if a consistent value for λ is used. It is important when comparing the results of two age determinations such as U-Pb and Rb-Sr. The uncertainty in the $^{238}\text{U}$ and $^{235}\text{U}$ half-lives is about ±0.1%. This translates into an accuracy error of
+3 Ma for U-Pb ages in the Archean. Decay constant uncertainties affect only absolute ages and cause problems in comparison of ages obtained by different decay schemes, but within each scheme the ages are correct relatively.

The values used in this study for the natural composition of Rb, Sr, U and Pb are those that are now widely accepted in geochronology. In this study the natural isotopic constants used are those recommended by the International Union of Geological Sciences Submission of Geochronology (Steiger and Jager, 1977) and are:

**Atomic Ratios**

\[
\begin{align*}
^{85}\text{Rb}/^{87}\text{Rb} &= 2.59265 \\
^{86}\text{Sr}/^{88}\text{Sr} &= 0.1194 \\
^{238}\text{U}/^{235}\text{U} &= 137.88
\end{align*}
\]

**Decay Constants**

\[
\begin{align*}
^{87}\text{Rb} &= 1.42 \times 10^{-11} \text{ yr}^{-1} \\
^{238}\text{U} &= 1.55125 \times 10^{-10} \text{ yr}^{-1} \\
^{235}\text{U} &= 9.8485 \times 10^{-10} \text{ yr}^{-1}
\end{align*}
\]

**Statistical Treatment of Data**

Age data obtained in this investigation were interpreted in the form of an isochron plot for Rb-Sr, and a concordia plot for U-Pb. In constructing these plots it is necessary to find the best straight line on a plot of Y versus X when both Y and X are subject to experimental errors. This involves the use of a two error linear regression technique. This technique must allow for known experimental error in both X and Y coordinates and must weight data to allow for nonuniform variance in the X coordinate. Such a regression
technique, while providing a realistic age measurement, must also
give a measurement of the uncertainty of the age which is equally
important in interpreting geologic data.

Isochron Regression

Four two-error linear regressions for isochrons have been
published, that of McIntyre et al. (1966), York (1966), York (1969)
and Brooks et al. (1968). A comparison of these treatments by Brooks
et al. (1972) and Birk and McNutt (1981) revealed that provided all
data scatter about the line is attributable to experimental error,
all treatments yield identical regression parameters and hence
identical age intercepts and uncertainties. This study uses the
treatment of McIntyre et al. (1966) and is outlined briefly below.
The calculation is done by a computer using a program written in
Fortran.

The McIntyre regression treatment is based on an equation for
the slope of the line, that on substitution of a preliminary
estimate of the slope, reduces to a cubic equation. The X and Y
coordinates are assigned constant weighting factors which are prior
estimates of the variances of the coordinates as determined by
replicate analysis. A constant absolute error in Y results in a
decreasing percentage for increased Y (\(^{87}\text{Sr}/^{86}\text{Sr}\)). For evaluating
whether or not a scatter of data points about a regressed isochron
is within the assigned limits, an index is calculated. This index,
a calculated mean square of weighted deviates (MSWD), is compared
to a theoretical value based on "Student's t" distribution. If
the MSWD is less than the theoretical value, the data are within
experimental error. If the MSWD exceeds the theoretical value then the source of scatter is attributable to geological error the magnitude of which is a reflection of the magnitude of the index. If the isochron cannot be fitted within experimental error (that is MSWD >> 1), then the isochron parameters are calculated using three geologic error models. The first geologic error model regresses the data on the assumption that the excess variance in the $^{87}$Sr/$^{86}$Sr is distributed proportionally to $^{87}$Rb/$^{86}$Sr. This model is considered applicable to rocks that have undergone open system behaviour. A second geologic error model assumes that the variation in $^{87}$Sr/$^{86}$Sr is independent of $^{87}$Rb/$^{86}$Sr. This model is considered applicable to samples that crystallized with different initial $^{87}$Sr/$^{86}$Sr values. The suitability of either of these models is determined by testing the trend of the Y residuals as a function of X. If neither of these two models is suitable, a third geological error model is invoked, in which the scatter is assumed to be a mixture of the first and second type distributions.

The nature of errors in Rb-Sr isochrons is such that the error in Y is essentially constant, while the error in X increases with the magnitude of X. The regression used in this study allows for this distribution of errors, hence, a point high up on the isochron does not carry as much weight as a point lower on the isochron, which is better determined. The Rb-Sr age is proportional to the slope, and therefore the error of the slope is the error in the age.

The errors, as variances estimated for Rb-Sr data are: 144.73 $\times$ 10$^{-6}$ for $^{87}$Rb/$^{86}$Sr and 0.220 $\times$ 10$^{-6}$ for $^{87}$Sr/$^{86}$Sr. This is roughly 1.2% and 0.07% respectively at the 95% confidence.
level. All uncertainties in Rb-Sr ages and initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are quoted at the 95% confidence level.

Concordia Regressions

On the U-Pb concordia, the ages are determined from the intercepts of the discordancy line with the concordia curve. The intercept ages are calculated by a computer program which first determines the parameters of the discordancy line using a York (1966) least squares regression and then finds the intercepts by iteration. The errors associated with the concordia intercept ages are based on the intersection of hypothetical cords with concordia having slopes equal to the slope of the original cord plus or minus its standard error. Error limits for U-Pb determinations at the isotope geochemistry laboratory at the University of Kansas are $^{207}\text{Pb}/^{206}\text{Pb} \pm 0.20\%$, $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U} \pm 1.0\%$ at the 95% confidence level.

Precision error estimates used for the error envelopes in Figs. 9, 11b, and 13b were obtained by taking plus or minus two times the standard error of the measured $^{204}\text{Pb}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios:

$$SE' = \sqrt{\frac{\sum_{i=1}^{N} (\bar{X} - X_i)^2}{N(N-1)}}$$

where $\bar{X}$ is the mean of $N$ values (usually ten) and $X_i$ = the $i$th value.
Errors in U-Pb data on concordia diagrams are unlike those of the Rb-Sr isochrons. Because of the way \( X (^{207}\text{Pb} / ^{235}\text{U}) \) and \( Y (^{206}\text{Pb} / ^{238}\text{U}) \) are measured, the error of an individual datum point is essentially ellipsoidal with the long axis subparallel to the discordancy line. Furthermore, there is no systematic increase of the error of a point with increasing X or Y coordinate, as is the case with Rb-Sr isochrons. In U-Pb age work the York (1966) treatment regresses the data points with allowance made for the individual errors in each datum point. There are no problems of a statistical nature in fitting a line to discordant points. However, a fundamental problem exists in using the error of the line as an estimate of the error in the age. The problem is that a line goes to infinity while a U-Pb age is the intersection of the discordancy line with the concordia curve. That is, the discordancy line is not mathematically a line, but is a finite length chord. As an example of the problem - if there are three evenly distributed points on a discordancy line, the error of the regression will be very small, and hence, the intercept error will also be very small. On the other hand, if the three points are very close to the concordia, the error of the discordancy line will be very large. Clearly this treatment is improper. The more concordant the data the greater should be the certainty in the age. Unfortunately, at the present time, there is no statistical treatment that takes these problems into account. For this reason in this study the error is estimated from the intercept of the slope of the discordancy line plus or minus the standard error of the mean. This treatment is rather conservative. The statistical treatment of U-Pb concordia data has not reached the level of sophistication achieved for Rb-Sr data.
Testing Differences Between Ages

When ages are calculated it is of prime importance to know whether or not a real or significant difference exists between them. Ages may be statistically distinct even though their confidence intervals overlap. Differences between ages and intercepts can be tested by the following method (Turek, 1966).

A test statistic \( t_{\text{calc}} \) is calculated for the two ages in question from their slopes \( m_1 \) and \( m_2 \), and standard errors \( SE_1 \) and \( SE_2 \):

\[
t_{\text{calc}} = \frac{m_1 - m_2}{\left( (SE_1)^2 + (SE_2)^2 \right)^{\frac{1}{2}}}
\]

The value of \( t_{\text{calc}} \) is then compared to a \( t \) value from a "students - t" distribution with \( K \) degrees of freedom, where \( K \) is found by the relation:

\[
\left( \frac{(SE_1)^2 + (SE_2)^2}{K} \right)^2 = \frac{(SE_1)^4}{N_1 - 2} + \frac{(SE_2)^4}{N_2 - 2}
\]

where \( N_1 \) and \( N_2 \) represent the number of samples comprising age 1 and age 2 respectively. If \( t_{\text{calc}} > t_K \) then age 1 and age 2 are significantly different. Age differences in this study were tested at the 95% confidence level.
ANALYTICAL RESULTS

Approximately 120 samples were collected for possible age determination by Rb-Sr and U-Pb methods. Sampling was designed to provide material suitable for age determinations from the oldest to the youngest rocks in the Michipicoten volcanic belt and also to obtain ages for the granitic rocks that intrude the greenstone and associated metasedimentary rocks. The limitation to this sampling scheme was the accessibility of outcrops, their precise stratigraphic position and also their technical feasibility for dating.

This thesis reports 35 Rb-Sr ages and 26 zircon ages from the following units:

1. Acid volcanics at the top of the lower volcanic sequence.
2. Acid volcanics near the bottom of the upper volcanic sequence.
3. Granitic rocks which are physically within the greenstone belt, i.e. internal granites, which intrude the lower volcanic sequence.
4. Granitic rocks which border the greenstone belt, i.e. an external granite.

For Rb-Sr ages samples of about 2 kg were collected, crushed, pulverized and analysed for their Rb and Sr content to determine their suitability in terms of their Rb/Sr ratios for age determination. Of the 120 samples collected, 35 were deemed suitable for age determination. For U-Pb ages on zircons there is no preliminary suitability test. There is no practical method to ascertain the presence or absence of zircons nor their quality or quantity in a given rock.
Fortunately, in this study all 5 samples proved to contain plentiful and high quality zircons.

All zircons measured indicate at least a two stage U-Pb evolutionary history and plot below the concordia curve. They have either lost Pb, or gained U. The lower intercepts on the concordia plots range from 100 Ma to 400 Ma for the felsic volcanic rocks and 350 Ma to 500 Ma for the granitic rocks. Such lower intercepts are consistent with other data reported in the literature for the Superior province. Wetherill (1956) considers this to represent an episodic loss of Pb (or gain of U); however, a major orogenic event in the Paleozoic era has not been demonstrated. On the other hand Goldich and Mudrey (1972) suggest uplift of the shield as the possible Pb loss event for the Superior province. Evidence for recent crustal uplift is seen in scattered patches of Paleozoic sediments in the area and the lower intercept may be reflecting this uplift, possibly combined with a superimposed continuous diffusion of Pb (Tilton, 1960) since crystallization of zircons.

The location of the samples used in the study is shown in Fig. 2. The analytical data is given in Tables 1, 2 and 3, and illustrated as isochron and concordia diagrams in Fig. 4 to 14.

Lower Volcanics

Helen Mine Fragmental Volcanics

Samples 659 and 660 were collected from the Helen Mine property about 2 km north of Wawa. The rocks are overturned to the north about an east-west trending fold axis and dip steeply to the south and face north. These are the mine footwall volcanic rocks and occupy a stratigraphic position just below the Helen iron range. Sample 659 was
<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Rb (ppm)</th>
<th>Sr (ppm)</th>
<th>$^{87}\text{Rb}/^{86}\text{Sr}$</th>
<th>$^{87}\text{Sr}/^{86}\text{Sr}$</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Acid Volcanic</strong>&lt;sup&gt;*&lt;/sup&gt;</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Quartz - Feldspar Crystal Tuff</td>
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<td></td>
<td></td>
<td></td>
</tr>
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<td>636(1)</td>
<td>44.83</td>
<td>188.2</td>
<td>0.690</td>
<td>0.7280</td>
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<td>636(4)</td>
<td>50.83</td>
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<td>0.802</td>
<td>0.7316</td>
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<td>636(3)</td>
<td>49.66</td>
<td>177.9</td>
<td>0.810</td>
<td>0.7335</td>
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<td>636(6)</td>
<td>54.59</td>
<td>149.3</td>
<td>1.062</td>
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<td>636(10)</td>
<td>58.30</td>
<td>137.3</td>
<td>1.234</td>
<td>0.7513</td>
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<tr>
<td>636(7)</td>
<td>61.79</td>
<td>131.5</td>
<td>1.365</td>
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<td><strong>Acid Volcanic - Wawa Tuff</strong>&lt;sup&gt;*&lt;/sup&gt;</td>
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<td>643(4)</td>
<td>74.98</td>
<td>67.77</td>
<td>3.259</td>
<td>0.8336</td>
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<td>643(7)</td>
<td>63.62</td>
<td>44.89</td>
<td>4.165</td>
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<td>66.13</td>
<td>42.54</td>
<td>4.572</td>
<td>0.8764</td>
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<tr>
<td>643(2)</td>
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<td>37.15</td>
<td>7.501</td>
<td>0.9793</td>
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<tr>
<td>643(6)</td>
<td>112.5</td>
<td>35.74</td>
<td>9.393</td>
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<td>643(5)</td>
<td>110.4</td>
<td>13.65</td>
<td>25.339</td>
<td>1.5597</td>
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<td><strong>Acid Volcanic I - Helen Mine</strong></td>
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<tr>
<td>659(5)</td>
<td>13.71</td>
<td>149.3</td>
<td>0.266</td>
<td>0.7129</td>
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<tr>
<td>660(1)</td>
<td>57.20</td>
<td>299.6</td>
<td>0.554</td>
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<td>660(2)</td>
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<td>227.9</td>
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<td>0.7347</td>
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<td>66.25</td>
<td>206.2</td>
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<td>0.7391</td>
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<td>2.220</td>
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<td><strong>Diorite - Jubilee Stock</strong></td>
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<tr>
<td>662(6)</td>
<td>28.61</td>
<td>288.8</td>
<td>0.287</td>
<td>0.7138</td>
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<td>662(3)</td>
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<td>0.531</td>
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<td>36.97</td>
<td>105.0</td>
<td>1.022</td>
<td>0.7400</td>
</tr>
<tr>
<td>Sample No.</td>
<td>Rb (ppm)</td>
<td>Sr (ppm)</td>
<td>$^{87}\text{Rb}/^{86}\text{Sr}$</td>
<td>$^{87}\text{Sr}/^{86}\text{Sr}$</td>
</tr>
<tr>
<td>------------</td>
<td>----------</td>
<td>----------</td>
<td>-------------------------------</td>
<td>-------------------------------</td>
</tr>
<tr>
<td>646 (1)</td>
<td>19.86</td>
<td>255.8</td>
<td>0.225</td>
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<tr>
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<td>180.0</td>
<td>0.641</td>
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<td>646 (2)</td>
<td>52.18</td>
<td>166.7</td>
<td>0.909</td>
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</tbody>
</table>

Trondhjemite - Hawk Lake

Note: For samples with *Rb and Sr ppm and $^{87}\text{Rb}/^{86}\text{Sr}$ were obtained by isotope dilution. For all other samples the Rb/Sr weight ratio was obtained by XRF, and was converted to an isotopic ratio by the formula:

$^{87}\text{Rb}/^{86}\text{Sr} = \text{(Rb/Sr) wt (0.2832)(9.514 +}^{87}\text{Sr}/^{86}\text{Sr)}$. 
Table 2. Analytical Data for Zircons from the Michipicoten Greenstone Belt

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Magnetism</th>
<th>Tyler Mesh Grain Size</th>
<th>Sample Weight (mg)</th>
<th>Concentration (ppm)</th>
<th>Pb isotopic composition (atom %)</th>
<th>Isotopic Ratios</th>
</tr>
</thead>
<tbody>
<tr>
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Notes: Magnetism, NM (non-magnetic) and M (magnetic), is a measure of magnetic susceptibility of zircons and indicates the horizontal inclination of the Franz Isodynamic Separator, field strength was 1.7 A. (L) indicates a leached fraction.
Table 3. U-Pb Age Data for Zircons from the Michipicoten Greenstone Belt

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<th>Sample No.</th>
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<th>Concordia Ages</th>
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<td>$^{206}\text{Pb}/^{238}\text{U}$</td>
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collected from an altered zone 45 m below the siderite member of the Helen iron formation. Sample 660 was collected south of 659, outside the altered zone. Sample 659 consists of a dense metamorphosed homogeneous dacite tuff. The original tuffaceous matrix has been recrystallized into sericite, epidote, calcite, and quartz. Sample 660 is more schistose and contains aligned flakes of biotite.

Seven samples of the meta-volcanics of this area gave a Rb-Sr whole rock isochron age of 2530 ± 90 Ma with an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7037 ± 0.0011 (Fig. 4). The error of regression is well within experimental error (MSWD = 0.05).

The 659 dacite was also sampled for zircon U-Pb analysis. Zircons from the unit are pastel brown or pink and clear. The U content of these zircons ranges from 218 to 309 ppm. Five zircon analyses from the unit define a discordancy line with an upper concordia intercept of 2749 ± 2 Ma and a lower intercept of 107 ± 20 Ma (Fig. 5). One of these analyses, a leached fraction is within analytical error of the concordia and has a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2750 Ma. The colinearity of the points, moderate discordance (maximum 30%), and uniform nature of the crystals indicate that only one primary igneous population of zircons is present. Thus, the age is considered to be a primary age. The Rb-Sr age therefore appears to give a younger age.

Wawa Tuff

Sample 643 was collected from a breccia dome located on the northwest shore of Wawa Lake. This rock is also a fragmental volcanic and is similar to the Helen Mine fragmental volcanics, in all probability these two sample sites represent two slightly different facies of a common volcanic episode. In thin section the rock shows quartz
Fig. 4. Isochron diagram for acid volcanic I unit - Helen Mine.
Fig. 5. Concordia diagram for acid volcanic I unit - Helen Mine. Closed circle indicates a leached analysis.
phenocrysts, quartz-calcite amygdules, and rock fragments set in a recrystallized quartz, sericite, and carbonate matrix.

Rocks in this unit are high in potassium and are enriched in radiogenic strontium. A six point isochron age for this unit yields a colinear array (MSWD = 2.26), with a slope corresponding to an age of 2285 ± 70 Ma and initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7275 ± 0.0052 (Fig. 6). This rock should give a much older age, it should at least agree with the Rb/Sr age obtained for the Helen Mine volcanics. The $^{87}\text{Sr}/^{86}\text{Sr}$ is exceptionally high, and therefore this isochron indicates a secondary age.

Quartz Feldspar Crystal Tuff

Sample 636 was collected from a shallowly dipping dacitic unit located on the southern shore of Wawa Lake. This unit belongs to the lower volcanics but appears to be stratigraphically younger than the steeply dipping Helen Mine fragmental volcanics and the Wawa tuff to the north (643, 659, 660) and is stratigraphically separated by an angular unconformity (Sage, personal communication). The rock is dark grey in colour and is characterized by large (1 cm) opalescent blue quartz eyes set in a fine grained quartz, feldspar, and calcite matrix. In thin section the quartz eyes are deeply embayed and are streaked with small opaque inclusions.

The Rb-Sr whole rock isochron, Fig. 7', shows considerable scatter of data points. The variance of regression is outside the limits of experimental error (MSWD = 5.26), therefore, the scatter of points is attributed to geological error. A geologic error model regression (assuming the variance in $^{87}\text{Sr}/^{86}\text{Sr}$ is distributed proportional to $^{87}\text{Rb}/^{86}\text{Sr}$) yields an age of 2680 ± 490 Ma with an intercept of 0.7014
Fig. 6. Isochron diagram for acid volcanic I unit - Wawa Tuff.
Fig. 7. Isochron diagram for acid volcanic II unit - Quartz feldspar crystal tuff.
0.0064. Most likely the scatter of the data points is the result of a younger thermal event. For example, if the 2284 Ma age registered by the Wawa tuff was widespread the indicated age here may be a mixed age. The U content for five zircon analyses range from 302 to 455 ppm and is similar to the Helen Mine zircons. A sixth analysis, from a -100 + 200 M + 1 fraction has an anomalously high U content, 903 ppm. Zircon analyses from 636 define a discordancy line with upper and lower intercepts of 2744 ± 10 Ma and 364 ± 126 Ma respectively (Fig. 8f). The uncertainty in the age is somewhat larger than reported for the previous volcanic horizon; this in part is a result of the closely spaced disposition on the discordancy line.

Upper Volcanics
Sample 641 is a felsic metavolcanic rock collected from a roadcut near Catfish Lake located in the western area of the belt. This unit overlies the Dore sediments and according to Attoh (1980) belongs to the upper volcanics. Rocks of this vicinity are moderately sheared, a phenomenon which is probably related to the emplacement of the granitic batholith a few kilometers to the west.

No Rb-Sr ages are reported for this unit. Zircons from this unit are transparent to pink and are easily distinguished from the honey-brown zircons of the lower volcanic units (Plate 1). However, in terms of U content there is no discernable difference from 659 (Table 2). Five zircon analyses from this unit yield a tightly colinear discordancy line with upper and lower concordia intercepts at 2696 ± 2 Ma and 385 ± 30 Ma respectively (Fig. 9). One of these analyses, a leached fraction, plots above the concordia curve. As
Fig. 9. Concordia diagram for acid volcanic III - Catfish Lake. Closed circle indicates a leached analysis.
the zircons are homogeneous and free from overgrowths or xenocrystic material there is little doubt that the upper intercept age represents the time of primary magmatic crystallization.

The U-Pb zircon analyses with their analytical uncertainties for all three volcanic units are plotted on a single concordia diagram in Fig. 10. It is clear that all points including the leached fractions are within analytical error of their discordancy lines. Furthermore, the sequence of ages for the volcanics agrees with their relative stratigraphic position, although the 2744 ± 10 Ma age for the crystal tuff (636) is not significantly different from the 2749 ± 2 Ma age for the Helen mine unit (659). However, a statistically significant time interval of 53 Ma separates the Helen mine unit from the upper volcanic unit (641) dated at 2696 ± 2 Ma.

Internal Granites

Jubilee Stock

Samples from the Jubilee stock are 662 and 663. Sample 662 consists of diorite porphyry and was collected in the vicinity of Minto Lake. Sample 663 is finer grained and was collected about 1 km north of Minto Lake. In thin section the feldspars from these rocks are zoisitized and partly sericitized. Other secondary minerals include carbonate and sulfides such as pyrite, pyrrhotite and chalcopyrite. Possibly, these secondary minerals are the result of pneumatolytic-hydrothermal reactions which occurred during the late stages of magmatic crystallization.

A five point Rb-Sr isochron for the stock gives an age of 2560 ± 270 Ma with initial ratio of 0.7034 ± 0.0024. The rather large error quoted is primarily a consequence of the lack of an adequate
Fig. 10. Concordia diagram for all volcanics. Error envelopes are analytical precision error estimates.
range of Rb/Sr ratios to define a good line. Despite the fact that the
663 sample was collected 1 km apart from sample 662 and lithologically
they are slightly different, isotopically the points are colinear as
shown in isochron diagram Fig. 11, indicating that the stock was
isotopically homogenous over a wide area. The age of the stock is the
same within experimental error as the Rb-Sr age obtained for the 659
samples at the Helen Mine. This age is considered to represent the
time that the rocks became a closed system to Rb and Sr migration.
This unit was not sampled for zircons.

Granodiorite Stock

Sample 664 was collected from the vicinity of the Norfolk
gold mine about 2.5 km south of the Jubilee Stock. The granodiorite is
characterized by very highly altered plagioclase laths and large
opalescent blue quartz grains similar to that of sample 636.
Unfortunately, the lack of variability in the Rb/Sr ratios for the
samples collected from this stock, preclude Rb-Sr dating.

Zircons from the unit are brown to clear, distinctly zoned,
euhedral and prismatic (Plate 2). Crystals from the larger grain
sizes are moderately fractured. The U contents of these zircons,
range from 279 to 1312 ppm and show a much broader range of values
than those of the volcanic rocks. Four analyses from the unit are
20% to 60% discordant and form a least squares discordancy line
intercepting the concordia curve at 2737 \pm 6 Ma and 345 \pm 19 Ma
(Fig. 12ab). A fifth analysis, a leached -100 + 200 mesh nonmagnetic
fraction, plots well above the concordia and to the right of the
extrapolated discordancy line indicating an older age. It plots
outside experimental error and is not included in the regression.
The position of this point with respect to the unleached parent
Fig. 11. Isochron diagram for the Jubilee stock.
Fig. 12a. Concordia diagram for granodiorite stock.
Closed circle indicates a leached analysis.
Fig. 12b. Concordia diagram for granodiorite stock showing analytical error envelopes.
fractions is a consequence of preferential dissolution of large quantities of U from radiation damaged areas of zircon crystals together with a disturbance of the isotopic equilibrium. A 50 mg portion of crystals was reduced to 5 mg in 15 minutes, demonstrating the extreme solubility of the altered areas.

Both this stock and the Jubilee stock are considered as cogenetic and coeval, being emplaced 2737 Ma ago, which is an age very similar to that obtained for the felsic extrusives of the lower volcanics.

External Granite

Hawk Lake Trondhjemite

The Hawk Lake trondhjemite is part of a large batholithic terrain, which embays the greenstone belt on the south and separates the Michipicoten from the Gamitagama belts. For this study, sample 646 was collected at the northwest corner of Hawk Lake, approximately 3 km from the supracrustal rocks. It is a medium grained, slightly foliated trondhjemite. The primary minerals are better preserved than those of the granitic stocks previously described. No calcite or sulfides were observed.

A six point Rb-Sr isochron for the trondhjemite is within experimental error (MSWD = 1.13) and yields an age of 2550 ± 175 Ma with initial ratio of 0.7048 ± 0.0014 (Fig. 13). The rather large uncertainty in the age is related to the limited spread in Rb/Sr ratios. The age and initial ratio are statistically indistinguishable from the Rb-Sr age obtained for the Jubilee Stock.

Zircons from the trondhjemite are typical of a granitic rock in that they are euhedral, zoned, and have high U contents, ranging from 1113 to 2002 ppm. Five zircon analyses are shown on the concordia
diagram in Fig. 14ab. The analysed fractions are from 25% to 60%
discordant; one analysis, a leach of the -100 + 200 MNM -1 fraction
is also plotted and lies well above and to the right of the concordia
in a fashion similar to the leached 664 fraction. A least squares
fit of fine grained -200 + 350 M 0° fraction, about 25% discordant,
with the two most discordant points (also fine grained) yields an age
of 2747 ± 6 Ma with lower intercept of 462 ± 14 Ma. This age is essen-
tially coeval with the 664 granodiorite stock. However, if the
-200 + 350 M 0 analysis is excluded, a regression of four zircon
analyses, including the leach, defines a discordancy line with
concordia intercepts of 2812 ± 3 Ma and 573 ± 8 Ma. Thus, it
would appear that the coarser grained zircons include an age older
than the finer grained zircons.

In view of the possible ambiguity in age, the 646 zircons were
examined microscopically under immersion oils to look for evidence
supporting the existence of two populations of zircons. Various
size and magnetic fractions are shown in Plates 3, 4 and 5. The
crystals are characterized by numerous fractures, especially the more
magnetic fractions. This is probably related to radiation damage
since an increase in U is paralleled by an increase in magnetic
susceptibility. Dark masses are common in the larger size fractions
and in some instances appear to be rounded older zircons which have
been overgrown by a younger zircon population.

The presence of older xenocrystic zircons would explain the
tendency of the coarser fractions to define older discordancy tra-
jectories. If this is so, then the 2812 Ma age is an age generated
by the mixing of two stage and three stage leads. Consequently, one
Fig. 14a. Concordia diagram for the Hawk Lake trondhjemite. Closed circle indicates a leached analysis.
Fig. 146. Concordia diagram for the Hawk Lake trondhjemite showing analytical error envelopes.
might expect that a leach of such a mix of primary and inherited zircons would plot off the discordancy line—presumably enriched or depleted in one of the populations. However, the residue of the leach is collinear with the original unleached point.

Chemical transitions that take place during the leaching of zircons are poorly understood. Leaching experiments by Krogh and Davis (1974, 1975) revealed that discordant zircons contain altered material that is extremely soluble in HF and that the unaltered residue remains a closed system to leaching. Analysis of the residue of the -100 + 200 fraction indicated that HF leaching has decreased the U content from 1113 ppm (unleached fraction) to 284 ppm, a significant reduction compared to leaches of other analyses, especially those from the volcanics (Table 2). Zircons are inhomogenous with regard to U concentration and it is evident that the radiation damaged (high U) zones have been preferentially leached from these crystals. The residual crystals are clearer and in some places appear somewhat skeletal (Plate 6). It is not known, however, whether or not the U-Pb system of these zircons was disturbed by the selective removal of U or Pb from the unaltered residue. Considering the quantity (about 95%) of material that was dissolved, this possibility must not be disregarded. Furthermore, the mechanism causing the 664 leach—also granitic zircons—to plot to the right of its discordancy line may have been operative on the 646 leach. If this is so, then the collinearity using the leached point may be fortuitous and hence misleading.

Therefore, I believe that the 2747 Ma age must be accepted as the best age estimate for this rock; however, the possibility that at
least in part some of the rocks in this batholith have a prehistory and are older than 2812 is conceded.
DISCUSSION

Zircon ages reported here are in agreement with relative ages based on the detailed stratigraphic mapping in progress by Attoh (1980) and Sage (1979, 1980). The ages determined are shown in relation to their stratigraphic position in Figure 15. Both the Rb-Sr and U-Pb zircon ages are shown diagramatically in Fig 16. It is apparent that the zircon ages have a very narrow range and very low uncertainties. On the other hand, the Rb-Sr data is systematically younger by about 8-10% and the uncertainties in ages are very large. The results here are in marked contrast to the Rb-Sr and U-Pb ages for the Gamitagama area (Turek et al., 1981; Krogh and Turek, 1981) where both systems' ages agree within analytical and decay constant uncertainties.

The successful determination of zircon ages for the upper and lower volcanic units of the Michipicoten belt has indicated that the belt evolved over a time period in excess of 53 Ma. This must be considered a minimum age since older mafic non zircon bearing strata underlies the lowest unit dated and overlies the youngest unit dated.

The 2696 Ma age for the upper volcanic unit is coeval with a 2703 Ma reported by Nunes and Pike (1980), and a 2703 Ma age reported by Nunes and Jensen (1980), both from upper volcanic cycles in the Abitibi belt. Also, an upper volcanic age of 2704 Ma is reported from the Wabigoon subprovince west (Davis et al., 1980). These ages are the youngest U-Pb ages of volcanics in greenstone belts of Ontario and probably reflect the cessation of Archean
Fig. 15. Generalized geochronology (this study), and stratigraphic sequence for the Michipicoten belt (based on Goodwin, 1962 and Attoh, 1980).
Fig. 16. Summary of Rb-Sr and U-Pb ages with respective error bars, Michipicoten belt.
volcanism in the Superior province.

The middle volcanics have not been dated in this study. However, considering the upper and lower ages obtained, a 2713 Ma age for a middle volcanic rhyolite in the Gamitagama area (Krogh and Turek, 1981) appears to correspond favourably to – and thus is a reasonable approximation of – the middle volcanics of the Michipicoten belt.

Goodwin (1964) extensively discussed structural and geochemical evidence that the acid volcanics and intrusive stocks of the belt are cogenetic. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios obtained for the volcanics and the Jubilee stock, 0.7037 and 0.7034 respectively, are statistically indistinguishable and further support this hypothesis. Ages obtained in this study prove that the extrusive volcanics and intrusive stocks are coeval.

The Hawk Lake trondhjemite, although perhaps genetically distinct, also shows evidence of intrusion coeval with the stocks. Although coarse grained zircons indicate a U-Pb age of greater than 2812 Ma, contact relations preclude the trondhjemite being older than the supracrustals, at least for that part of the batholith sampled in this study. It is possible, however, that this trondhjemite contains rafts or enclaves of an older igneous rock. Indeed, where sialic basement complexes have been recognized in the Superior province, they have been found to be tonalitic in composition and exist in relatively small enclaves within younger granitic plutons (Baragar and McGlynn, 1978). On the other hand, if the 2812 Ma age is the true age of intrusion, then the mafic units underlying the acid
pyroclastics of the lower volcanics must be at least this age. This would extend Michipicoten volcanism another 60 Ma. This possibility cannot be discounted considering the fact that these greenstones may range up to 6000 m in stratigraphic thickness (Goodwin 1966).

Alternatively, evidence of xenocrystic zircons from the trondhjemite indicate that this is an inherited age, influenced by a yet older granitic (since it bears zircons) source. However, it is unlikely that most of this trondhjemite has this derivation since it is not possible to obtain a trondhjemite magma from ensialic anatexis unless total melting of the parent is accomplished without the early segregation and escape of low melting eutectic magma (Glickson, 1979). Thus, the trondhjemite is unlikely to have been formed by the remelting of a similar or more differentiated precursor. A more likely source of this rock is the partial melting of basic volcanics near the bottom of the stratigraphic pile - trondhjemite may be derived by 30 - 50% melting of basic rocks (Barker and Arth, 1976). Early basalts could be metamorphosed to amphibolites at shallow depths as a result of the high geothermal gradient proposed for the Archean by Lambert (1976). As the rocks were further depressed, partial melting of the amphibolites would produce a trondhjemite liquid with hornblende as a residual phase (Hunter, 1979). The low initial$^{87}\text{Sr}/^{86}\text{Sr}$, 0.7048, supports this hypothesis. If the parent material was sialic it cannot have been segregated from the mantle for a long period of time (not more than 400 Ma) otherwise a much larger enrichment in the initial $^{87}\text{Sr}/^{86}\text{Sr}$ would result. The lack of amphiboles in the trondhje-
mite is further support for this origin. More conclusive evidence would be obtained by a study of the REE pattern. This origin of the trondhjemite is consistent with the notion of Glikson (1978) that the basement to Archean greenstone belts was earlier greenstone—granite terrain and early Na rich intrusives were derived from the fusion of subsiding crustal root zones of these earlier greenstones. Although earlier greenstones, 2950 to 3000 Ma have been identified in greenstone belts in northwestern Ontario (Nunes and Thurston, 1979; Nunes and Wood, 1979), none have been found in the eastern Superior province.

It is noteworthy that Krogh and Turek (1981) report a concordia diagram for the Mijinemungshing trondhjemite in the Gamitagama area whose discordant points indicate a similar dichotomy of ages. This unit lies between the Michipicoten belt and the Gamitagama belt and is believed to be part of Brulé Bay batholith described by Coleman (1906). The relationship, if any, of this trondhjemite to the Hawk Lake trondhjemite is unknown, except that these rocks are part of the granitic terrain separating the two greenstone belts. The Mijinemungshing trondhjemite would seem to be younger as it intrudes metavolcanics dated at 2713 Ma (Krogh and Turek, 1981); its zircon age is estimated at about 2675 Ma.

This study provided absolute ages for the Michipicoten greenstone belt and a direct comparison of U-Pb and Rb-Sr dating techniques for Archean metavolcanics and granitic rocks. The results indicate that the whole rock Rb-Sr ages are 8-10% younger than the zircon U-Pb ages. Furthermore, the Rb-Sr ages tend to
cluster around an age of 2550 Ma. The discrepancy between the ages can be explained in terms of the behaviour of the two systems under metamorphic conditions and the uncertainty of the $^{87}\text{Rb}$ decay constant. Metamorphism can cause migration and loss of parent and/or daughter components of the Rb-Sr system resulting in younger non primary ages. The zircon U-Pb system, however, although rarely a closed system, is insensitive to such disturbances and in most cases gives reliable crystallization ages. This pattern of disparate ages between the two systems is well recognized in the geochronology of Archean rocks.

The conditions by which rocks became open systems with respect to Rb, $^{87}\text{Sr}$ or Sr is poorly understood. It is notable, however, that calcite – a secondary mineral thought likely to be indicative of disturbed systems – is found widespread in both acid volcanics and intrusive stocks. Yet the Hawk Lake trondhjemite which shows a similar pattern of disturbance contains no calcite. On isochrons having highly scattered points, such as 636, individual samples from the rock have undergone variable decreases in the daughter-parent ratio $^{87}\text{Sr}/^{87}\text{Rb}$. An explanation for young ages from rocks that yield colinear isochrons, such as 659, requires a loss in Sr which is systematically related to the Rb/Sr ratios of the samples, or more likely, a rehomogenization and complete resetting of the radioactive clock after Sr loss.

Isochrons from three different units that are within or slightly outside experimental error are statistically indistinguishable and indicate an age of about 2550 Ma. It would appear
then, that the metamorphic event responsible for producing younger Rb-Sr ages is applicable to the acid volcanics as well as granitic intrusives and had regional distribution. If the Rb-Sr ages from the Gamitagama belt (Turek et al., 1981) are added to the data here, this event now becomes 2560 Ma. This event is considered to represent the Kenoran orogeny as defined by Stockwell (1972). Thus the stratigraphic section, Fig. 15, shows the zircon ages of the volcanic and plutonic rocks as representing the crystallization ages of these rocks. And the Rb-Sr data obtained here, together with data from the Gamitagama area, define an age of 2560 Ma, which is shown in Fig. 15 as the Kenoran orogeny.

The much younger 2285 Ma age shown in Fig. 15 as a post-Kenoran event is difficult to explain — no such event of this age has been formally recognized in the Superior province. However, the Rb-Sr isochron in Fig. 6 certainly reflects a metamorphic event and cannot be dismissed. It is of interest that this event has been recorded in an aplite dike in granitic terrain near White River about 75 km northwest of Wawa (Turek, personal communication). Thus the event may be related to dike intrusion. Other evidence for this event is a biotite K-Ar age of 2169 Ma (Fig. 2) from the Rossport area, 225 km northwest of Wawa.

Wanless et al. (1974), who reported the age, suggested that it may be the result of reheating of the region in post-Kenoran time. On a larger more regional scale this event has been identified in northern Wisconsin and the upper peninsula of Michigan (W.R. Van Schmus, personal communication). In addition, Turek and Peterman (1971) report an age of 2295 Ma (recalculated using $\lambda^{87} \text{Rb} = 1.42 \times 10^{-11} \text{yr}^{-1}$) for the age of cataclasis in greenstone terrain near Rice Lake, Manitoba. Thus a post-Kenoran thermal event has been recognized in the Superior province, however, its exact nature and extent remain to be documented and interpreted.
Attempts to identify and isolate the 2285 Ma event in some of the other data in this study have been unsuccessful. For example, the quartz-feldspar crystal tuff (Fig. 7) gives an age of $2680 \pm 490$ Ma. Points 636 (3), (6) and (7) would give an age of 2466 Ma, with a reasonable degree of fit. However, the remaining three points would give an old age with an initial ratio well below .70, below the meteoritic value. Hence, this analysis is unacceptable. Another treatment tested was to assume that the 636, 660 and 659 samples are coeval and can be pooled together. From the plot of thirteen points (i.e. Fig. 4 and Fig. 7) it is possible to select eight points which would give an age of $2450 \pm 155$ Ma ($\text{initial} \frac{^{87}\text{Sr}}{^{86}\text{Sr}} = 0.7040$) the remaining five points are above the line and are not colinear. Another possible evuluation is to take ten of the thirteen points; this gives $2500 \pm 80$ Ma ($\text{initial} \frac{^{87}\text{Sr}}{^{86}\text{Sr}} = 0.7035$) and the remaining three points are below the line and would give an age of 2485 Ma. These and similar permutations of the data for the acid volcanics were tested and have simply failed to yield any evidence of the 2285 Ma event (with a high or even low initial ratio). Thus the 2285 Ma event being postulated in this thesis must be a sporadic one, possibly related to dike emplacement, faulting or cataclasis.

The lower intercepts of the U-Pb data are difficult to interpret. The results are: $107 \pm 19$ Ma, $364 \pm 126$ Ma, $385 \pm 30$ Ma, $345 \pm 19$ Ma, $464 \pm 14$ Ma and/or $573 \pm 8$ Ma. (These results are given in Table 3). The uncertainties here are perhaps underestimated because the statistical treatment was more concerned with giving a conservative estimate for the upper intercepts. Explanations relating to theses ages are:

1. There are several outlying Paleozoic rocks in north central Ontario;
this implies uplift in the Superior province.

2. In the Michigan basin are evaporites of Silurian age. Uplift could have caused the basin to become a restricted lagoon which led to the formation of these evaporites.

3. In the Devonian and Silurian of the Michigan basin there are several unconformities also indicating uplift.

Possible explanations all have in common uplift, hence, this is considered supportive evidence for the dilatancy model of Goldich and Mudrey (1972).

Goodwin (1977) discussed the similarities of Archean greenstone belts in the Superior province and suggested a common parental origin for the volcanic rocks based on chemical characteristics. Furthermore, Goodwin (1968), observed a pattern in three of the larger belts — Uchi, Wabigoon and Abitibi. This pattern is an increase in length and width, an increase in acid volcanic rocks and an increase in the alkalinity ($K_2O/Na_2O$) of the acid volcanics toward the south. Based on these observations he suggested that the greenstone belts were progressively younger to the south. Similarly, but on a much smaller scale, Goodwin (1977) pointed out that there are four distinct subgroups within the Abitibi belt which are progressively more fractionated in ascending stratigraphic order and overlap eastward with time, indicating westward movement of a lithospheric plate or the eastward migration of a vent.

With the data for the Michipicoten belt obtained in this study, and data from other recent studies of greenstone belts of the Superior province (Davis et al., 1980; Nunes and Jensen, 1980; Nunes and
Thurston, 1980; Nunes and Pyke, 1980; Nunes and Wood, 1980; Krogh and Turek, 1981) it is possible for the first time to compare the chronology of the belts based on U-Pb zircon data. These ages are shown in Fig. 17 and reveal some important patterns. The greenstone belts in Fig. 17 are arranged from West (LHS) to East (RHS). The bar diagrams represent volcanic episodes; for example, the Michipicoten belt is represented by the Upper Volcanics (2696 Ma), Middle Volcanics (no age - but probably equivalent to 2713 Ma of the Gamitagama belt) and Lower Volcanics (2744 Ma and 2749 Ma). It is clear from this that there are correlative volcanic episodes in the different greenstone belts. Moreover, it is also evident that the belts become progressively younger from West to East. Volcanism appears to have commenced about 3000 Ma followed by major extrusive episodes at about 2800 Ma, 2750 Ma, 2720 Ma and 2700 Ma, the latter age marking the cessation of volcanism. Fig. 17 also gives the difference in age between youngest and oldest age in each belt. This difference is a minimum age for the formation of the belt. Thus the belt at North Spirit Lake took at least 280 Ma to develop while the belt at Kirkland Lake took more than 7 Ma to form. Again there is a trend from West to East; the younger the belt the shorter the time required to form it.

This regional age pattern suggests a progressive and diachronous trend of Archean crustal evolution in the Superior province. The apparent eastward younging of the greenstone belts suggests that the original Superior protocontinent of Goodwin (1968), originally situated in the northwest Superior province, grew via plate tectonics,
Fig. 17. Bar diagram for U-Pb zircon ages of Archean volcanic rocks of the Superior Province, including ages from the Michipicoten belt, this study (Hawa). Numbers below refer to stratigraphic thickness and minimum times of formation of the respective greenstone belts. Ages in parentheses are plutonic ages delimiting the minimum ages of volcanism.
to the southeast, by lateral accretion of new crustal material, probably island arcs. The plate tectonic model envisioned may have involved an eastwardly migrating subduction zone or the westwardly movement of a lithospheric plate over a fixed plume. However, to prove such mechanisms were operative more collaborative studies, similar to the petro-chemical studies of the Abitibi belt (Goodwin, 1977), are required for the other greenstone belts.
CONCLUSIONS

1. The precise zircon ages reported in this study confirm the relative age relationships based on stratigraphic mapping.

2. An acid volcanic from the upper volcanics of the Michipicoten belt has a U-Pb zircon age of 2696 ± 2 Ma and is the youngest date reported for volcanics in the Superior province. This age marks the end of Archean volcanism in the greenstone belt. The oldest acid volcanics dated from the belt have an age of 2749 ± 2 Ma. The ages from the upper and lower cycles bracket a minimum time period of 53 Ma for the evolution of the belt.

3. Granitic stocks internally intruding Michipicoten supracrustals are cogenetic and coeval with lower cycle acid volcanics. The age of emplacement of one of these stocks, as determined by zircon U-Pb dating, is found to be 2737 Ma.

4. The Hawk Lake trondhjemite, an intrusive rock external to the greenstone belt was probably emplaced coeval with early volcanism, about 2747 Ma. However, zircons from this rock show evidence of an inherited population of older zircon crystals, which possibly are derived from an older silicious source rock.

5. Rb-Sr ages for the volcanic and intrusive rocks are significantly younger than the zircon ages and I believe the Rb-Sr data defines the Kenoran orogeny at about 2560 Ma in the Wawa belt. A younger event, 2285 Ma, is registered by acid volcanics (which should have a zircon age of 2749 Ma). This event is referred to as a post Kenoran event, the
significance of which is not fully understood, but it may be a period of dike intrusion.

6. The reliability of leaches used to obtain more concordant points on U-Pb discordancy lines has been demonstrated by the fact that two Michipicoten volcanic leaches are colinear with their respective discordancy lines. Leaching of zircons analysed from two granitic units has yielded points that plot above and to the right of the concordia curve and their validity is less certain.

7. The regional pattern of U-Pb zircon ages from greenstone belts of the Superior province show a progressive younging of ages to the southeast with upper volcanic cycles in the west having their time equivalents in the lower volcanic cycles in the east. In addition, the time spans covered in the formation of the greenstone belts become progressively shorter eastward. These apparent trends of the greenstone belts may be related to, and be evidence for, Archean crustal accretion toward the southeast by plate tectonics.

8. The lower intercepts on the U-Pb concordia diagrams are more narrowly defined between the 345 to 385 Ma. These are typical of the Superior province. Possible explanations all have in common uplift related to deposition, erosion, or evaporite formation in the Paleozoic Michigan basin. Hence, this may be supportive evidence for the dilatancy model of Goldich and Mudrey (1972).
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APPENDIX I

Petrography

1. **636 Quartz-Feldspar Crystal Tuff**
   
   Rock is dark grey, dacitic in composition consisting of phenocrysts of rounded quartz (5 mm) and plagioclase (2 mm) with anhedral biotite imparting a porphyrytic appearance to the rock. Opalescent blue quartz crystals have been embayed by resorption. Occasional large (1 mm) laths of calcite. Matrix is composed of quartzofeldspathic material with subordinate sericite and epidote. Collected from roadcut on Highway 101, 500 m east of Wawa, Latitude: 47° 59' 27" N, Longitude: 84° 45' 30" W.

2. **643 Fragmental Volcanic**
   
   Yellow grey, pyroclastic rock of rhyolitic composition consisting dominantly of fine grained matrix (0.03 mm) of quartz and albitic plagioclase and sericite (70% of rock) with larger subhedral quartz grains, calcite and amygdules. Occasional angular lithic fragments (2 mm) of andesitic composition. Iron oxides and limonite are predominant accessory minerals. Collected from "Wawa Tuff" on southwest shore of Wawa Lake just north of Wawa. Latitude: 48° 00' 01" N, Longitude: 84° 48' 17" W.

3. **646 Trondhjemite**
   
   Black and white, medium grained equigranular, slightly foliated rock composed of laths of slightly sericitized plagioclase (2 mm) (An 15) and subordinate orthoclase in a matrix of anhedral quartz (2 mm). Mafic constituents include biotite with pleocroic haloes and overgrowths of epidote on allanite. Accessories: sphene, apatite, zircon, magnetite. Collected from Algoma Central railway tracks at NE corner
of Hawk Lake. Latitude: 48° 04' 18" N, Longitude: 84° 32' 55" W.

4. **659 Dacite**

Dense grey pyroclastic rock consisting of medium size (0.5 mm) sieve-like porphroblasts of chlorite-like mineral (ottrelite?) in a matrix of fine grained recrystallized quartz, sericite and calcite. Accessory iron oxides and epidote disseminated throughout matrix. Collected from Helen Mine property north of Spud Lake. Latitude: 48° 01' 25" N, Longitude: 84° 44' 29" W.

5. **662-6 Diorite**

Medium grained massive rock consisting of highly sauceritized subhedral plagioclase (An$_{25}$) (2 mm) in a matrix of hypidiomorphic granular quartz (2 mm) and patches of ragged chloritized biotite. Secondary minerals include occasional large laths of calcite as well as disseminated epidote, zoisite, sericite and sulfides. Accessory iron oxides, apatite and zircon. Collected about 200 m west of Minto Lake. Latitude: 47° 58' 21" N, Longitude: 84° 44' 48" W.

6. **663-2 Quartz Diorite**

Embayed quartz (3 mm) and sericitized plagioclase (2 mm) and flaky anhedral biotite set in fine grained granular matrix of quartz, plagioclase and epidote. Minor hornblende. Secondary muscovite and calcite in veinlets. Accessory magnetite and zircons. Collected south of Jubilee Lake. Latitude: 47° 58' 21" N, Longitude: 84° 44' 48" W.

7. **664 Granodiorite**

Coarse grained massive rock consisting of blue opalescent quartz grains (5 mm); plagioclase oligoclase in composition crowded
with secondary epidote, carbonate and sericite, and irregular patches of ragged biotite, altered in some places to chlorite. Secondary calcite occurs in veinlets with quartz epidote and sericite. Accessory iron oxides, apatite and zircon. Collected about 700 m north of Michipicoten River. Latitude: 47° 55' 20" N, Longitude: 84° 43' 30" W.
APPENDIX II

Zircon Isolation

The following flow chart illustrates in ten steps the systematic procedure used for zircon isolation. The actual assemblage of minerals indicated in the technique varies, of course, depending upon the type of rock being processed thus making some of the parameters in the latter steps of the separation empirically determined. The following abbreviations are used:

- quartz = qtz
- feldspar = fd
- sulfides = sulf
- mafic minerals = maf
- apatite = apt
- sphene = sph
- fluorite = flu
- magnetite = mag
- muscovite = musc
- barite = bar
- zircon = zrc
- density = p
- magnetic susceptibility = k
- Franz Isodynamic Separator = Franz

Sample 100 kg

1. Crush to 80 mesh (using steel pulverizing plates)

2. Wilfley table separation (inclination = 5°, stroke = 220 RPM)
3. Franz separation (tilt = 10°, inclination = 10°, current = 0.3 amp)

4. HNO₃ bath (7 N, 30 min)

5. C₅H₂Br₄ (p = 2.95)

6. Franz separation (inclination = 5°, current = 0.8 amp)
non magnetic (apt, sph, sulf, flu, bar, zrc)
magnetic (maf)

7. CH₂I₂ (p = 3.33)

heavies (sulf, bar, zrc) lights (apt, flu)

8. Size fractionation (nylon seives)
   Tyler seive sizes: 70; 100; 200; 350

9. Franz fractionation (magnetic fractions based on k)
   constant current = 1.7 amp
   inclination angle 10° (impurities) 7° 6° 5° 4° 3° 2° 1° -1°

10. hand picking (using binocular microscope)

ready for chemistry
APPENDIX III

Rb and Sr Ion Exchange

Figs. A and B show the elution histograms for the ion exchange of Rb and Sr respectively for a column of Dowex 400W-X8 200-400 mesh using 2 M HCl. Rb was isolated by collecting the 30-60 ml fraction. Sr was isolated by collecting the 60-90 ml fraction.
Figure A - Calibration graph for Rb ion exchange.
Figure B - Calibration graph for Sr ion exchange
APPENDIX IV

Sample Suitability Studies

Rb-Sr samples used for dating in this study were analysed further in an attempt to find a correlation between goodness of fit of a point on an isochron and any other measurable parameter. For instance, sample 662(1) on the isochron diagram in Fig. 10 plots above the isochron; the remaining samples are colinear. The non-linearity of this point may be reflected by an anomaly in another parameter. If so, this information would be useful in prescreening nonrepresentative or altered samples before proceeding with age work, or, it would be helpful in understanding why certain coeval and cogenetic points are not colinear. The parameters that have been tested here are Na$_2$O, K$_2$O and the data is summarized in the tables below. The tests proved unsuccessful in determining convincing correlations but the data is reported here in order to provide a background for further research.
Table A - CO₂ Data for Selected Whole Rock Samples Used in this Study

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Fig. C - plot of % Na₂O vs. K₂O for selected samples.

Plate 2. Photomicrograph of zircons from the granodiorite stock (664, 100-200 mesh, M +1°). Note euhedral nature and magmatic zoning of the crystals (cross polarized light).
Plate 3. Photomicrograph of magnetic zircons from the Hawk Lake trondhjemite (646, 200-350 mesh, M + 6°). Note numerous fractures and rounded exteriors (plain polarized light).

Plate 4. Photomicrograph of non-magnetic coarse grained zircons from the Hawk Lake trondhjemite (646, 100-200 mesh, MMN -1°). Note xenocrystic core in upper right crystal (cross polarized light).
Plate 5. Photomicrograph of a thin section from the Hawk Lake trondhjemite showing coarse grained, fractured zircon with dark coloured core. Magnification: x 200 (cross polarized light).

Plate 6. Photomicrograph of leached coarse grained zircons from the Hawk Lake trondhjemite (646, 100-200 mesh, MNM -1°). Note skeletal appearance compared to Plate 4 (cross polarized light).
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