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Precambrian Geology of the Central Wabigoon Subprovince Area, Northwestern Ontario

2010

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ONTARIO GEOLOGICAL SURVEY

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by

D. Stone

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Miscellaneous Release—Data 242

Geochemical Analyses of Rocks, Minerals and Soil in the Central Wabigoon Subprovince Area, Northwestern Ontario

by D. Stone

This digital release is a compilation of geochemical analyses from rock, mineral and soil samples collected from 1995 to 2007 during regional mapping of the central Wabigoon Subprovince area of northwestern Ontario. These data are being released in conjunction with Open File Report (OFR) 5422, *Precambrian Geology of the Central Wabigoon Subprovince Area, Northwestern Ontario* and Preliminary Map P.2229, *Precambrian Geology, Central Wabigoon Subprovince Area, Northwestern Ontario*. The data are provided as 18 Microsoft® Excel® (*.xls*) spreadsheets. Also provided is a compilation of references (Microsoft® Word *.doc* format file) listing geotechnical work done by Atomic Energy of Canada Ltd. at the Atikokan Research Area. Location data are provided in the Universal Transverse Mercator (UTM) projection and grid system, zones 15 and 16, North American Datum 1927 (NAD27). The data are available on 1 CD.

These digital data are available separately from this report.

Abstract

The central Wabigoon Subprovince area (17 500 km^2) is 150 km west northwest of Thunder Bay, Ontario, and includes the towns of Atikokan, Mine Centre, Ignace and Upsala. It is underlain mainly by Archean supracrustal rocks (greenstone and sedimentary belts) and felsic plutonic rocks of the western Superior Province of the Canadian Shield.

For more than a century, the area has been studied by geologists and has featured prominently in growth of knowledge about very old rocks. Here, chronostratigraphic series of Precambrian rocks were defined and debated, greenstone belts were classified according to lithostratigraphic groups and formations and later, key advances were made in defining lithotectonic subprovinces and mechanisms by which the subprovinces developed. The area also has a long history of mining for gold at Mine Centre and Atikokan, iron at the famous Steep Rock locality and platinum group metals at Lac des Iles.

This study investigates the remarkable variation in age (from 3.0 to 2.7 Ga) of rocks in the central Wabigoon Subprovince. Work included regional geologic mapping supported by geochemistry, thermobarometry and geochronology. Soil surveys were done to study the distribution of gold in soil near known bedrock gold occurrences.

On the order of 8 Archean magmatic events are recorded by U/Pb geochronology in the central Wabigoon Subprovince area. In most instances, magmatic events are geographically localized and comprise, for example, certain parts of greenstone belts and adjacent plutonic areas. These geographically constrained magmatic events correspond with tectonic domains (volumes of rock bounded by compositional and structural discontinuities within which there is structural and chronologic homogeneity) and a major focus of geologic research has been to define the probable extent of these domains as well as magmatic and tectonic mechanisms by which the domains formed and evolved.

The Archean magmatic rocks are geochemically comparable to modern rocks formed in plate-tectonic environments such as magmatic arcs and ocean floors. The geochemical data provide evidence that the various tectonic domains can be parts of ancient volcanic arcs, ocean floors and continental arcs such as occur in a modern environment. Packages of domains show local age-zonation marked by an outward younging sequence of domains. This pattern of mainly outward-younging crustal fragments is similar to what is observed in modern continental masses thought to have developed by accretion of crustal fragments through plate tectonic processes. A notable feature is that the possible Mesoarchean crustal fragments are an order of magnitude smaller than late Neoarchean fragments.

Aluminum-in-hornblende barometry combined with geochronology shows that old rocks are more deeply eroded than young rocks. Slow uplift during the Mesoarchean was followed by an episode of rapid uplift during the late Neoarchean. The rapid uplift is possibly related to the growth in size of crustal plates whose accretion led to correspondingly large uplifts in tectonically shortened and overthrust terranes.

Volcanic rocks in the Mine Centre greenstone belt have the same geochemistry as volcanic rocks near known volcanogenic massive sulphide deposits. The host-rock geochemistry and presence of numerous base metal showings suggest that the Mine Centre area has good potential for base metal deposits. Several soil surveys show gold anomalies near known gold occurrences at Atikokan and have identified blind soil anomalies worthy of exploration. The crescentic western end of the Marmion batholith is cut by gabbro dikes and a complex braided fault and has greenschist-facies alteration. This area has many gold showings and is recommended for gold exploration.

Precambrian Geology of the Central Wabigoon Subprovince Area, Northwestern Ontario

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Open File Report 5422
2010

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Introduction

The central Wabigoon Subprovince area has a long history of geologic research, mineral exploration and mining through the past century. In this area, geologists made key advances in understanding complex Archean rocks and the development of Archean cratons (*see* "Previous Mapping and Exploration"). At a regional scale, the western Superior Province was subdivided into belt-like lithotectonic subprovinces including the volcano-plutonic Wabigoon Subprovince (western Wabigoon and Marmion terranes of Figure 1) and the metasedimentary Quetico Subprovince (e.g., Card and Ciesielski 1986). These subprovinces were interpreted to be an island arc and sedimentary prism accreted together about 2.7 billion years ago (Percival and Williams 1989) during a major stage in growth of the Superior Province known as the Kenoran Orogeny. However, detailed geochronology (e.g., Davis and Jackson 1988) showed that there are variations in age of up to 300 million years for rocks of the central Wabigoon Subprovince area. This alluded to a magmatic and tectonic history for the area more complex than could be explained by a single orogeny. The present study was initiated in the 1990s to characterize the regional geology and to provide a detailed analysis of the magmatic and tectonic evolution of the central Wabigoon Subprovince area. The work included geologic mapping with an emphasis on geochronology and provides insight on Archean crustal development with implications for mineral exploration.

LOCATION AND GEOLOGIC SETTING

The central Wabigoon Subprovince study area (17 500 km^2) extends 60 to 280 km west-northwest of Thunder Bay, Ontario. The area includes the towns of Atikokan, Mine Centre, Upsala and Ignace. It is crossed by Highways 11 and 17, numerous secondary roads and both the Canadian Pacific and Canadian National railways. Access to various parts of the study area was achieved by driving on logging roads, using boats and canoes on lakes and by traversing through the bush.

The central Wabigoon Subprovince area is situated mainly within the western Wabigoon and Marmion terranes of the western Superior Province of the Canadian Shield (*see* Figure 1). The area is underlain by Archean greenstone belts (mainly metavolcanic rocks) interspersed with plutonic domains (mainly granitic rocks). The area also includes the northern margin of the Archean metasedimentary Quetico Subprovince. Proterozoic gabbro of the Nipigon Sill Complex intrudes Archean rocks in the east.

FIELD WORK, MAPPING AND SAMPLING

The central Wabigoon Subprovince area was mapped at a scale of 1:50 000 in the summers of 1995 to 2004 to produce a series of 18 regional maps (Figure 2, Chart A, back pocket). From 2005 to 2007, several of the early maps were revised due to the construction of new logging roads that provided better access and improved exposure of the bedrock. Two areas were mapped in the Steep Rock Lake area at a scale of 1:10 000 (Stone and Lennox 2007; Stone 2008a). All maps were compiled into a 1:250 000 scale geologic map that accompanies this report (P.2229, back pocket; Stone 2010a).

Air photographs taken at a scale of 1:50 000 from 1976 to 1996 and supplied by the National Airphoto Library were used for navigation and as a plotting base in the field. Air photographs of 1:20 000 scale and 1995 to 1996 vintage were obtained from the Ontario Ministry of Natural Resources as a detailed plotting base for the mapping of many greenstone belts. Enlarged air photographs (1:5000 scale) were used for mapping the Steep Rock Lake area. Geologic information was transferred from air photographs and field notes to digital base maps and several databases using AutoCAD® version 14 and

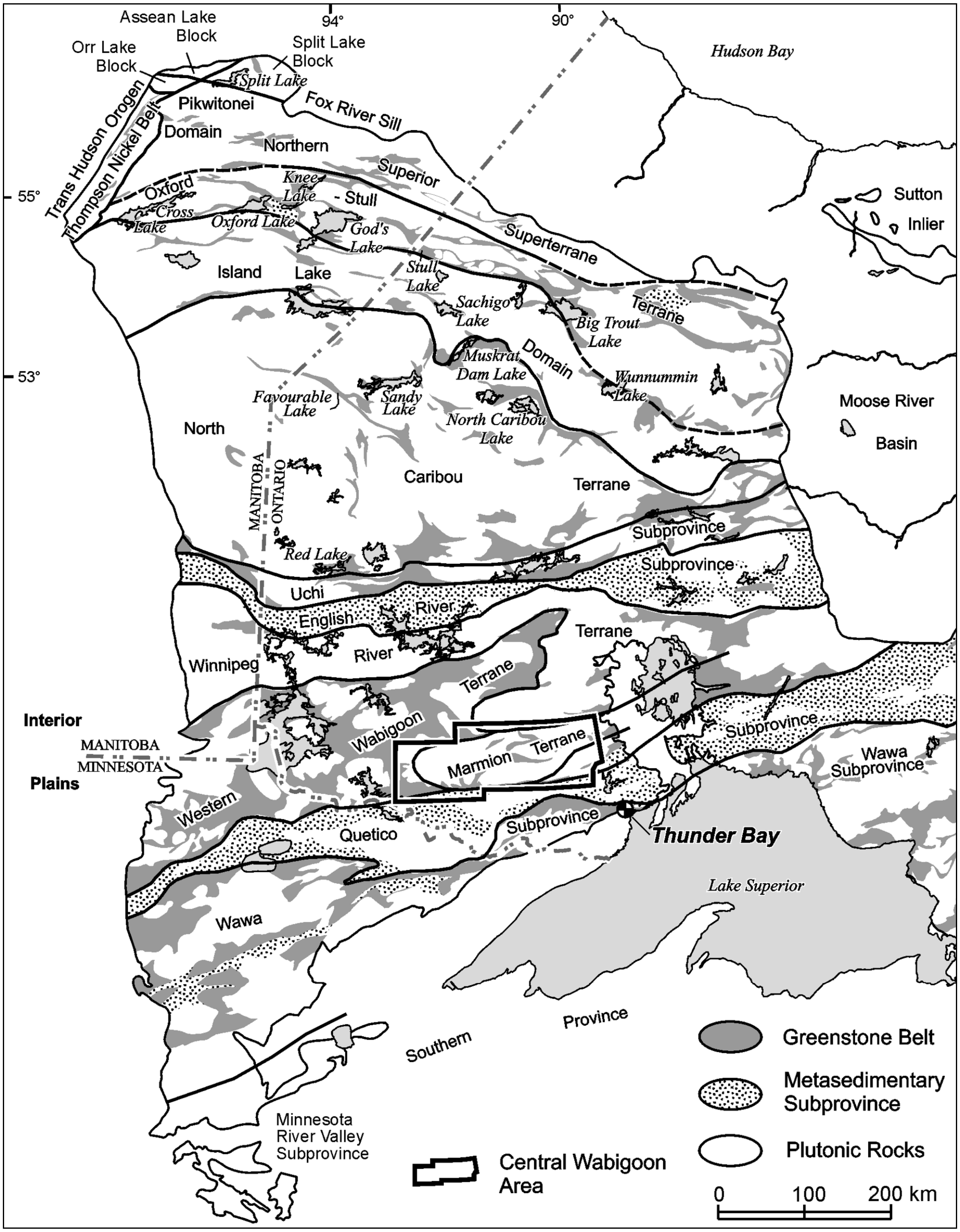

Figure 1. Tectonic subdivisions of the western Superior Province.

Fieldlog version 3.0. The digital base maps were derived from the Natural Resources and Values Information System (NRVIS), supplied by the Ontario Ministry of Natural Resources, Geospatial Data Exchange with cultural information added by the authors. The base maps are projected by the Lambert Conformable method with easting and northing co-ordinates specified in metres by the Universal Transverse Mercator (UTM) grid for Zone 15 using North American Datum 1927 (NAD27). The Armistice Lake and Lac des Iles maps, which are east of longitude 90°, have co-ordinates in UTM Zone 16, NAD27.

Samples of rock and soil were collected throughout the area and a list of samples, together with analytical results, are provided in a separate publication (Miscellaneous Release—Data 242: Stone 2010b). Rock samples were assayed for metals of economic interest or for a range of major and trace elements. The compositions of minerals were obtained by microprobe for many rock samples and the data used to establish mineral assemblages and for thermobarometry. Samples of soil (mainly till and modern alluvium) were analyzed for a range of metals including gold and some samples were processed for indicator grains of gold, metamorphosed or magmatic sulphide minerals and kimberlite indicator minerals (*see* "Rock Samples and Geochemical Analyses" for description of analytical methods).

DRAINAGE AND TOPOGRAPHY

The central Wabigoon Subprovince area represents a topographic highland spanning the divide between easterly and northerly flowing river systems. The extreme eastern parts of the area are drained by the Gull and Dog rivers to Lake Nipigon and Lake Superior and, hence, eastward to the Atlantic Ocean. Central and western parts of the area are drained westward through the Seine and English river systems to Lake Winnipeg and, hence, north to Hudson Bay.

Topography is subdued and slopes more-or-less eastward in the extreme eastern part of the area (Armistice Lake and Lac des Iles areas, *see* Figure 2) and westward through the remainder of the area. Muskeg Lake, which is located near the drainage-divide, has an elevation of 473 m above sea level. Going westward from Muskeg Lake, major lakes along the Seine River decrease in elevation and include Lac des Mille Lacs (457 m asl), Steep Rock Lake (395 m asl) and Rainy Lake (291 m asl). Locally, hills can exceed 50 m elevation above the regional peneplain. This scale of local relief is developed in the Steep Rock greenstone belt at Atikokan where resistant gabbro bluffs are juxtaposed with valleys underlain by more easily eroded supracrustal rocks. In contrast, large areas underlain by horizontally fractured granite east of Upsala have extremely subdued topography marked by muskeg swamps.

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Unless otherwise noted, all data for geochemical analyses in this report were provided by the Geoscience Laboratories, Ontario Geological Survey in Sudbury.

Previous Mapping and Exploration

EARLY STUDIES

The central Wabigoon Subprovince area is an important locality where, more than a century ago, geologists took up the challenging task of subdividing Precambrian rocks. Sir William Logan had initiated the subdivision of Precambrian rocks into chronostratigraphic series. Working in the Lake Huron area, Logan et al. (1863) recognized supracrustal sequences of the Huronian Series unconformably overlying older granitic and gneissic rocks of the Laurentian Series. Bell (1873) was one of the first geologists to study the present area and, using Logan's two-fold subdivision, he described rocks of the Huronian (which were probably Archean greenstone sequences) and Laurentian (Archean felsic plutonic rocks) series along a canoe route leading west from Lake Superior to Lac des Mille Lacs and west on the Seine River. Bell's route extended north from the Seine River across what is now known as the Lumby Lake greenstone belt to the headwaters of the English River and then westward. Later, following completion of the Canadian Pacific Railway, Bell (1883) described rocks exposed along the railway between the stations of English River and Ignace.

Following earlier work in the Lake of the Woods area (Lawson 1885) and field work at Rainy Lake in 1885 to 1887, A.C. Lawson (Lawson 1888) described the general geology of the Rainy Lake region where he applied the term "Keewatin Series" for volcanic rocks of the area and introduced the Coutchiching Series for metasedimentary sequences of what is presently known as the Quetico Subprovince (*see* Figure 1). Lawson considered the Coutchiching Series to be older than the Keewatin Series. In 1891, H.L. Smyth (Smyth 1891) published a map of the Steep Rock Lake area and identified a rare example of a Precambrian unconformity where the Steep Rock Series overlies the Laurentian Series. The reports of Lawson and Smyth prompted other geologists to visit the Rainy Lake–Atikokan area with the result that considerable debate arose over the relative ages of the various Precambrian series (*see* "Previous Work" *in* Stone, Kamineni and Jackson 1992). Nonetheless, by the end of the nineteenth century, Coleman (1898) was able to compile a regional map showing the distribution of Keewatin greenstone belts through much of the present study area.

Prospectors were active in the 1890s and discovered several gold showings in the Lower Seine River and Upper Seine River localities. These early gold mining camps near Mine Centre and Atikokan were described by Coleman (1895) and Blue (1900). Several properties including the Golden Star, Foley and Olive mines in the Mine Centre area were developed and produced a few thousand ounces of gold (*see* Poulsen (2000a, 2000b) for detailed descriptions of past-producing mines in the Mine Centre area). Similarly, several properties in the Atikokan area including the Harold, Sawbill and Hammond (*see* Figure 2) were developed and milled a few thousand tons of ore (Wilkinson 1982). Although these gold mines ceased operation by 1900, several new gold and base metal occurrences were identified and underwent limited development in the early years of the twentieth century. These include the Elizabeth and Sunbeam mines, west and east of Steep Rock Lake, respectively, and the Port Arthur Copper Mine at Mine Centre (Wilkinson 1982; Poulsen 2000b). In reviewing the status of gold deposits in the Kenora–Rainy River area, Bruce (1925) reported renewed exploration at the Foley Mine near Mine Centre in 1924 although no new production resulted from this work.

Prospecting in the late nineteenth century led not only to the discovery of gold at Mine Centre and Atikokan, but also iron within a discontinuous belt of magnetite + iron sulphide deposits extending between Atikokan and Sapawe. One of the larger iron orebodies at Sapawe was developed as the Atikokan Iron Mine and, from 1907 to 1911, 82 300 tonnes of ore averaging 60% iron were shipped by rail from this locality to a smelter in Port Arthur (Hawley 1930a). As was the case with the early gold mining operations, the production of iron was uneconomical and the Atikokan Iron Mine ceased operation within a few years of commencement.

Table 1. Chronostratigraphic subdivisions of Precambrian rocks in the Atikokan–Lake Superior area (*after* Lawson 1912).

Series	Rock Type
Keweenawan	Volcanic rocks and intrusions
Animikie	Sedimentary rocks
Algoman	Granitic intrusions
Seine	Conglomerate
Steep Rock	Sedimentary and metavolcanic rocks
Laurentian	Felsic plutonic and gneissic rocks
Keewatin	Volcanic rocks
Coutchiching	Sedimentary rocks

Iron would eventually play an important role in development of the Atikokan area. In 1890, W.H.C. Smith of the Geological Survey of Canada noted boulders of hematite ore on the south shore of Steep Rock Lake. Upon publishing the results of Smith's observations some years later, McInnes (1899) concluded that the source of the rich iron ore boulders lay beneath the waters of Steep Rock Lake. McInnes's report prompted staking and exploration; however, development of the orebody would wait until the Second World War when a perceived shortage of iron gave the impetus to drain the lake and mine the iron.

Also in the early decades of the twentieth century, both Canadian and American geologists studied and debated the correlation of Precambrian series in the central Wabigoon area. A.C. Lawson, then a professor at the University of California, continued to visit the Rainy Lake–Atikokan area and contributed memoirs on the geology of Steep Rock Lake (Lawson 1912) and the Rainy Lake area (Lawson 1913). In these reports, Lawson interpreted the Steep Rock Series as a synclinal sequence overlying the Laurentian and Keewatin series. He introduced the Seine Series to represent coarse clastic sediments somewhat younger than those of the Steep Rock Series and overlying the Laurentian, Keewatin and Coutchiching series. Lawson (1913) also identified a late "Algoman" Series of massive granites and reaffirmed the interpretation that he had made 24 years earlier wherein the volcanic Keewatin Series overlies the metasedimentary Coutchiching Series (Table 1).

In the era between the 2 world wars, the geologic literature recorded much debate over the relative ages of the Keewatin and Coutchiching series. Grout (1925) used graded sedimentary beds to determine that sedimentary rocks of the Coutchiching Series overlie volcanic rocks of the Keewatin Series at several localities in Minnesota and Ontario. Tanton (1926) disagreed with Grout on the basis of sedimentary beds in the Coutchiching that young north toward the Keewatin near Atikokan. Hawley (1930b) and Gill (1931) debated relative ages of the Keewatin and Coutchiching at the faulted boundary between these series in the Atikokan area. Resolution of the problem would not come until much later when Poulsen, Borradile and Kehlenbeck (1980) showed that the stratigraphy was overturned in Lawson's type locality at Bear Pass on Rainy Lake and that even though the Keewatin is structurally on top of the Coutchiching, the Keewatin is stratigraphically lower and chronologically older.

During the 1930s, T.L. Tanton of the Geological Survey of Canada systematically mapped the central Wabigoon Subprovince area adding detail to the known distribution of supracrustal and plutonic series. Tanton (1936, 1938a, 1938b, 1940) produced a set of geologic maps at a scale of 1:253 440 and a summary compilation map (Tanton 1939) at a scale of 1:506 880. Beginning also in the 1930s, the Ontario Department of Mines initiated detailed mapping and remapping of greenstone belts in and around the central Wabigoon Subprovince area—a task that would continue until the present. Thomson (1934a, 1934b) and Satterly (1943) mapped adjoining parts of the Manitou–Wabigoon lakes greenstone belt, which lies northwest of the present area and described several molybdenum occurrences discovered by prospectors south of Ignace.

THE MID-TWENTIETH CENTURY AND STEEP ROCK MINING

In 1939, E.S. Moore published a report and map on the geology of the Atikokan area and introduced the Timiskaming Series for late clastic sedimentary rocks. Earlier, Miller (1913) and Cooke (1922) had described the classic section of conglomerates, sandstones, greywackes and intercalated trachytes at Lake Timiskaming in northeastern Ontario. By the 1930s, many sequences of coarse clastic sedimentary rocks unconformably overlying the Keewatin Series and occurring at scattered localities throughout Ontario were included with the Timiskaming Series (e.g., Pettijohn 1935). Moore (1940) interpreted the Seine Series and Steep Rock Series to be local components of the Timiskaming Series.

Moore (1940) reviewed the episodic development work on old gold properties in the Atikokan area and together with Bartley (1940) and Brant (1940) focussed on the geology and recent exploration for iron at Steep Rock Lake. Much earlier, McInnes (1899), Miller (1903) and Tanton (1927) had predicted iron ore beneath Steep Rock Lake on the basis of glacially transported boulders of hematite on the south shore of the lake. During the winter of 1937–1938, Jules Cross and a consortium later known as the Steerola Exploration Company followed up on dip needle surveys by diamond drilling from the ice and succeeded in outlining several bodies of hematite beneath the middle arm of Steep Rock Lake. A shaft was sunk in an attempt to reach the ore, but had to be abandoned due to flooding.

Overlain by 23 to 46 m of clay and 30 to 60 m of water, the iron orebodies might have remained undeveloped beneath Steep Rock Lake except for the onset of World War II.

Fearing shortages of iron ore, the United States government agreed to provide 10 million dollars toward development of the Steep Rock iron range (Coffin 1957) and this together with privately raised money allowed the newly formed Steep Rock Iron Mines Ltd. to begin the immense mining project. The Seine River was diverted by means of a series of dams and rock cuts, to flow around Steep Rock Lake. Once isolated from the regional watershed, the big lake was dewatered using 14 barge-mounted electric pumps. Immense deposits of clay and gravel were dislodged using hydraulic monitors and pumped as slurries from the basin. By September 1944, the first iron ore was dug from the former lake bed.

Three open pits (Errington, Roberts and Hogarth) were eventually developed along the former middle arm of Steep Rock Lake with limited underground production coming from 2 shafts. In 1949, the deposit at Falls Bay was optioned to the Inland Steel Company and went into production as the Caland Mine in 1960. The Steep Rock and Caland mines operated until 1979 with a total production of approximately 100 million tonnes of iron ore. From 1958 to 1964, Canadian Charleston Limited concentrated 0.71 million tonnes of iron ore from gravel deposits in the vicinity of the Atikokan airport (Shklanka 1972). This represents the only placer mining operation in Ontario.

The Steep Rock Lake area captivated the attention of engineers and geologists for decades and was a prime locality for developments in dredging and open-pit mining methods. The dredging operation provided excellent exposures of glacial deposits and prompted pioneering research on the origin and physical properties of varved clays (e.g., Antevs 1951; Legget and Bartley 1953; Eden 1955; Hardy and Legget 1960). The dispersion of bedrock material by the movement of glaciers (Dreimanis 1956) was also studied at Steep Rock Lake.

Precambrian geologists were attracted to the new bedrock exposed by the mining operation at Steep Rock Lake resulting in numerous interpretations of the geologic units in the area and the origin of the iron ore. Building upon the work of Smyth (1891) and Lawson (1912), Moore (1940) and Bartley (1940) confirmed that the Steep Rock Series unconformably overlies the Laurentian basement complex. The Steep Rock series consisted of 1) a basal conglomerate and coarse arkosic grit; 2) limestone–dolostone;

3) ferruginous carbonate and 4) upper igneous formations of lavas, dikes and pyroclastic rocks. These authors believed that the iron ore developed by oxidation and leaching of the ferruginous carbonate supplemented by the oxidation to hematite of pyrite from the ashrock horizon. Other workers (e.g., Tanton 1941; Roberts and Bartley 1943; and Tanton 1946) favoured hydrothermal replacement whereby the iron was introduced from an igneous source at depth. Smith (1942) thought that the goethite-hematite deposits resulted from secondary alteration of an iron formation.

A.W. Jolliffe, a professor at Queen's University, came to study the Steep Rock Lake area in the years after mining began. Aided by new exposures and assisted by the research theses of his graduate students, Jolliffe applied lithostratigraphic terminology to the sequence and defined 4 formations (Conglomerate, Dolomite, Orezone and Ashrock) within the Steep Rock Group. Jolliffe (1955, 1966) recognized an unconformity on top of the Dolomite Formation and proposed that the ferruginous material of the Orezone Formation was precipitated chemically or biochemically as limonite from seawater within a shallow restricted basin environment and subsequently underwent hydrothermal alteration to goethite and hematite. Later workers dissented, however, and generally concurred with Smith (1942). McIntosh (1972) noted scattered small units of iron formation within and adjacent to the Orezone Formation in the Caland orebody and Shklanka (1972) cited this and a variety of factors including mineral assemblages of low metamorphic grade as evidence that the goethite-hematite deposits originated by alteration of a pre-existing iron formation by downward percolating meteoric waters. Kimberley and Sorbara (1976) and Wilks (1986) supported the model of Shklanka and suggested that alteration could have occurred as late as the Cretaceous.

Geologists continued to study the Steep Rock Lake area after the cessation of mining. Wilks and Nisbett (1988) renamed and described the formations of the Steep Rock Group (Wagita, Mosher, Jolliffe and Dismal formations) and described stromatolites within the Mosher Formation. These authors suggested that the Steep Rock Group was deposited in a rifted continental margin environment and noted a structural break between the upper Dismal Formation and adjacent pillow lavas (Witch Bay Formation) of the western Steep Rock Lake greenstone belt. Schaefer and Morton (1991) provided detailed petrographic descriptions of the Dismal Formation, but perhaps the most significant recent developments in Steep Rock area geology came from Kusky and Hudleston (1999) who proposed that the Dismal and Witch Bay formations were allochthonous and had been thrust into place onto the lower formations of the Steep Rock Group. This interpretation required a previously unrecognized thrust fault between the Jolliffe and Dismal formations.

LATE REGIONAL WORK

Although the Steep Rock Lake area had been the focal point of geologic research through the middle decades of the twentieth century, other parts of the central Wabigoon Subprovince area began to receive attention in the 1960s. The Ontario Department of Mines, later to become the Ontario Geological Survey, began systematic detailed mapping of greenstone belts in the area. Young (1960), Woolverton (1960), Irvine (1963) and Kaye (1967) mapped greenstone belts at Calm Lake, Lumby Lake and western and eastern Lac des Mille Lacs. Milne (1964) mapped the Garden Lake greenstone belt and Kaye (1969) did a geologic survey of the southwestern Lac des Iles greenstone belt (*see* Figure 2 for locations of these greenstone belts). Somewhat later, parts of the Mine Centre and Western Wabigoon greenstone belts at Otukamamoan Lake, Fort Frances and Rainy Lake were mapped by Blackburn (1973), Davies (1973) and Harris (1974). The Steep Rock Lake and Finlayson Lake greenstone belts were studied by Shklanka (1972) and Fenwick (1976) and parts of the Lac des Mille Lacs greenstone belt at Crooked Pine Lake and Sapawe were mapped by Pirie (1978) and McIlwaine and Larsen (1981a, 1981b).

The new generation of geologic mapping led to gradual abandonment of the controversial chrono-stratigraphic terminology that had been adopted by earlier geologists. Instead, most mappers focussed on the supracrustal sequences and adopted a simple lithologic terminology. Volcanic sequences were subdivided into mafic, intermediate and felsic varieties that were broadly correlative with basalt, andesite and dacite-rhyolite compositions. Efforts were made to distinguish mafic intrusive rocks (gabbro) from mafic extrusive rocks within the greenstone belts. Similarly, sedimentary sequences were subdivided into rare chemical sedimentary units including iron formation and limestone and more widespread units of greywacke and conglomerate. Detailed mapping typically stopped at greenstone belt margins with the result that maps produced during the 1960s and 1970s showed greenstone sequences set in unsubdivided felsic plutonic rocks.

In the early 1960s, airborne magnetic surveys were done over most of Ontario. The results published as a series of 1:250 000 scale aeromagnetic maps (e.g., Geological Survey of Canada–Ontario Department of Mines 1965a-e and Geological Survey of Canada 1989) provided a new source of information and greatly aided in the interpretation of regional geology. Features such as the boundaries of greenstone belts and major faults could be identified in the aeromagnetic maps leading to refinement of regional geologic maps, particularly in areas of poor exposure. The aeromagnetic maps would continue to be used for decades, but a significant early impact was in the identification of intermediate to mafic intrusions within previously unmapped areas of felsic plutonic rocks. The intrusions such as the Entwine stock of the sanukitoid suite and the mafic intrusions of the Lac des Iles suite were identified as oval magnetic "highs" in the aeromagnetic maps and many were staked after the maps were released. Davies (1965), Kaye (1966) and Pye (1968) subsequently mapped the Entwine, Tib and Lac des Iles stocks, respectively. Although early exploration led to the discovery of base metal occurrences in these and other intermediate to mafic intrusions, later work would reveal platinum group metal mineralization as well.

The focus of mineral exploration shifted from iron to precious, base and platinum group metals in the latter part of the twentieth century. The historic gold areas at Mine Centre and Atikokan received sporadic but more-or-less continuous attention by prospectors during this era and, although some new gold occurrences were discovered, the Sapawe Mine 15 km east of Atikokan represents the only significant gold production. The Sapawe Mine milled 38 000 tonnes of ore averaging 5 g/t Au from 1964 to 1966 (Wilkinson 1982). Discovery of the Kidd Creek Mine near Timmins in 1964 was an impetus for base metal exploration through Ontario. Major companies including Amax Exploration Inc., Canadian Nickel Co. Limited, Dome Exploration (Canada) Limited, Newmont Mining Corp. of Canada Limited and Phelps Dodge Corp. (Canada) Limited were active in the present area. The style of base metal exploration typically involved drilling of magnetic and electrical anomalies identified in airborne geophysical surveys. Sage, Breaks, Stott, McWilliams and Robertson (1974) reviewed the results of base metal exploration projects of the 1960s and early 1970s in eastern parts of the present area.

Since the 1940s, the Ontario government required prospectors and exploration companies who hold mining claims to do exploration work and file reports in order to maintain their claims. With the growth in exploration in the 1970s, these reports, known as assessment files, grew in volume and became a valuable record of mineral exploration. The assessment files for the central Wabigoon Subprovince area are held in offices of the Resident Geologists in Kenora and Thunder Bay and are available online through GeologyOntario (www.ontario.ca/geology/).

The intensity of mineral exploration followed a cyclical pattern as the surge of base metal exploration in the 1970s was followed by a lull in the early 1980s and renewed activity in the mid 1980s. Exploration in the 1980s was focussed partly on gold, with the result that substantial resources were defined at several historic gold properties and many new gold occurrences were identified in the Atikokan area. Wilkinson (1982) and Schnieders and Dutka (1985) provided comprehensive descriptions of gold occurrences of the Atikokan area. By 1990, exploration had identified a resource 250 000 tonnes of ore grading 4.8 g/t Au at the Elizabeth property (Lavigne, Scott and Sarvas 1990) and, by 1997, a resource of 88 million tonnes of ore averaging 0.93 g/t Au was estimated for the Hammond Reef property (Brett Resources Inc., www.brettresources.com, Hammond Reef, "Exploration History" [accessed April 14, 2010]).

Also by the mid 1980s, exploration had defined significant palladium and platinum mineralization within the gabbroic Mine Block intrusion of the Lac des Iles suite at Lac des Iles. In 1993, Madeleine Mines Limited, later renamed North American Palladium Ltd., began mining the Roby zone of the Mine Block intrusion and a major exploration program for platinum group metals was initiated in mafic intrusions of the Lac des Iles suite through eastern parts of the present area (Lavigne and Michaud 2001). By the end of 2004, the Lac des Iles Mine had milled 5 298 544 tonnes of ore and had produced 8758 kg palladium with subsidiary platinum, gold, copper and nickel (Scott, Magee and Komar 2006).

Beginning in the late 1960s, the Ontario Department of Mines did regional helicopter-supported reconnaissance mapping in remote areas of the province. In many instances, these surveys focussed on felsic plutonic areas that had received little or no previous mapping. In 1974, one such reconnaissance survey (Operation Ignace–Armstrong) was done over central and northern parts of the study area and published as a series of maps at a scale of 1:126 720. For the first time, the maps (Sage, Breaks, Stott, McWilliams and Ali 1974; Sage, Breaks, Stott, McWilliams and Atkinson 1974; Sage, Breaks, Stott, McWilliams and Robertson 1974) and report (Sage 1998a, 1998b) of Operation Ignace–Armstrong showed and described subdivisions of felsic plutonic rocks broadly through the central Wabigoon Subprovince area. For example, several large massive intrusions such as the White Otter granite batholith were distinguished from gneissic rocks and foliated tonalite.

Selective parts of greenstone belts within the central Wabigoon Subprovince area continued to be mapped by the Ontario Geological Survey in the latter part of the twentieth century. New tools including routine geochemical analyses of rocks were used in the identification of rock units. Geophysical surveys, showing greater detail than airborne surveys of the 1960s, were flown over many greenstone sequences of the central Wabigoon Subprovince area (Ontario Geological Survey 2001, 2002, 2003) and helped to identify geologic features such as lithologic units, geologic contacts and faults. Geochronologic studies (e.g., Davis and Jackson 1988; Davis, Poulsen and Kamo 1989) permitted the distinction of rock units by age and provided insight on the timing of magmatic, depositional and tectonic events in the area.

The new geologic work emphasized the identification of economic mineralization and mineralized environments in aid of exploration programs by prospectors and mining companies. Wood et al. (1980a, 1980b) restudied the mineralized greenstone sequences of the Mine Centre area. Although his work was not published until later (Poulsen 2000a, 2000b), K.H. Poulsen mapped the Mine Centre–Bear Narrows area in the 1980s with emphasis on description of base- and precious-metal occurrences. Blackburn (1982), Berger (1991) and Smith (1993) mapped parts of the Manitou Lake greenstone belt at the western margin of the present area. Blocks of the Mine Centre greenstone belt between Mine Centre and Atikokan were studied by Fumerton (1986) and Shanks (1994). Jackson (1985a, 1985b) provided a new set of geologic maps of the Lumby Lake greenstone belt.

Advanced exploration for platinum group metals in the Roby zone of the Mine Block intrusion prompted renewed mapping of the Lac des Iles area in the 1980s. Sutcliffe and Smith (1988) produced a regional map of the Lac des Iles area as detailed studies were made of the Lac des Iles–Mine Block and Tib intrusions by Sutcliffe and Sweeny (1986) and Smith and Sutcliffe (1987). Also in the late 1980s, staff of the Ontario Geological Survey began a comprehensive review of the geology of Ontario. Published in 1991 and 1992, the *Geology of Ontario* volumes including a series of 1:1 000 000 scale maps, provided a comprehensive compilation and description of the geology of the province. Blackburn et al. (1991) described the geology, metallogeny and structural evolution of the Wabigoon Subprovince that includes the central Wabigoon Subprovince area. Later, in 1995, the present project was initiated to study the distribution of Mesoarchean and Neoarchean rocks in the central Wabigoon Subprovince area. Other mapping conducted in the years following *Geology of Ontario* includes the work of Hart et al. (2000) and Hart, MacDonald and Lepine (2001a, 2001b) in the Garden Lake, Lac des Iles and Heaven Lake greenstone belts at the east side of the study area.

In addition to routine government mapping and mineral exploration, the central Wabigoon Subprovince area has been the site of 3 special geologic research projects. The earliest of these was the University of Toronto's Geotraverse Project that, in the mid 1970s, set out to study a large area of the western Superior Province extending north over a distance of 500 km from the central Wabigoon Subprovince area. Shepherded by Professors A.M. Goodwin and G.F. West, the Geotraverse Project focussed the attention of numerous researchers and graduate students on the lithologic, metallogenic and geophysical characteristics of the Archean rocks of the western Superior Province (e.g., Goodwin 1976a; Goodwin and Sabag 1976; West 1976). Major goals of the Geotraverse Project were to constrain tectonic models for the development and regional disposition of the belt-like subprovinces (Goodwin 1976b) and to understand the structural evolution of the area (e.g., Schwerdtner et al. 1979). Although there was generally insufficient geologic information available in the 1970s to completely achieve these goals, the Geotraverse Project nonetheless drew attention to the regional lithotectonic architecture of the western Superior Province and was a foundation for future research.

A second specialized research project developed from a 1978 decision by the governments of Canada and Ontario to establish the Nuclear Fuel Waste Management Program to assure the safe and permanent disposal of nuclear fuel waste. Atomic Energy of Canada Limited was made responsible for research and development on a disposal concept that involved placing nuclear waste in a deep underground repository in intrusive igneous rock. To assess the disposal concept, Atomic Energy of Canada Limited needed to evaluate a broad range of geotechnical problems including the stability of an underground vault, the effects of radiogenic heat on the disposal vault and the extent to which groundwater might interact with the waste and transport the waste back to the biosphere. In 1979, research was initiated at several sites in the Canadian Shield, one of which was located within the Eye–Dashwa granite pluton, 25 km north of Atikokan. The geotechnical research at Atikokan included regional geologic mapping (Stone, Kamineni and Jackson 1992), diamond drilling and detailed fracture studies (Kamineni, Thivierge and Stone 1988) and a range of geophysical and hydrogeological work designed to identify faults and fractures and to evaluate the movement of groundwater through these discontinuities. References to reports done at the Atikokan research area are provided in Stone (2010b: Miscellaneous Release—Data 242). The geotechnical research continued for a decade at Atikokan and the results were used with those from other research areas to constrain the Geosphere Model, which is a numerical simulation of three-dimensional groundwater flow, heat transport and solute transport in a fractured/porous media (Davidson et al. 1994). The Geosphere Model, together with Vault and Biosphere models were designed to predict the expected long-term performance of a disposal system situated in plutonic rocks of the Canadian Shield.

The Lithoprobe and NATMAP transects represent the latest multidisciplinary research project overlapping the central Wabigoon Subprovince area. Funded largely by the Government of Canada and active from 1996 to 2003, the objectives of the Lithoprobe and NATMAP transects were to define the extent of continental and oceanic domains in the western Superior Province and to test the accretionary tectonic model for growth of the Superior craton. The Lithoprobe and NATMAP projects provided three-dimensional geophysical imaging of the earth's crust and new data through field mapping, geochemical and geochronological studies of selected area. Products included sets of 1:250 000 scale geology and tectonic assemblage maps (e.g., Stone, Tomlinson et al. 2002) and a special volume of the *Canadian Journal of Earth Sciences* documenting the contributions of many workers (e.g., Percival and Helmstaedt 2006). In summarizing results of the Lithoprobe and NATMAP research, Percival et al. (2006) endorsed the accretionary model whereby the Superior craton was assembled from fragments of continental and oceanic crust during 5 discrete accretionary events.

Subdivisions of Archean Rocks and Tectonic Evolution of the Western Superior Province: A Review

CHRONOSTRATIGRAPHIC AND LITHOSTRATIGRAPHIC SUBDIVISIONS

R. Bell, A.C. Lawson and H.L. Smyth (Bell 1873; Lawson 1888; Smyth 1891) initiated the classification of Precambrian rocks of the central Wabigoon Subprovince area according to chronostratigraphic series and this terminology (*see* Table 1) was widely used for the first half of the twentieth century. A chronostratigraphic series is essentially a supracrustal or plutonic rock unit whose relative age is established (or thought to be established) on the basis of stratigraphic superposition or crosscutting relations, with respect to rock units in adjacent chronostratigraphic series. In many ways, the chronostratigraphic terminology was an outgrowth or extension of lithostratigraphic terminology that had been successfully used to subdivide Paleozoic rocks in Europe and North America in the nineteenth century. Unfortunately, Precambrian rocks were generated within numerous ancient magmatic and depositional events and are much more complexly deformed than the layered strata of Paleozoic sequences.

Early geologists did not fully appreciate the complexity of Precambrian rocks and that the stratigraphic or crosscutting relations between rock units at one locality were not necessarily consistent from place to place and could not be extrapolated widely. For example, although felsic plutonic rocks were subdivided into an older gneissic series (Laurentian Series) and a younger massive series (Algoman Series), a compilation of radiometric age determinations (e.g., Stone and Davis 2006) shows that there are as many as 8 generations of felsic plutonic rocks varying in age by up to 300 million years in the central Wabigoon Subprovince area. Similarly, there are numerous generations of mafic volcanic rocks within what was broadly considered to be the Keewatin Series. Hence, conflicting age relations were often observed between the Laurentian, Keewatin and other series. Literature of the early half of the twentieth century, some of which is reviewed in the previous section, recorded much debate over the assignment of various units to certain series and the relative ages of the series.

Prior to the Second World War, Wilson (1939a) lamented the shortcomings of classifying Precambrian rocks according to chronostratigraphic series and acknowledged that many previous chronostratigraphic subdivisions of Archean rocks were, in fact, lithostratigraphic or lithologic subdivisions. In succeeding years, the chronostratigraphic terminology was gradually abandoned in favour of lithostratigraphic and lithologic subdivisions. By the 1930s, F.J. Pettijohn (Pettijohn 1935) had applied a mix of chronostratigraphic and lithostratigraphic terminology in the Vermilion Township area by dividing the Abram Series into a lower conglomeratic formation and an upper greywacke formation. Lithostratigraphic terminology was applied to the Proterozoic Animikie Group (Tanton 1939) at this time. Locally, some supracrustal sequences such as the Steep Rock Series became known as the Steep Rock Group (Jolliffe 1955, 1966). The Steep Rock Group was later subdivided into 4 formations including the basal Wagita Formation (conglomerate), Mosher Formation (carbonate), Jolliffe Formation (ironstone) and Dismal Formation (ultramafic lapilli tuff; Wilks and Nisbet 1988). Following publication of the *Code of Stratigraphic Nomenclature* in 1961 (American Commission on Stratigraphic Nomenclature 1961), it became increasingly common for various chronostratigraphic series to be renamed as lithostratigraphic groups (e.g., Hodgkinson 1968). For example, the Seine Series became known as the Seine Group.

Perhaps the widest usage of lithostratigraphic classification was that of Blackburn et al. (1991) who systematically subdivided greenstone sequences of the western Wabigoon Subprovince into groups, some of which were further divided into formations. Unfortunately, although various groups and formations of Archean supracrustal strata can be defined locally, few can be traced for more than a few tens of kilometres or correlated from greenstone belt to greenstone belt. This has resulted in limited application of lithostratigraphic terminology to Archean supracrustal strata. An exception is the Seine Group for which distinct coarse clastic sedimentary facies lithology, greenschist metamorphic grade and unconformable relation to underlying supracrustal strata allows isolated units of this group to be recognized and correlated widely in the central Wabigoon Subprovince area.

THE SUBPROVINCES

With the gradual accumulation of regional geologic data, geologists in the mid-twentieth century began to recognize large-scale subdivisions of the Canadian Shield. M.E. Wilson (Wilson 1939b) provided an early mainly geographic subdivision in which he distinguished the Arctic, Churchill, Ungava and St Lawrence provinces of the Canadian Shield. The St. Lawrence Province (which would eventually become known as the Superior Province) was further subdivided into the Northwest, Southern, Timiskaming and Grenville Subprovinces. Later, Gill (1949) used structural trends to define the Slave, Churchill, Ungava, Superior and Grenville provinces. Gill (1949) distinguished several structural subprovinces that he termed "plains", "belted plains" and "mountain belts" within the Superior Province. Many of Gill's subprovinces correspond with areas of Proterozoic rocks.

Also in the mid-twentieth century, techniques for determining the absolute ages of rocks based on the nuclear decay of naturally occurring radioactive isotopes were developed and refined. The radiometric age determinations provided a new set of data (rock ages) for the subdivision of Precambrian rocks. Radiometric age determinations were initially made mainly by the potassium–argon (K/Ar) and rubidium–strontium (Rb/Sr) methods. Although it would be shown later that these ages tend to reflect cooling and metamorphic events more so than crystallization ages for igneous rocks (Beakhouse, McNutt and Krogh 1988), the K/Ar and Rb/Sr methods nonetheless provided estimates of Archean orogenic events. By 1964, H.C. Stockwell (Stockwell 1964) had compiled extensive K/Ar age determinations and used these to define several orogenies or periods of mountain building in the Canadian Shield.

Stockwell (1964) subdivided the Superior Province into subprovinces by distinguishing the belt-like English River and Quetico subprovinces on the basis of easterly structural trends and a predominance of paragneiss within the broader volcano-plutonic Cross Lake Subprovince. In succeeding years, Douglas (1973) defined the Sachigo, Berens River, Uchi, Wabigoon and Wawa–Abitibi subprovinces and Goodwin (1978) provided detailed descriptions of the lithology, metamorphism, structure and metallogeny of the subprovinces and described subprovince boundaries. Beakhouse (1977) and Card and Ciesielski (1986) further refined the names and extent of subprovinces in the Superior Province. Card and Ciesielski (1986) identified 4 types of subprovinces distinguished by predominantly plutonic, volcano-plutonic, sedimentary and high-grade gneiss lithologies. Later, Percival et al. (1992) substantially modified the pattern of subprovinces in northern Quebec based on mapping.

By the 1980s, the subprovinces were widely recognized as regional-scale elements of the Archean Superior Province and were typically cited by many authors to describe the location of their study area. Although subprovinces were distinguished on the basis of several factors including structural trends, metallogenic characteristics and magnetic fabric, the predominant lithology (e.g., plutonic, volcano-plutonic and sedimentary) is the main distinguishing feature. Subprovince boundaries were defined mainly by lithologic transitions locally accentuated by major transcurrent faults (Mackasey, Blackburn and Trowell 1974; Schwerdtner et al. 1979) and changes in metamorphic grade. Elsewhere, subprovince boundaries were drawn at intrusive contacts or lithologic transitions (Goodwin 1978).

TECTONIC SUBDIVISIONS: THE SUPERIOR PROVINCE

The development of high-precision U/Pb dating methods (Krogh 1982) and the gradual accumulation of a database of U/Pb age determinations provided new insight on the distribution of various generations of Archean rocks. At a local scale, many greenstone belts that were previously thought to represent homoclinal strata were shown to be composed of several sequences of volcanic rocks varying in age by hundreds of millions of years (e.g., Corfu and Wallace 1986). These supracrustal packages comprising parts of greenstone belts became known as tectonic assemblages (Ontario Geological Survey 1992; Thurston 1991). Thurston (1991) defined a tectonic assemblage as stratified volcanic and/or sedimentary rock units built during a discrete interval of time in a common depositional or volcanic setting. The rock units typically share a common or dominant lithofacies; they may also share some additional attributes, such as structural, metamorphic, geochemical and geophysical features. Thurston and Chivers (1990) identified 4 types of tectonic assemblages including 1) platform-type assemblages dominated by quartz arenite and carbonate-bearing sequences; 2) mafic plain assemblages dominated by mafic to ultramafic sequences and deep-water argillites; 3) mafic to felsic volcanic assemblages typically consisting of deep-water tholeiitic flows overlain by shallow-water intermediate to felsic pyroclastics and flows; and 4) "Timiskaming-type" assemblages of fluviatile sediments and calc-alkaline to alkaline volcanic rocks. It was generally recognized that at least some assemblages must have been tectonically juxtaposed.

At a regional scale, U/Pb geochronology showed age variations of up to 500 million years from 3.2 Ga to 2.7 Ga for rocks in various subprovinces (*see* summary of Corfu and Davis 1992). Older rocks (pre 3.0 Ga) were identified within the central Sachigo, Winnipeg River and Minnesota River Valley subprovinces of Card and Ciesielski (1986), whereas other subprovinces such as the English River, Wabigoon and Quetico contained mainly younger (circa 2.7 Ga) ages. These variations in age and lithology from one subprovince to another, together with the overall striped pattern of subprovinces, are striking features of the western Superior Province. By the 1980s, geologists were considering various tectonic models to explain development of the subprovince architecture within the world's largest Archean craton.

The origin of continents or sialic crust was considered by Lawson (1932), Shand (1943) and Buddington (1943) and discussions in the middle part of the twentieth century (e.g., Umbgrove 1947) typically addressed whether a sialic crust had developed when the earth formed or progressively after formation of the earth. In the 1940s, continents were considered to have formed by removal of part of an original thick sialic crust of the earth, by coalescence of a thin primordial sial or by nucleation of continental masses within a simatic crust and progressive outward growth of the continents through time. In reviewing models for the origin of continents, Wilson (1949) cited the available information on the distribution of ages of rocks in the Superior Province as evidence that continents must have grown through time. Building upon models of Lake (1931) and Vening Meinesz (1947), J.T. Wilson studied island arcs (Wilson 1950) and the motion and interaction of oceanic and continental crust (Wilson 1963, 1965) and eventually proposed the modern theory of plate tectonics (Wilson 1973).

Wilson's hypothesis held that convection plumes rise from the lower mantle to spread out in the asthenosphere imposing a shear-couple on the base of the crust, which results in slow motion of crustal plates. Where 2 plates approach each other at a subduction zone, the continental plate will generally override the oceanic plate. Collisions between 2 continental plates result in Appalachian-type mountain building and cause overall accretionary growth of the continental mass. Eventually, where a continental mass migrates over a rising mantle plume the continent can be rifted or torn apart producing a new ocean to complete the "Wilson cycle" (e.g., Burke and Dewey 1973).

Although plate tectonic theory was well established for modern orogens by the 1970s, its application to the Archean was somewhat slower. At a local scale, Archean granite-greenstone terrains are characterized by mainly subvertical folds, faults and mineral fabrics. These structures were interpreted to have developed due to vertical displacements probably resulting from gravitational sinking of dense material such as greenstone belts and diapiric rise of low-density material such as granite (e.g., Eskola 1948; Ramberg 1981). Greenstone belts were considered to have developed by eruption of mafic lavas along fissures in a thin sialic crust. The mafic material underwent subsequent gravitational induced downwarping attendant with diapiric rise of granitic plutons to produce the characteristic pattern of thin and variably curved and tapered greenstone belts interspersed with oval plutonic masses (Anhaeusser et al. 1969).

Clues to the role of horizontal deformation and accretionary growth of the Superior craton through plate tectonic processes came initially from regional-scale considerations. Noting the characteristic striped pattern of subprovinces, Goodwin (1968) proposed that the Superior Province could have grown by accretion of juvenile oceanic material. Talbot (1973) also presented a model of Archean crustal development based on plate tectonic processes and Langford and Morin (1976) noted similarities in geometry, scale and lithology between the subprovinces of the western Superior craton and major crustal belts of the western Canadian cordillera. In correspondence with an island-arc model for the cordillera, Langford and Morin (1976) proposed that the Uchi, Wabigoon and Wawa subprovinces represented island arcs that moved northward against the Berens River Subprovince. The sedimentary English River and Quetico subprovinces were interpreted to represent sedimentary basins trapped between the merging arcs.

The application of plate tectonic models to the Superior Province advanced rapidly by the 1990s. Hoffman (1988, 1989) proposed that the North American craton (Laurentia) is an aggregate of 7 former microcontinents, one of which is the Superior Province. These Archean microcontinents were welded together by Paleoproterozoic collisional orogens. Within the Superior Province, Percival and Williams (1989) proposed that the Quetico sedimentary sequences represent detritus eroded from adjacent Wabigoon and Wawa volcanic arcs and deposited on oceanic crust that was being subducted northward beneath the Wabigoon arc. The sediments were subsequently compressed as the Wawa arc docked against the Wabigoon arc at about 2.695 Ga after which strike-slip faulting occurred at the Quetico–Wabigoon boundary and Quetico sedimentary rocks were metamorphosed. Williams (1990) advocated an arc–arc accretion model for development of the Wabigoon, Quetico and Wawa subprovinces based in part on imbricated faults and folds at the Quetico–Wabigoon subprovinces boundary in the Beardmore–Geraldton area. Card (1990) and, later, Percival et al. (1994) applied accretionary tectonic models to the entire Superior Province.

The accumulation of geologic and geochronologic data for the far northern part of the Superior Province (e.g., Davis and Moore 1991) provided a basis for revised tectonic models within the broad Sachigo Subprovince. Thurston, Osmani and Stone (1991) noted a preponderance of 3.0 Ga ages for rocks in the North Caribou Lake area and proposed that the southern part of the Sachigo Subprovince represents an ancient crustal block (North Caribou terrane) against which other terranes were accreted to both the south and north sides. In this context, many of the former subprovinces became known as terranes in which a terrane is defined as a fault-bounded package of strata that is allochthonous to, and has a geologic history distinct from, the adjoining geologic units (Thurston 1991).

Stott (1997) reviewed the geology of the Superior Province and proposed that certain blocks of crust or superterranes such as the combined Uchi, Berens and North Caribou terranes (North Caribou Superterrane) developed as independent microcontinents and were subsequently amalgamated together with other island arcs and sedimentary prisms into the Superior craton. Areas of high-grade gneiss such as the Pikwitonei and Kapsukasing zones were interpreted as sections of lower Archean crust uplifted during Neoarchean to Proterozoic orogenesis.

Further mapping and geochronologic studies (Böhm et al. 2000; Skulski et al. 2000; Stone 2005b) identified the Paleoarchean Northern Superior Superterrane and added detail to the distribution of terranes in the far northern Superior Province (*see* Figure 1). In reviewing tectonic evolution of the western Superior

Province, Percival et al. (2006) identified 5 accretionary events between 2.72 and 2.60 Ga during which fragments of continental and oceanic crust were assembled through plate-tectonic processes into the coherent Superior craton. On the basis of samarium–neodymium (Sm/Nd) isotopic studies, volcanic rocks of mainly oceanic origin such as in the western Wabigoon and Oxford–Stull terranes were distinguished from volcanic rocks of mainly continental origin such as in the North Caribou Superterrane. Also, certain blocks of crust previously thought to be fault-bounded allochthonous terranes were reinterpreted as having developed autochthonously and were redefined as tectonic domains. A tectonic domain is considered to be a volume of rock, bounded by compositional or structural discontinuities, within which there is structural and chronologic homogeneity. A tectonic domain may contain both plutonic and supracrustal rocks, is generally smaller than a subprovince and may be either allochthonous or autochthonous with respect to adjoining geologic units. An example includes the Uchi domain (Stott and Rainsford 2006) that appears to have developed over a 300 million year period on the margin of the North Caribou microcontinent. Stott and Rainsford (2006) extended the terrane/domain architecture of the Superior Province under the Paleozoic cover rocks of the Hudson Bay Lowland based on geophysical interpretation.

TECTONIC SUBDIVISIONS: THE WABIGOON SUBPROVINCE

By 1990, the Wabigoon Subprovince had been interpreted as a volcanic arc that developed through eruption of mafic to felsic lavas from about 2.75 to 2.72 Ga followed by plutonism as late as 2.69 Ga. Sedimentary rocks of the Quetico Subprovince were rapidly eroded from the Wabigoon and Wawa arcs and accreted to the Wabigoon volcanic arc at about 2.70 Ga (Percival and Williams 1989; Williams 1990). To a large extent, the available geochronology supported this interpretation although some dates hinted at older crustal material within the central Wabigoon Subprovince area. The older crustal material could not be easily explained within the context of a tectonic model involving the accretion of a sedimentary prism to a volcanic arc.

Supracrustal and plutonic rocks of the Kakagi–Savant lakes volcanic belt, or what is presently known as the western Wabigoon terrane, were extensively dated in the 1980s. Although basal mafic sequences at Savant Lake and Sturgeon Lake had ages of 2745 and 2775 Ma, the majority of volcanic rocks were shown to have erupted between 2735 and 2702 Ma. Plutonism began contemporaneous with volcanism and protracted after volcanism including intrusion of tonalitic plutons at 2732 to 2716 Ma, granodioritic plutons at 2709 to 2699 Ma and late monzodiorite plutons at 2688 to 2686 Ma (Davis and Trowell 1982; Davis, Blackburn and Krogh 1982; Davis et al. 1985; Davis and Edwards 1986; Davis, Sutcliffe and Trowell 1988; Davis, Poulsen and Kamo 1989; Davis, Pezzutto and Ojakangas 1990). Studies of detrital zircon grains by the above authors showed that most major sedimentary sequences were deposited late in the evolution of the area. For example, the youngest detrital zircon grains, which are equated with the maximum depositional age are 2698 Ma for sedimentary units near Sioux Lookout, 2704 Ma for the Coutchiching sequences at Mine Centre, 2696 Ma for the Seine conglomerate at Mine Centre and 2698 Ma for Quetico sediments at Atikokan.

Davis, Sutcliffe and Trowell (1988) obtained ages of 3051 and 3075 Ma for tonalite gneiss at Tannis Lake (north of Kenora) and Caribou Lake (north of Armstrong). Although these localities would later be shown as part of an old crustal block represented by the Winnipeg River Subprovince (e.g., Melnyk et al. 2006), old rocks were also found in the central Wabigoon Subprovince area. Davis and Jackson (1988) obtained ages of 3000 and 2999 Ma for volcanic sequences of the Lumby Lake greenstone belt and an age of 3003 Ma for tonalite gneiss of the Marmion batholith. Davis (1993) and Stone, Kamineni and Jackson (1992) reported ages of 2931 to 2936 Ma for felsic tuffs of the western Finlayson Lake greenstone belt and tonalite and tonalite gneiss of the adjacent Dashwa gneiss complex. Supporting evidence for an old protolith came from detrital zircon grains as old as 3059 Ma in Coutchiching sediments at Mine Centre and 3007 Ma in Quetico sediments at Atikokan (Davis, Poulsen and Kamo 1989; Davis, Pezzutto and Ojakangas 1990).

These data provided clear evidence for fairly young (2.72 Ga) volcanic sequences within the Kakagi Lake–Savant Lake greenstone belt and old (3.01 to 3.0 Ga) crustal material within the Winnipeg River and central Wabigoon subprovinces areas. In a seminal interpretation, Davis, Sutcliffe and Trowell (1988) noted a lack of inheritance in volcanic sequences of the Kakagi Lake–Savant Lake greenstone belt and suggested that these volcanic rocks had developed in a simatic environment after which they were tectonically juxtaposed with older rocks of the Winnipeg River and the central Wabigoon subprovince areas. This interpretation added new complexity to the Wabigoon Subprovince, which was shown to be divisible into the juvenile western Wabigoon terrane (Kakagi Lake–Savant Lake greenstone belt), the central Wabigoon Subprovince area and the eastern Wabigoon Subprovince area lying east of Lake Nipigon.

By the mid 1990s, the central Wabigoon Subprovince had become one of the more enigmatic areas of the Superior Province on the basis of data that was available at the time. The Wabigoon Subprovince was known to contain a diverse mixture of rocks including older 2.93 to 3.0 Ga largely sialic material, as well as 2.72 Ga mainly simatic greenstone sequences, and yet little was known of the extent and boundary relations of the older crust. Were there other magmatic events between 2.93 and 2.72 Ga represented in the area? What are the ages and tectonic significance of platform sequences such as the Steep Rock Group and greenstone belts such as the Heaven Lake, Garden Lake and Obonga Lake greenstone belts, which apparently occur within the older crustal domain? Is the central Wabigoon domain an extension of the Winnipeg River Subprovince or, if not, where is the boundary between the two? Finally, if the Wabigoon Subprovince is not a simple volcanic arc, then what are its various components and how and when were they assembled?

The questions outlined in the previous paragraph are the basis for the present study that was initiated in the mid 1990s and continued for a decade with emphasis on bedrock mapping, geochemistry and geochronology. In the meantime, other research including neodymium (Nd) isotopic studies and geochronology shed new light on tectonic evolution of the central Wabigoon Subprovince area. For example, the Nd isotopes of the 2.70 to 2.74 Ga volcanic rocks of the Lake of the Woods greenstone belt (western Wabigoon terrane) are characterized by positive ε_{Nd} values (Ayer and Dostal 2000). These data support the earlier interpretation of Davis, Sutcliffe and Trowell (1988) in which rocks of the western Wabigoon terrane were derived from an isotopically juvenile source and probably erupted in a simatic environment. Similarly, the Nd isotopic study of Henry et al. (2000) showed that the gneisses and sedimentary rocks of the 2.9 to 3.0 Ga Finlayson Lake greenstone belt and Marmion batholith plot close to juvenile depleted mantle values at 3.0 Ga (i.e., they have positive ε_{Nd} values) indicating that they were largely derived from new crust at that time. In contrast, the pre 3.0 Ga gneisses of the Winnipeg River Subprovince have generally negative ε_{Nd} values indicating that they could have assimilated older LREE-enriched crustal material of approximately 3.4 Ga age. The work of Henry et al. (2000) provided evidence that the old rocks in the Wabigoon Subprovince had evolved independent of those in the Winnipeg River Subprovince and, therefore, in a tectonic sense, these crustal blocks probably represent separate microcontinental fragments that were joined at some time after 3.0 Ga.

In the late 1990s, K.Y. Tomlinson studied the central Wabigoon Subprovince area as part of the Lithoprobe project and added to the geochemical, geochronological and isotopic data for the area. Tomlinson et al. (1999) analyzed the geochemistry of mafic to ultramafic volcanic rocks in the Steep Rock Lake, Finlayson Lake and Lumby Lake greenstone belts and proposed that volcanism resulted from one or more mantle plumes rising and impinging on the base of the crust. Tomlinson et al. (2003) provided extensive new age determinations for greenstone belts and plutonic rocks of the central Wabigoon Subprovince area (Table 2 discussed below). Among these are the ages of 2727 Ma for the Lac des Iles greenstone belt, 2729 and 2954 Ma for the Whistle and Whitton assemblages of the Heaven Lake greenstone belt, 2726 Ma for the Garden Lake greenstone belt, 2956 Ma for the Phyllis Lake greenstone belt, new ages including 2828 Ma for the Lumby Lake greenstone belt, 2926 Ma for the western Steep Rock Lake greenstone belt, 2830 Ma for the Lac des Mille Lacs greenstone belt, and a maximum age of 2780 Ma for the Ashrock assemblage of the Steep Rock Lake greenstone belt. On the basis of Nd isotope systematics, Tomlinson et al. (2004) defined the Marmion terrane to represent 3.0 Ga plutonic and supracrustal sequences of the south-central Wabigoon Subprovince area (*see* Figure 1). The Winnipeg River terrane was extended around the eastern end of the western Wabigoon terrane at Savant Lake to include northern parts of the central Wabigoon Subprovince area. The boundary between the Marmion and Winnipeg River terranes was placed south of the Garden Lake greenstone belt in view of isotopic evidence that magmas in the Garden Lake, Obonga Lake and Caribou Lake greenstone belts had interacted with older crustal material typical of the Winnipeg River terrane.

In summarizing work in the Savant Lake–Sturgeon Lake greenstone belt, Sanborn-Barrie and Skulski (2006) interpreted the Jutten assemblage at Savant Lake (quartzite and conglomerate overlain by mafic lavas) as a continental margin deposit developed between 2880 and 2750 Ma on a basement represented by the >3.0 Ga Winnipeg River continental terrane. These authors reiterated the interpretation that the western Wabigoon terrane including 2745 to 2718 Ma volcanic rocks of the Savant Lake–Sturgeon Lake greenstone belt (Handy Lake and South Sturgeon assemblages) represent an ocean island arc–rift complex formed on 2775 Ma oceanic crust (Fourbay assemblage). Collision of the western Wabigoon arc with the Winnipeg River continent is bracketed between 2703 Ma (the maximum age of marine forearc deposits of the Warclub group) and 2696 Ma (the minimum age of a late tectonic pluton).

In succeeding sections, the geology, geochronology and thermobarometry resulting from this study are reviewed. Later, in “Structural and Magmatic Evolution of the Central Wabigoon Subprovince Area”, this information is assessed in terms of the new insight that it provides on the structural and magmatic evolution of the central Wabigoon Subprovince area.

Geochronology

Age determinations were made on approximately 120 rock samples collected within or near the central Wabigoon Subprovince area. The results of the geochronologic investigations (Table 2) are compiled from earlier studies as well as a series of annual reports issued by the Jack Satterly Geochronology Laboratory at the University of Toronto for samples collected as part of the present study. In the majority of cases, ages are determined for zircon grains, which were separated from the rock and analyzed by U/Pb systematics using isotope dilution thermal ionization mass spectrometry (ID–TIMS) at the Jack Satterly Geochronology Laboratory. About 20 samples were analyzed using sensitive high-resolution ion microprobes (SHRIMP) at the laboratories of the Geological Survey of Canada in Ottawa and the Chinese Academy of Geological Sciences in Beijing. The analytical techniques and interpretations are discussed in the references cited in Table 2, with the exception of 5 samples analyzed at the Beijing SHRIMP Center discussed further below.

The geochronologic data are keyed by number in Table 2 to locations shown on Figure 3 (Chart A, back pocket) and on Map P.2229 (back pocket). The geochronologic data include the age of crystallization for intrusive and extrusive magmatic rocks. In some samples, zircon grains or parts of zircon grains, which are older than those representing the crystallization age are identified. The ages of these grains (*see* Table 2, column “Inherited Age”) are interpreted to represent refractory remnants of older crustal material that was assimilated by the magma. In other instances, the age of metamorphism (*see* Table 2, “Metamorphic Age”) is interpreted from U/Pb age determinations on titanite and from zircon grains with low Th/U ratios as well as the rims of zircon grains. Detrital zircon grains are analyzed from clastic sedimentary rocks and help to constrain the age at which the sediments were deposited. The youngest age obtained for a suite of detrital grains (*see* Table 2, “Detrital Grain Age”) represents the maximum age at which the sedimentary rock was deposited.

Figure 3 shows the interpreted variation in age of rocks through the central Wabigoon Subprovince area based on available geochronology (*see* Table 2). The age domains are discussed further in the sections describing various greenstone belts and plutonic suites and in "Structural and Magmatic Evolution of the Central Wabigoon Subprovince Area".

Five samples were initially processed using equipment of Laurentian University in Sudbury and were subsequently analyzed using the SHRIMP at the Beijing SHRIMP Center, Institute of Geology, Chinese Academy of Geological Sciences. The samples were crushed using a jaw crusher followed by a disk mill. Initial separations of heavy minerals was carried out by multiple passes on a Wilfley table after which the concentrate was further refined by density separation using methylene iodide to concentrate zircon. Final sample selections were hand picked under a microscope to choose the freshest least-cracked grains of zircon.

Zircon grains were analyzed using the SHRIMP II analytical facility at Beijing SHRIMP Center on April 16, 2007. Zircon grains were cast in epoxy grain mounts (BJ-SHRIMP mount G1688), and polished sufficiently with diamond compound to reveal the grain centres. The mount was then cleaned, coated with approximately 10 nm of high-purity gold, and individual zircon grains were imaged with a Hitachi S-3000N scanning electron microscope equipped with Gatan cathodoluminescence (CL) detector, as well as a petrographic microscope (for transmitted and reflecting images) to identify complexities of internal structure and compositional zoning, cracks, inclusions and fractures. Reference standards used were SL13, with an age of 572 Ma and U content of 238 ppm and TEMORA with $^{206}Pb/^{238}U$ isotope dilution age of 417 Ma (Williams 1998; Black et al. 2003). Uranium-lead isotopic spot analyses within the zircons were carried out following the general SHRIMP methods outlined in Williams (1998) and Wan et al. (2005). Data were acquired using a mass-filtered O_2^- primary beam and apertures of sufficient size to effect sputtering diameter of approximately 30 μm. Primary O_2^- beam currents during these sessions were roughly 7 to 8 nA, and mass resolution was routinely around 5000. Peak count rates were measured sequentially over 9 mass stations for Zr^+, Pb^+, U^+, UO^+, ThO^+ and Th^+ (plus background) averaged through 5 scans, using an electron multiplier in pulse-counting mode. Results of the SHRIMP analyses are summarized in Table 3 and concordia diagrams (Figure 4) were produced using the software SQUID 1.02 with ISOPLOT, version 2.49 (Ludwig 2001).

Sample 06DS10 represents an intermediate volcanic breccia from the Raleigh Lake greenstone belt. The heavy mineral concentrate is rich in pale green epidote and honey-brown titanite with about 1% zircon. The zircon grains are small, clear to pale pink and stubby to elongate and prismatic. Cathode luminescent images show oscillatory zonation within the grains; a few grains have dark cores and others have thin bright rims. Uranium-lead dating of 11 spots shows broadly similar isotopic characteristics for dark cores of grains and zoned grains (*see* Table 3; Figure 4a). Most analyses vary from slightly discordant through concordant to reversely discordant and appear to define a single population. The most strongly discordant analysis (06DS10-9) represents a spot on the edge of a grain and is excluded. The 9 most concordant analyses give an average age of 2730±7 Ma (*see* Table 2, No. 82), which is interpreted to represent the crystallization age of the rock. This age is similar to those for rocks in the Lower Keewatin Assemblage at Lake of the Woods (Ayer and Davis 1997) and, together with lack of evidence for older inherited material, suggests that the Raleigh Lake greenstone belt is part of the juvenile western Wabigoon terrane.

Sample 06DS22 represents a large mass of biotite tonalite at the western margin of the central Wabigoon Subprovince area. The sample yielded an epidote-rich heavy mineral concentrate with a few percent pyrite and about 1% zircon. The zircon grains are small elongate doubly terminated prisms with a few oval to football shapes. Grains are mainly clear to smokey-pink with local reddish stain and dark inclusions. Cathode luminescent images show oscillatory zonation with no clear evidence of cores in the zircon grains. Some grains have bright rims and patches. The bright rims, represented by analysis 06DS22-21 of Table 3 provide unreliable age determinations due to extremely low U and Th concentrations. Seven analyses of grains with diverse morphologies as well as cores and rims of grains are variably discordant, but appear to represent a single population of grains (Figure 4b). Five analyses, comprising a collinear array of data points provide a mean age of 2716±6 Ma (*see* Table 2, No. 83). This age is representative of the ages of plutonic rocks in the western Wabigoon terrane (*see* Blackburn et al. 1991, Table 9.8). Together with the lack of evidence for inheritance of older crustal material, this age is interpreted to indicate that the tonalitic mass at the northwestern margin of the central Wabigoon Subprovince area is part of the western Wabigoon terrane. Alternatively, these comparatively young plutonic rocks could represent late magmatism on the western margin of the central Wabigoon domain. A careful search for older inherited material within these 2.71 Ga tonalitic rocks might help to resolve whether they are allochthonous or autochthonous.

The Entwine intrusion of the sanukitoid suite is situated at the western margin of the central Wabigoon Subprovince area within the tectonic boundary zone between the western Wabigoon terrane and older crustal blocks to the east. Sample 06DS23 represents a strong horizontally lineated monzodioritic phase of the Entwine intrusion that produced a heavy mineral concentrate rich in apatite with subequal components of amphibole, ilmenite, titanite, epidote and zircon. The zircon grains are all angular fragments possibly representing chunks of large crystals that were broken in the disking process. Cathode luminescent images show complex zonation within the grain fragments and a few grains have bright rims. The distribution of ages appears to be simple. Five analyses representing dark and light zoned parts of grains give approximately overlapping ages with an average of 2685±8 Ma (*see* Table 2, No. 120; Figure 4c). This age is typical of other late plutonic rocks of the central Wabigoon Subprovince area and the widespread linear fabric indicates that the Entwine intrusion was probably deformed during crystallization.

The oval Joe pluton of biotite granodiorite composition intrudes tonalite gneisses and Quetico-type sedimentary sequences at the south margin of the central Wabigoon Subprovince area. Sample 06DS50 of the Joe pluton yielded a heavy mineral concentrate rich in ilmenite with a small amount of zircon. The zircon grains are tiny clear to pale pink cracked elongate prisms with a few larger cloudy and stubby grains. Cathode luminescent images show complex zoned patterns within the grains. A few grains have thin bright rims. Due to the small size and acicular shape of the grains, the analyses were constrained to a variety of locations along the axes of grains ranging from the centres to the tips of clear and crack-free grains. The isotopic ratios, listed in Table 3, are scattered probably due to alteration and can be arranged into 2 groups according to Th/U ratios. Three analyses with low Th/U are reversely discordant (Figure 4d). Among the analyses with higher Th/U ratios, 2 are nearly concordant, whereas 2 are strongly discordant perhaps due to alteration-induced lead loss. A subset including the nearly concordant analyses and the reversely discordant grains define a linear array and yield an average age of 2686±6 Ma (*see* Table 2, No. 113). This age is geologically reasonable because the Joe pluton intrudes Quetico-type sedimentary sequences whose maximum depositional age is 2698 Ma (Davis, Pezzuto and Ojakangas 1990). The young age of this intrusion suggests that it is part of the posttectonic biotite granite suite (*see* Table 2).

Sample 06DS52 represents foliated biotite tonalite of the Hillyer domain in the south-central Wabigoon Subprovince area. The sample gave an epidote-rich concentrate with a few percent zircon. The zircon grains are small and somewhat oval to stubby and elongated in shape. The colour of grains ranges from clear to pale pink in small varieties to cloudy reddish in larger grains. The larger grains tend to have cracked rims. Cathode luminescent images show regular dark and light zoned patterns possibly representing magmatic zonation in many grains and small oval domains possibly representing inherited cores are identified in a few grains. The isotopic ratios for the analyses of sample 06DS52 are variable (*see* Table 3). A group of 5 analyses representing small clear zoned grains are nearly concordant and define a mean age of 2866±8 Ma (Figure 4e). Another group of 5 analyses representing oval grains and possible cores of larger grains are variably discordant and scatter to the left in Figure 4e. The crystallization age of the rock is interpreted to be 2866±8 Ma (*see* Table 2, No. 42) because this mean age is obtained from a population of small clear zoned grains with no overgrowths and the least evidence of alteration. The other group of analyses with younger intercept ages correspond with relatively more altered parts of zircon grains although analysis 06DS52-49 has a distinctly high Th/U ratio and possibly indicates a younger magmatic event approximately coeval with emplacement of the nearby Joe pluton at 2686 Ma. The discordant analysis 06DS52-50 has a low Th/U ratio and possibly represents a metamorphic zircon.

Table 2. U/Pb geochronology data for volcanic, plutonic and sedimentary units of the central Wabigoon Subprovince area.

No. [1]	Sample Number	Area	UTM Easting (m)	Location Northing (m)	Rock Type	Greenstone Belt or Domain	Assemblage or Plutonic Suite	Age of Crystallization	Detrital Grain Age, Max. Depositional Age (in bold)	Inherited Age	Metamorphic Age	Ref. (*see* end of table)
Marmion Domain												
1	DD84-10	S of Lumby Lake	624430	5427561	tonalite gneiss	Marmion domain	Tonalite Gneiss	3009±8				1
2	DD85-10	N of Steep Rock Lake	598425	5409110	mafic (hornblende) tonalite	Marmion domain	Hornblende Tonalite	3001.6+3.4/–2.3				1
3[r]		Little Falls Lake	596880	5409971	sandstone	western Marmion domain	Finlayson East		**2997.1±0.9**			3
4	Z1	Finlayson Lake	601040	5414724	conglomerate, sandstone	western Marmion domain	Finlayson East		3002, 3001, 3001, 2999, 2999, **2997**			2
5	05DS126	Elizabeth property	593700	5406040	quartz-feldspar porphyry dacite or subvolcanic intrusion	western Marmion domain	Finlayson East	2999.4±0.7				4
6	Sample 1	W of Lumby Lake	621029	5432763	felsic tuff	Lumby Lake	Lumby South	2999.4±2.9				5
7	DD97-4	Lumby Lake	624030	5432510	rhyolite	Lumby Lake	Lumby South	3001.1±1.1				1
8	Sample 2	Jefferson Lake	632109	5432335	quartz porphyry	Lumby Lake	Lumby South	2999.1±2.0				5
9	DD97-5	S of Pyramid Lake	645328	5441694	rhyolite	Lumby Lake	outlier of Marmion domain	3013.6±1.3				1
10	06DS81	E of Finlayson Lake	601160	5413760	quartz porphyritic felsic volcanic	western Marmion domain	Finlayson East	3003.3±0.7				19
11	06DS80	N of Crooked Pine	639112	5406190	leucogabbro	Marmion domain	mafic tonalite/gabbro	3004.5±1.1				19
12	07DS04	Little Dragon Lake	583601	5409319	felsic volcanic	western Marmion domain	unknown assemblage	2998.5±0.8				20
13	07DS39	E of Crooked Pine	647759	5407779	quartz-porphyry felsic volcanic or intrusion	Marmion doman	unknown assemblage	3007.6±1.2				20
14	07DS36	W of Owl Lake	606516	5457250	tonalite gneiss	Marmion domain	Tonalite Gneiss	2998.1±0.9				20
15	07DS46	N of Steep Rock Lake	598541	5409158	gabbro dike	Steep Rock Lake (Marmion domain)	dike cuts Marmion batholith	3002.1±0.7				20
16[r]	00KYT-25	W of Raven Lake	622735	5461789	tonalite gneiss	western Marmion domain	Tonalite Gneiss	2989 +5/–3				6
17	07DS40	Crooked Pine Lake	633232	5405360	felsic fragmental	Lac des Mille Lacs (Marmion domain)	unknown assemblage	3006.1±0.7				20
18	07DS47	S of Lumby Lake	623608	5429579	hornblende tonalite	N rim of Marmion batholith	Hornblende Tonalite	2998.6±0.7				21

Notes: UTM co-ordinates are NAD83, Zone 15 (except ***, *see* below);
[1] refers to numbers on Figure 3 (*see* Chart A, back pocket) and Preliminary Map P.2229 (*see* back pocket);
[r] location approximate.

Table 2. continued.

No. [1]	Sample Number	Area	UTM Easting (m)	Location Northing (m)	Rock Type	Greenstone Belt or Domain	Assemblage or Plutonic Suite	Age of Crystallization	Detrital Grain Age, Max. Depositional Age (in bold)	Inherited Age	Metamorphic Age	Ref. (*see* end of table)
Whitton Domain												
140[t]	96KYT-249	N of Whitton Lake	756394	5476046	felsic dike	Heaven Lake (Whitton domain)	Whitton	2953.1±1.6				1
141[t]	96KYT-252	N of Whitton Lake	757844	5475642	felsic tuff	Heaven Lake (Whitton domain)	Whitton	2953.7±1.3				1
142	96KYT-50	Tiny Lake	643036	5440069	felsic tuff	Lumby Lake (Whitton domain)	Lumby North	2963.3±4.5		3016±5		1
143	00KYT-24	N of Phyllis Lake	620809	5454850	felsic tuff	Phyllis Lake (Whitton domain)	Phyllis	2955.8±0.8				1
144	06DS78	W of Red Paint	616363	5437305	tonalite gniess	Whitton domain/ western Wabigoon terrane?	Tonalite Gneiss	2713±4[s]		2950±15[s]	2700±8[s]	19
145	07DS35	N of Trewartha Lake	661214	5445587	biotite tonalite	Whitton domain/ Marmion domain	Biotite Tonalite	2952±4		2993±11		20
146	08DS20	N of Turtle Lake	582280	5425092	tonalite gniess	Whitton domain	Tonalite Gneiss	2957±1				21
Central Wabigoon Domain												
22	03DS92	S of Garden Lake	742745	5486637	biotite tonalite	eastern central Wabigoon	Biotite Tonalite	2924.1±9			2690 (titanite, lower intercept)	8
23	03DS93	E of Armistice Lake	733130	5474740	tonalite gneiss with wacke and gabbro	eastern central Wabigoon	Tonalite Gneiss	2937.1±1.2*			2722 (monazite)	7
24	05DS56	NW of Graham	659562	5465150	biotite tonalite	central Wabigoon	Biotite Tonalite	2929±4				4
25	DD88-15	Finlayson Lake	604450	5420873	felsic tuff	Finlayson Lake	Finlayson West	2931.4±2				1
26		NW of Finlayson Lake	602020	5424440	biotite tonalite	central Wabigoon	Biotite Tonalite	2936				9
27		NW of Finlayson Lake	597830	5424500	tonalite gneiss	central Wabigoon	Tonalite Gneiss	2928				9
28	00KYT14	N of Perch Lake	577155	5405380	felsic volcanic fragmental	Perch Lake	unknown assemblage	2925.6±1.2				1
29	00KYT16	N of Mine Centre	519920	5409812	tonalite gneiss	west-central Wabigoon	Tonalite Gneiss	2924±2				6
30	06DS77	N of Red Paint Lake	616997	5441522	biotite tonalite	west-central Wabigoon	Biotite Tonalite	2722±8[s]		2932±5		19
31	06DS73	W of Tib Lake	727850	5457880	biotite tonalite	central Wabigoon	Biotite Tonalite	2917±12[s]				19
Bar Assemblage												
35	PT1	Bar Lake	629412	5434560	felsic tuff	Lumby Lake	Bar	2897.6±2.1		2903 to 2999		1

Table 2. continued.

No. [1]	Sample Number	Area	UTM Easting (m)	Location Northing (m)	Rock Type	Greenstone Belt or Domain	Assemblage or Plutonic Suite	Age of Crystallization	Detrital Grain Age, Max. Depositional Age (in bold)	Inherited Age	Metamorphic Age	Ref. (*see* end of table)
Hillyer Domain												
40	00KYT15	N of Perch Lake	577300	5405566	hornblende tonalite	Hillyer domain margin	Hornblende Tonalite	2883±2				10
41	00KYT11	N of Mine Centre	541452	5415004	hornblende tonalite	Hillyer domain margin	Hornblende Tonalite	2869±4				6
42	06DS52	E of Manion Lake	557750	5415800	biotite tonalite	Hillyer domain	Biotite Tonalite	2866±8				22 (this study)
43	07DS43	west of Manion Lake	541762	5417697	leuco tonalite gneiss	Hillyer domain margin with Marmion domain inheritance	Tonalite Gneiss	~2880		~3000		21
44	08DS27	E of Nevison Lake	585400	5411544	tonalite gneiss	Hillyer/Marmion domains	Tonalite Gneiss	2886±7		2875 to 2965		21
Pinecone-Savoy and Marmion South Margin Domain												
45	DD98-4	Lac des Mille Lacs	693300	5412160	felsic volcanic fragmental	Lac des Mille Lacs belt	unknown assemblage	2830.2±1.1				1
150	00KYT02	S Gargoyle Lake	637385	5437460	felsic tuff in komatiite	Lumby Lake	Pinecone	2828.3±1.8				1
151	05DS57	E of Savoy Lake	650020	5462475	biotite tonalite	Savoy domain	Biotite Tonalite	2814.0±1.2				4
152	05DS60	SE of Raven Lake	630017	5457740	biotite tonalite	Savoy domain	Biotite Tonalite	2817 +4/–2			2694 (titanite)	4
Dog Domain												
55	DD98-6	Lac des Mille Lacs	685620	5418565	hornblende tonalite	Dog domain	Hornblende Tonalite	2780.7±1.8				1
56	05DS85	W of Shelby Lake	736906	5440760	tonalite gneiss	Dog domain	Tonalite Gneiss	2773.6±1.2				4
57	05DS73	S of Buck Lake	723380	5445360	biotite tonalite to granodiorite	Dog domain	Biotite Tonalite	2777±3				4
58	DD85-3	S of Lac des Iles			tonalite gneiss	Dog domain	Tonalite Gneiss	2775±8			2722±3	11
59	06DS74	SW of Buck Lake	715177	5442745	tonalite gneiss	Whistle domain/ Dog domain	Tonalite Gneiss	2729±12[s]		2796±8[s]		19
Steep Rock Assemblage / Dismal Assemblage												
46	04DS57	S of Bedivere Lake	653680	5410603	conglomerate	Lac des Mille Lacs	unknown (possibly Steep Rock assemblage)		3007, 3006, 3004, 3002, 2981, **2927**			7
47		Steep Rock Mine	599316	5409238	tonalite cobble in conglomerate	Steep Rock Lake	Steep Rock		**2994**			9
48		Steep Rock Mine	599320	5406790	sandstone	Steep Rock Lake	unknown (possibly Seine assemblage)		**3002.6±1.4**			12
49[r]	97KYT-60	Steep Rock Mine	599320	5407880	ultramafic lapilli tuff	Steep Rock Lake	Dismal			2997, 2995, 2780		1

Table 2. continued.

No. [1]	Sample Number	Area	UTM Easting (m)	Location Northing (m)	Rock Type	Greenstone Belt or Domain	Assemblage or Plutonic Suite	Age of Crystallization	Detrital Grain Age, Max. Depositional Age (in bold)	Inherited Age	Metamorphic Age	Ref. (*see* end of table)
50	DD07-5	SE arm, Steep Rock Lake	604023	5406559	sandstone (Wagita Formation)	Steep Rock Lake	Steep Rock		2984±11, **2779±22**			20
51	DD07-7	SE arm, Steep Rock Lake	603490	5406575	ultramafic lapilli tuff	Steep Rock Lake	Dismal			2995±11, 2782±15		20
52	DD07-8	Steep Rock Lake	598554	5409164	wacke (Wagita Formation)	Steep Rock Lake	Steep Rock		2997±4 (84 grains by LA-ICP–MS), **2939±20**			21
Whistle and Mine Centre Domains, Puffy Pillow and Witch Bay Assemblages, Winnipeg River Terrane												
60[t]	97KYT-75	E of Whistle Lake	763285	5469530	felsic tuff	Heaven Lake	Whistle	2729±2		2813		1
61[t]	DD95-4	E of Whistle Lake	762323	5467740	quartz + feldspar porphyry intrusion	Heaven Lake	Whistle	2727.8±1.6		2758 to 2779		1
62[t]	TH01-149	E of Legris Lake	760980	5454350	dacite tuff	Lac des Iles	unknown assemblage	2727.8±1.3				1
63	DD86-25A	Lac des Iles	745093	5452405	hornblende tonalite	Whistle domain	Hornblende Tonalite	2727.8±1.5				11
64	05DS87	Lac des Mille Lacs	692750	5409820	dacite breccia	Lac des Mille Lacs	Puffy Pillow	2725.0±1.1				4
65	05DS154	Atikokan airport	600330	5403550	leucogabbro	Steep Rock Lake	Witch Bay	2735.2±0.8				4
66	05DS133	W of Steep Rock Lake	594058	5402278	quartz porphyritic tonalite	Steep Rock Lake	dike cutting Witch Bay	2729.7±1.9		2865		4
69[r]	11	Calm Lake	569210	5399360	felsic volcanic	Mine Centre	unknown assemblage	2722.3+1.3/–1.1				13
70	10	E of Mine Centre	533350	5399380	quartz-phyric volcanic rock	Mine Centre	unknown assemblage	2727.0±1.2				13
71	9	NW of Bad Vermillion Lake	520420	5398310	tonalite	Mine Centre	Biotite Tonalite	2728.1+1.8/–1.3				13
72	8	NW of Bad Vermillion Lake	519320	5399550	quartz porphyritic rhyolite	Mine Centre	unknown	2727.8+4/–2.6				13
73	03DS94	S of Upsala	687712	5432346	biotite tonalite	Whistle domain	Biotite Tonalite	2716.8±1.0			2707±37 (titanite)	7
74	06DS76	N of Upsala	680430	5445957	biotite tonalite	Whistle domain	Biotite Tonalite	2712±6[s]				19
75	00KYT06	S Garden Lake	714296	5484615	rhyolite	Garden Lake (Winnipeg River)	unknown assemblage	2725.5±1.3				1
76	00KYT34	NE Wawang Lake	687280	5482630	tonalite gneiss	Winnipeg River terrane?	Tonalite Gneiss	2712±2				10
77	06DS75	E of Upsala	696700	5431600	biotite tonalite	Whistle domain/ Savoy domain?	Biotite Tonalite	2725±6[s]		2811±24[s]	2721±6[s], 2678±9[s]	19
78	09DS251	Finlayson Lake	606110	5421654	wacke	Finlayson	Witch Bay		61 grains by LA-ICP–MS ranging from **2678** to 3031			23

Table 2. continued.

No.[1]	Sample Number	Area	UTM Easting (m)	Location Northing (m)	Rock Type	Greenstone Belt or Domain	Assemblage or Plutonic Suite	Age of Crystallization	Detrital Grain Age, Max. Depositional Age (in bold)	Inherited Age	Metamorphic Age	Ref. (*see* end of table)
Western Wabigoon Terrane												
80	00KYT19	NW of Eltrut Lake	537761	5436468	tonalite gneiss	western Wabigoon terrane	Tonalite Gneiss	2732				10
81	00KYT23	N of Irene Lake	605622	5454407	hornblende tonalite	late intrusion in Marmion domain	Hornblende Tonalite	2721.3±1				6
82	06DS10	W of Raleigh Lake	574950	5471340	intermediate volcanic breccia	Raleigh Lake	unknown assemblage	2730±7[s]				22 (this study)
83	06DS22	W of Entwine Lake	518247	5448150	biotite tonalite	western Wabigoon terrane	Biotite Tonalite	2716±6[s]				22 (this study)
84	07DS42b	W of Clearwater West Lake	567573	5422363	mafic tonalite	western Wabigoon terrane	Biotite Tonalite	2697.4±0.8				20
85	DD07-3	W of Pekagoning Lake	552086	5444622	biotite tonalite	western Wabigoon terrane	Biotite Tonalite	2709±9			2703±5	20
86	07DS44	N of Eltrut Lake	546722	5435910	biotite tonalite	western Wabigoon terrane with Winnipeg River inheritance	Biotite Tonalite	2688±5		2740, 3293±30		20
87	07DS37	E of Ignace	610464	5466497	biotite tonalite	western Wabigoon terrane with inherited central Wabigoon?	Biotite Tonalite	2719		~2923		21
Quetico and Seine Sequences												
90[t]		E of Legris Lake	765886	5452545	conglomerate	Lac des Iles	Seine assemblage		**2696**			14
91	DD86-11	S of Atikokan	604211	5397904	wacke	Quetico Subprovince	Quetico assemblage		3009, 3009, 3004, 3004, 2701, **2698**			15
92	3AZ	S of Atikokan	602950	5398943	wacke	Quetico Subprovince	Quetico assemblage		**2707**			15
93	8BZ	S of Atikokan	602610	5397940	wacke	Quetico Subprovince	Quetico assemblage		**2704**			15
94	11AZ	S of Atikokan	602064	5398020	wacke	Quetico Subprovince	Quetico assemblage		**2739**			15
95	DD86-10	S of Atikokan	600812	5398020	wacke	Quetico Subprovince	Quetico assemblage		2901, **2701**			15
96	15CZ	S of Atikokan	599510	5398130	wacke	Quetico Subprovince	Quetico assemblage		**2796**			15
97	27BZ	SW of Atikokan	594866	5399500	wacke	Quetico Subprovince	Quetico assemblage		2768, **2706**			15
98	33AZ	SW of Atikokan	592060	5398600	wacke	Quetico Subprovince	Quetico assemblage		2719, **2709**			15

Table 2. continued.

No. [1]	Sample Number	Area	UTM Easting (m)	Location Northing (m)	Rock Type	Greenstone Belt or Domain	Assemblage or Plutonic Suite	Age of Crystallization	Detrital Grain Age, Max. Depositional Age (in bold)	Inherited Age	Metamorphic Age	Ref. (*see* end of table)
99	35BZ	SW of Atikokan	588187	5398384	wacke	Quetico Subprovince	Quetico assemblage		**2792**			15
100	45CZ	S of Calm Lake	573300	5397680	wacke	Quetico Subprovince	Quetico assemblage		2709, **2704**			15
101	101DZ	S of Calm Lake	567275	5397826	wacke	Quetico Subprovince	Quetico assemblage		2913, 2787, **2704**			15
102	101EZ	S of Calm Lake	566962	5397558	wacke	Quetico Subprovince	Quetico assemblage		2724, 2705, **2700**			15
103	103BZ	S of Calm Lake	565867	5398206	wacke	Quetico Subprovince	Quetico assemblage		**2788**			15
104	12	E of Mine Centre	538409	5399392	tonalite clast in conglomerate	Mine Centre	Seine assemblage		**2696.1 +5/–3**			13
105	6 and 7	S of Rainy Lake	504500	5395650	wacke	Mine Centre	Coutchiching/Quetico assemblage		3059, 3057, 3056, 2940, 2830, 2726, 2721, 2714, 2712, 2704, **2704**			13
106	00KYT22	N of Irene Lake	604990	5455570	conglomerate	small outlier in Marmion domain	Seine assemblage		2718, 2710, 2707, **2700**			6
107	Sample 5	Bear Pass	501770	5395540	felsic sill in Couchiching sediments	Mine Centre	Coutchiching/Quetico assemblage			2717, 2825		13
Late Plutonic Rocks												
Biotite Granite Suite												
110	05DS48	Savanne	707560	5423735	potassium feldspar megacrystic granite	Muskeg batholith	Biotite Granite	2686±2		2690		4
111	WOB86	White Otter Lake	587900	5432889	coarse-grained biotite granite	White Otter batholith	Biotite Granite	2685±2			2667 (titanite)	16
112[t]	00KYT28	N of Ignace	613745	5488309	biotite granite	Indian (Cecil) Lake batholith	Biotite Granite	2671				10
113	06DS50	N of Joe Lake	568130	5413120	biotite granodiorite	Joe pluton	Biotite Granite	2686±6[s]				22 (this study)
Sanukitoid Suite												
115[t]	00KYT09	Roaring River			monzonite	Roaring River complex	Sanukitoid	2697				10
116	03DS91	S of Lac des Iles	747519	5447943	hornblende diorite	Shelby Lake batholith	Sanukitoid	2690±0.9			2674, 2667 (titanite)	8
117	ON+200	Sapawe	622784	5402685	quartz monzonite	Blalock pluton	Sanukitoid	2688±4				15
118		N of Atikokan	595442	5417355	biotite-hornblende granite	Eye–Dashwa pluton	Sanukitoid	2684±25**			2665 (titanite)	17

Table 2. continued.

No. [1]	Sample Number	Area	UTM Easting (m)	Location Northing (m)	Rock Type	Greenstone Belt or Domain	Assemblage or Plutonic Suite	Age of Crystallization	Detrital Grain Age, Max. Depositional Age (in bold)	Inherited Age	Metamorphic Age	Ref. (*see* end of table)
119	13	Rainy Lake	509600	5397240	quartz monzodiorite	Ottertail pluton	Sanukitoid	2686.1+1.5/–1.4				13
120	06DS23	Entwine Lake	532330	5443920	monzodiorite	Entwine stock	Sanukitoid	2685±8[s]				22 (this study)
Tonalite Gneiss Suite												
125	04DS80	Raith	726017	5411947	tonalite to granodiorite gneiss	Raith gneisses	Tonalite Gneiss	2673.0±1.1			2657 (titanite)	7
Lac des Iles Suite												
130	DD86-22	Tib Lake	733478	5458252	gabbro pegmatite	Tib stock	Lac des Iles	2685.9±1.6				11
131	Murphy 1	Lac des Iles	747853	5452420	gabbro pegmatite	Mine Block intrusion	Lac des Iles	2693.3±1.3				18
132	DD85-2	Lac des Iles	746668	5451861	gabbro pegmatite	Mine Block intrusion	Lac des Iles	2689.0±1.0				11
133[t]	JP01-04		346305***	5504619	gabbro	Johspine gabbro	Lac des Iles?	2699.0±0.9				18
134[t]	GK01-08		335900***	5485503	gabbro	Geike gabbro	Lac des Iles	2687.6±1.2				18
135[t]	99TRH-474/82	Garden Lake	315182***	5492528	gabbro	Graden Lake gabbro	Lac des Iles	2686.1±2.2				18
136[t]	00TRH-092		331113***	5453513	gabbro	Lac des Iles east gabbro	Lac des Iles	2671±11				18

Notes: UTM co-ordinates are NAD83, Zone 15 (except ***, *see* below);
[1] refers to numbers on Figure 3 (*see* Chart A, back pocket) and Preliminary Map P.2229 (*see* back pocket);
* the dated zircon grains are rounded and probably are derived from the sedimentary component of the rock;
** U/Pb whole rock age;
[s] analyzed by sensitive high-resolution ion microprobe (SHRIMP);
[r] location approximate;
[t] located outside the area of this study, but included for comparative purposes;
*** NAD83, Zone 16, mainly east of the study area.

Abbreviation: LA-ICP–MS, laser ablation inductively coupled plasma mass spectrometry

References: 1) Tomlinson et al. (2003); 2) Fralick and King (1995); 3) P. Fralick (Lakehead University, unpublished data, 2001; personal communication, 2007); 4) Davis (2006); 5) Davis and Jackson (1988); 6) Tomlinson et al. (2001); 7) Davis (2005); 8) Kamo (2004); 9) D.W. Davis (Jack Satterly Geochronology Laboratory, unpublished data (reported *in* Stone, Kamineni and Jackson 1992)); 10) K.Y. Tomlinson (unpublished data (reported *in* Tomlinson et al. 2004)); 11) Davis (2003); 12) D.W. Davis (Jack Satterly Geochronology Laboratory, unpublished data, 1986 (reported *in* Stone, Tomlinson et al. 2002)); 13) Davis, Poulsen and Kamo (1989); 14) D.W. Davis (Jack Satterly Geochronology Laboratory, unpublished data, 1994; personal communication, 2007); 15) Davis, Pezzutto and Ojakangas (1990); 16) Davis (1993); 17) Zartman and Kwak (1990); 18) Heaman and Easton (2006); 19) Hamilton, Davis and Kamo (2007); 20) Davis (2008); 21) Davis (2009); 22) this study; 23) D.W. Davis (Jack Satterly Geochronology Laboratory, written communication, 2010).

Table 3. U/Pb sensitive high-resolution ion microprobe (SHRIMP) analyses on zircon from the central Wabigoon Subprovince area. Samples analyzed at the Beijing SHRIMP Center (Institute of Geology, Chinese Academy of Geological Sciences).

Spot Name	$^{206}Pb_c$ %	U ppm	Th ppm	$^{232}Th/^{238}U$	$^{206}Pb^*$ ppm	$^{204}Pb/^{206}Pb$	% error	$^{206}Pb/^{238}U$ Age (1)	$^{207}Pb/^{206}Pb$ Age (1)	$^{208}Pb/^{232}Th$ Age (1)	% Disc	$^{207}Pb^*/^{235}U$ (1)	± %	$^{206}Pb^*/^{238}U$	± %	Error Corr.
A. 06DS10	**volcanic breccia**															
06DS10-*9*	*0.67*	*106*	*64*	*0.62*	*41.6*	*0.000487*	*17*	*2403 ±20*	*2691 ±16*	*957 ±47*	*11*	*11.47*	*1.4*	*0.4517*	*1*	*0.729*
06DS10-10	0.12	164	50	0.32	73.5	0.000088	27	2708 ±18	2732 ±9.1	2550 ±63	1	13.59	0.98	0.5222	0.81	0.827
06DS10-12	0.18	176	72	0.42	80.3	0.000131	21	2737 ±22	2727 ±9.0	2598 ±41	0	13.72	1.1	0.5288	0.99	0.874
06DS10b-1	0.36	70	19	0.28	32.3	0.000268	24	2761 ±26	2712 ±16	2533 ±100	-2	13.75	1.5	0.5347	1.2	0.769
06DS10b-2	0.09	216	70	0.34	100	0.000066	30	2785 ±15	2734 ±8.3	2684 ±36	-2	14.09	0.84	0.5405	0.67	0.801
06DS10b-3	0.09	141	116	0.85	65.5	0.000063	50	2787 ±23	2733 ±16	2777 ±40	-2	14.09	1.4	0.5408	1	0.732
06DS10b-4	0.44	122	44	0.37	54.7	0.000327	16	2695 ±19	2735 ±12	2377 ±77	1	13.54	1.1	0.5191	0.86	0.765
06DS10b-5	0.14	155	66	0.44	70.2	0.000099	27	2724 ±18	2722 ±10.0	2353 ±40	0	13.6	10	0.5258	0.79	0.794
06DS10-9b	0.17	301	138	0.47	138	0.000129	16	2759 ±12	2753 ±7.5	2660 ±31	0	14.081	0.69	0.5341	0.52	0.749
06DS10-9c	0.16	132	70	0.54	64.6	0.000120	25	2898 ±17	2727 ±9.4	2540 ±35	-6	14.73	0.94	0.5677	0.74	0.795
06DS10-11	0.23	112	38	0.35	51.9	0.000169	22	2778 ±30	2742 ±10	2626 ±59	-1	14.11	1.5	0.5386	1.3	0.908
B. 06DS22	**biotite tonalite**															
06DS22b-14	0.12	197	116	0.61	87.2	0.000085	24	2678 ±17	2712 ±8.2	2597 ±34	1	13.25	0.91	0.515	0.76	0.837
06DS22b-15	0.14	275	176	0.66	120	0.000102	22	2646 ±13	2693 ±7.3	2617 ±24	2	12.902	0.73	0.5074	0.59	0.798
06DS22b-16	0.22	507	172	0.35	231	0.000164	11	2738 ±10	2717 ±5.2	2940 ±30	-1	13.653	0.56	0.5291	0.46	0.825
06DS22b-17	0.08	490	286	0.60	210	0.000059	20	2610 ±9.2	2698 ±5.2	2652 ±17	3	12.733	0.53	0.4992	0.43	0.807
06DS22b-18	0.23	133	46	0.36	56.5	0.000170	22	2582 ±26	2704 ±10	2524 ±61	5	12.6	1.4	0.4925	1.2	0.894
06DS22-19	0.12	313	212	0.70	142	0.000090	19	2732 ±12	2716 ±6.0	2687 ±21	-1	13.608	0.66	0.5277	0.55	0.835
06DS22-21	*26.41*	*1*	*0*	*0.01*	*0.386*	*0.021424*	*20*	*2256 ±230*	*3271 ±430*		*31*	*15.3*	*30*	*0.419*	*12*	*0.401*
06DS22-23	0.11	209	132	0.65	92.1	0.000083	21	2664 ±13	2724 ±7.3	2599 ±23	2	13.263	0.75	0.5118	0.6	0.806
C. 06DS23	**monzodiorite**															
06DS23-26	0.17	110	110	1.03	49.5	0.000122	25	2708 ±17	2675 ±11	2604 ±28	-1	13.14	1	0.5222	0.77	0.756
06DS23-24b	0.04	105	72	0.71	46.6	0.000031	77	2685 ±26	2695 ±10	2673 ±39	0	13.15	1.3	0.5166	1.2	0.880
06DS23-25	0.09	191	231	1.25	83.8	0.000065	33	2656 ±13	2692 ±7.7	2610 ±21	1	12.96	0.77	0.5099	0.62	0.798
06DS23-28	0.17	124	146	1.21	56.2	0.000123	32	2719 ±43	2687 ±11	2676 ±56	-1	13.29	2.1	0.525	2	0.950
06DS23-30	0.17	278	278	1.04	120	0.000121	17	2631 ±11	2678 ±6.6	2581 ±18	2	12.694	0.65	0.5039	0.52	0.794
D. 06DS50	**biotite granodiorite**															
06DS50b-31	0.06	728	218	0.31	322	0.000048	29	2675 ±7.6	2691 ±4.3	2595 ±21	1	13.064	0.44	0.5144	0.35	0.800
06DS50b-36b	0.26	332	100	0.31	147	0.000190	17	2670 ±19	2690 ±8.6	2780 ±51	1	13.02	1	0.5132	0.87	0.858
06DS50b-33	0.11	256	20	0.08	119	0.000078	30	2785 ±16	2674 ±9.0	2885 ±120	-4	13.58	0.89	0.5404	0.7	0.792
06DS50b-34	0.40	944	336	0.37	382	0.000290	8	2477 ±8.5	2607 ±5.7	2405 ±39	5	11.31	0.54	0.4685	0.42	0.770
06DS50b-37	0.03	393	9	0.02	185	0.000021	55	2815 ±20	2682 ±6.7	2844 ±200	-5	13.83	0.97	0.5474	0.88	0.909
06DS50-40	0.44	121	72	0.62	48.9	0.000322	21	2475 ±22	2728 ±15	2684 ±58	9	12.16	1.4	0.4682	1.1	0.754
06DS50-41	0.09	231	4	0.02	109	0.000069	28	2811 ±14	2678 ±7.4	2485 ±410	-5	13.77	0.75	0.5467	0.61	0.806

Notes: (1) Common Pb corrected using measured ^{204}Pb. Errors are 1σ; Pb_c and Pb^* indicate the common and radiogenic portions, respectively; error in Standard calibration was 0.35% (not included in above errors, but required to compare data from different mounts). Analyses in italics are excluded from the mean age.

Table 3. continued.

Spot Name	$^{206}Pb_c$ %	U ppm	Th ppm	$^{232}Th/^{238}U$	$^{206}Pb^*$ ppm	$^{204}Pb/^{206}Pb$	% error	$^{206}Pb/^{238}U$ Age [(1)]		$^{207}Pb/^{206}Pb$ Age [(1)]		$^{208}Pb/^{232}Th$ Age [(1)]		% Disc	$^{207}Pb^*/^{235}U$ [(1)]	± %	$^{206}Pb^*/^{238}U$	± %	Error Corr.
E. 06DS52	**tonalite gneiss**																		
06DS52b-45	0.16	80	38	0.49	39	0.000119	35	2882	±32	2877	±18	2923	±63	0	16.03	1.8	0.5636	1.4	0.787
06DS52b-46	0.19	97	54	0.57	46.4	0.000142	24	2835	±19	2859	±9.8	2728	±39	1	15.55	1	0.5525	0.84	0.814
06DS52b-47	0.31	137	59	0.45	67	0.000237	16	2889	±17	2866	±9.0	2810	±45	-1	15.98	0.91	0.5655	0.72	0.791
06DS52b-48	0.09	129	96	0.77	61.7	0.000068	35	2850	±17	2864	±9.7	2857	±41	1	15.69	0.96	0.5559	0.75	0.785
06DS52b-49	0.21	56	101	1.88	25	0.000155	39	2709	±26	2686	±15	2727	±39	-1	13.23	1.5	0.5224	1.2	0.781
06DS52b-49b	0.07	141	61	0.45	67.2	0.000049	35	2841	±17	2825	±8.4	2899	±34	-1	15.25	0.9	0.5537	0.74	0.821
06DS52-50	0.62	71	7	0.11	31.5	0.000457	16	2684	±22	2750	±15	2777	±270	2	13.59	1.3	0.5164	10	0.745
06DS52-51	0.14	167	81	0.50	80.6	0.000104	20	2870	±14	2869	±7.1	2848	±30	0	15.88	0.75	0.5609	0.61	0.812
06DS52-55	*0.41*	*251*	*135*	*0.56*	*112*	*0.000307*	*11*	*2693*	*±15*	*2830*	*±15*	*2673*	*±47*	*5*	*14.33*	*1.2*	*0.5186*	*0.69*	*0.593*
06DS52-56	*0.33*	*56*	*20*	*0.36*	*26.5*	*0.000246*	*25*	*2811*	*±25*	*2781*	*±14*	*2701*	*±77*	*-1*	*14.66*	*1.4*	*0.5465*	*1.1*	*0.785*

Notes: [(1)] Common Pb corrected using measured ^{204}Pb. Errors are 1σ; Pb_c and Pb^* indicate the common and radiogenic portions, respectively; error in Standard calibration was 0.35% (not included in above errors, but required to compare data from different mounts).
Analyses in italics are excluded from the mean age.

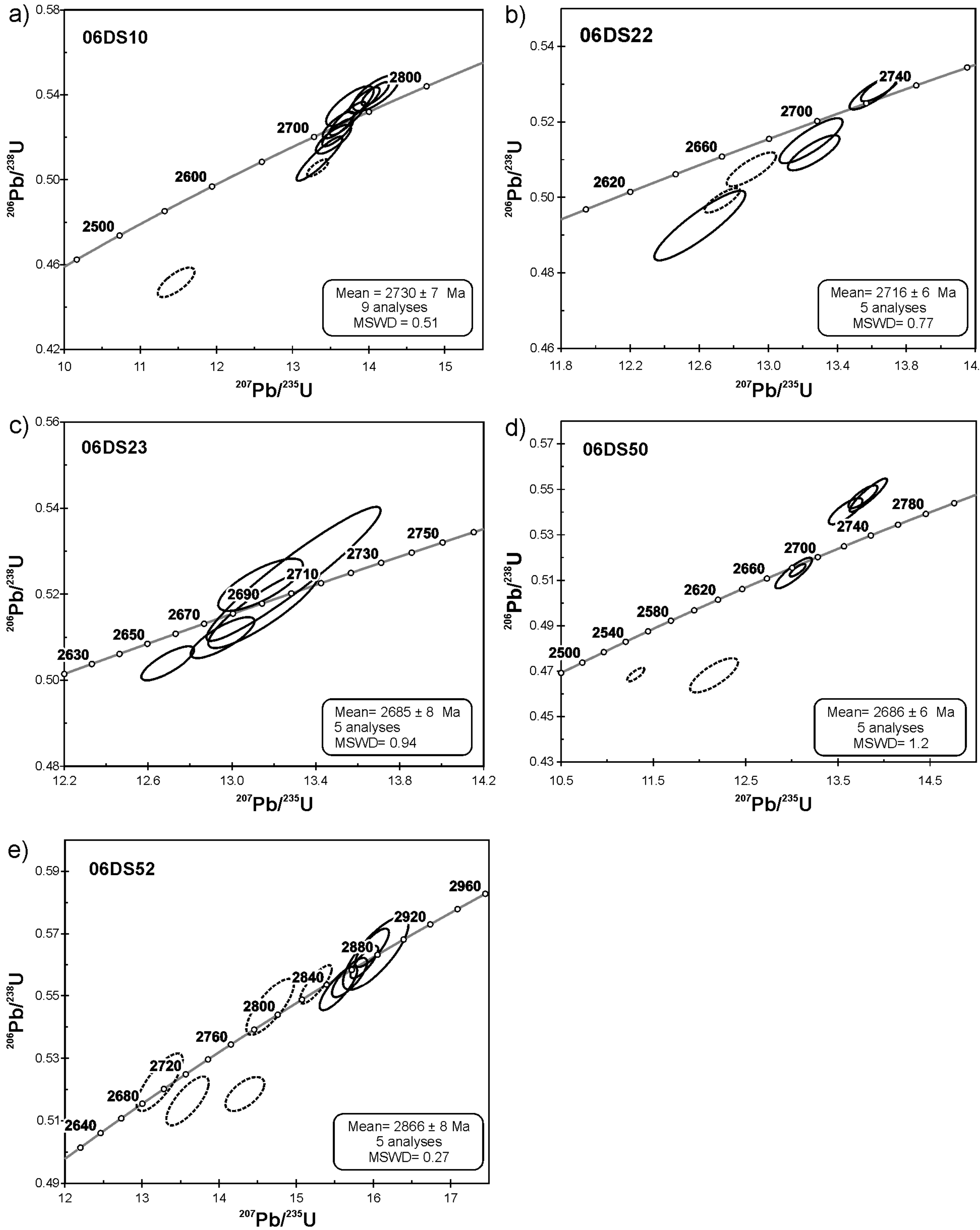

Figure 4. Concordia diagrams for geochronology samples analyzed at the Beijing SHRIMP Center (Institute of Geology, Chinese Academy of Geological Sciences). Analyses represented by dashed ellipses are excluded from the mean age.

Rock Samples, Geochemical Analyses and Mineral Compositions

Approximately 1280 rock samples were collected from the central Wabigoon Subprovince area. Many rock samples were assayed for economic metals; these assay results are listed with the sample file. Of the samples, approximately 388 were analyzed for major and trace elements and are used to characterize the chemistry of greenstone sequences and plutonic rock suites.

The results of the various analyses that support this study are available separately as Miscellaneous Release—Data (MRD) 242 (Stone 2010b). Although data in these 18 Microsoft® Excel® (*.xls*) spreadsheets are mentioned, referred to and/or discussed in the report, they are only provided in MRD 242 (*see* "Contents"). However, to simplify references in the report, the spreadsheets have been labelled as "Table xx" and are cited in the report as "*see* MRD 242: Table xx". The files comprise sample information and locations (Table 4), geochemical analyses (Tables 8 and 9), electron microprobe data (Tables 11 to 22: mineral compositions) and geochemistry on soil samples (Tables 27 to 29).

Polished thin sections were made from many samples for petrographic analysis and for determination of mineral compositions by microprobe. Mineral compositions are grouped by mineral name (e.g., biotite, chlorite, pyroxene; *see* MRD 242: Tables 22 to 29). Amphiboles and feldspars were analyzed extensively for aluminum-in-hornblende barometry and thermometry and are discussed in "Aluminum-in-Hornblende Barometry" (*see* MRD 242: Tables 11 to 14 for supporting data).

Approximately 389 samples of soil (mainly unweathered till) were dug from shallow pits in the Bolton Bay of Lac des Mille Lacs, Sapawe and Harold Lake areas. The soil samples (*see* MRD 242: Tables 27 to 28) were assayed for various metals and are discussed further in "Economic Geology".

Representative pieces of the rock samples, powders of the samples used for chemical analyses and the polished thin sections are stored in the archives of the Ontario Geological Survey in Sudbury.

General Geology: The Central Wabigoon Subprovince Area

The central Wabigoon Subprovince area is underlain by bifurcated and anastomosed greenstone belts separated by large, commonly oval masses of felsic plutonic rocks. The greenstone belts (shown on Figure 5 (Chart A, back pocket) and on the 1:250 000 scale Map P.2229 (back pocket)) are composed primarily of mafic metavolcanic rocks with lesser parts of intermediate to felsic metavolcanic rocks, gabbro and metasedimentary rocks. The Mine Centre and Lac des Mille Lacs greenstone belts have large components of gabbro (*see* Figure 5). The supracrustal strata of greenstone belts typically show a subvertical foliation and are metamorphosed from greenschist to amphibolite facies.

Well-bedded wacke–siltstone sequences (Quetico sedimentary sequences) of the Quetico Subprovince extend along the south margin of the central Wabigoon Subprovince area. Southward, the metasedimentary rocks undergo increasing metamorphic grade giving way to migmatites intruded by peraluminous granite plutons. Coarse conglomerate of the Seine sequence occurs in the Shoal Lake area and as small, well-foliated units unconformably overlying Quetico sediments along the Quetico–Wabigoon subprovinces boundary.

Seven major suites of felsic to intermediate and mafic plutonic rocks occur in the central Wabigoon Subprovince area. Among the oldest plutonic rocks are intrusions of the biotite tonalite and gneissic suites, examples of which include the Marmion batholith and Eltrut, Dashwa and Raven gneiss complexes (*see* Figure 5).

Coarse-grained hornblende tonalite to granodiorite of the hornblende tonalite suite constitutes a few percent of the central Wabigoon Subprovince area and occurs as oval to elongate masses at scattered localities. Muscovite±biotite granite of the peraluminous suite occurs as dikes, masses and plutons intruding metasedimentary rocks of the Quetico Subprovince and in greenstone belts.

Late, massive and mainly posttectonic intrusive rocks include the biotite granite, sanukitoid and Lac des Iles suites. Pink biotite granite to granodiorite of the biotite granite suite is among the most common rock types in the area occurring in forms ranging from dikes and stocks to large batholiths. Examples of large biotite granite bodies include the White Otter, Cecil and Muskeg batholiths. Oval plutons of the sanukitoid suite occupy 5% of the central Wabigoon Subprovince area. The sanukitoid plutons are heterogeneous and are variably composed of diorite, monzodiorite, monzonite and granite with gabbroic to hornblendite inclusions. Gabbroic to pyroxenitic and peridotitic dikes and stocks of the Lac des Iles suite, many of which are too small to be shown in Figure 5, occur within a broad band extending from south of Atikokan to Lac des Iles. Larger examples of this suite include the Tib and Lac des Iles stocks at Lac des Iles. The Mine Block intrusion, which adjoins the southern Lac des Iles stock has been recently mined for palladium, platinum and gold.

Two major sets of Neoarchean subvertical, transcurrent faults transect the area. The dextral Quetico and Shelby Lake faults extend easterly along the Quetico–Wabigoon subprovinces boundary and are marked by zones of schist and mylonite up to 1 km in width. Northeast- to north-northeast-trending faults include the Entwine, Marmion, Pakashkan and Gull River faults. At its southern end within the Marmion batholith, the Marmion fault broadens to a complex braided zone of fault segments, some of which are mineralized with gold.

Archean rocks of the central Wabigoon Subprovince area are cut by Proterozoic diabase dikes and sills. The larger masses of Proterozoic diabase are associated with the Nipigon sill complex and occur in the eastern part of the area.

SUPRACRUSTAL BELTS

The geology, geochemistry and geochronology of the major supracrustal belts are described below.

Lumby Lake Greenstone Belt

The Lumby Lake greenstone belt (“Lumby belt”) tapers easterly over a distance of 60 km in the centre of the central Wabigoon Subprovince area (*see* Figure 5; *see* Map P.2229) and wraps around oval plutons of biotite granite and monzonite. Narrow arms of the Lumby belt extend southwesterly to the Finlayson Lake greenstone belt and northeasterly to the Graham greenstone belt. The Lumby Lake greenstone belt was mapped by Jackson (1985a, 1985b) and subsequently re-interpreted by Tomlinson et al. (2003) who added geochemical and geochronologic data. The western half of the greenstone belt as later mapped by Buse, Lewis and Magnus (2009). The Lumby South assemblage extends along the south margin of the belt in contact with gabbro, hornblende tonalite of the Diversion stock and biotite tonalite of the Marmion batholith. The Lumby South assemblage youngs north and contains basal felsic metavolcanic sequences overlain by mafic metavolcanic flows with rare komatiite. Three felsic units in this assemblage have ages of 3001 to 2999 Ma (*see* Table 2, Nos. 6, 7 and 8) on the basis of which it is included with the Marmion domain (*see* Figure 3). South of Pyramid Lake on the north side of the Lumby belt, a felsic unit in mafic metavolcanic sequences has an age of 3013 Ma (*see* Table 2, No. 9). This locality is interpreted to represent a small outlier of the Lumby South assemblage within the Lumby North assemblage in view of the marked difference in age between the 3013 Ma felsic unit and a nearby 2963 Ma felsic unit (*see* Table 2, No. 142) representative of the Lumby North assemblage.

The Lumby North assemblage extends along the northern side of the Lumby Lake greenstone belt and is intruded by monzodiorite, hornblende granodiorite and biotite granite plutons. Near Upsala, mafic sequences of the Lumby North assemblage are interlayered with iron formation and quartz arenite adjacent to the unexposed contact with the Marmion batholith. This locality possibly represents a thin platform sequence developed on the Marmion batholith. The Lumby North assemblage contains a high proportion of massive mafic flows or gabbro interspersed with thin felsic metavolcanic units and rare pillowed mafic units. The mafic units are metamorphosed to black amphibolite and the assemblage generally youngs north and is characterized by a relatively low residual magnetic field in comparison with that of the Lumby South assemblage (Ontario Geological Survey 2009a). A thin felsic unit in the Lumby North assemblage has an age of 2963 Ma on the basis of which this assemblage is included with the Whitton domain.

The Pinecone assemblage occupies the central axis of the western part of the Lumby Lake greenstone belt (*see* Figure 3). The Pinecone assemblage, represented by rock units 5 and 7 on Map P.2229, is dominated by south-younging pillowed mafic flows with komatiite-iron formation units and is capped by a 500 m thick unit of clastic metasedimentary rocks and iron formation (Fralick and King 1995). A thin felsic unit associated with a komatiite flow has an age of 2828 Ma (*see* Table 2, No. 150).

The Bar assemblage is a package of mafic flows with thin units of felsic tuff and wacke at the south side of the Pinecone assemblage in the western Lumby Lake greenstone belt (*see* Figure 3). The southern boundary of the Bar assemblage, representing the contact with the Lumby South assemblage is not well defined, but is possibly marked by a discontinuous metasedimentary unit. A thin felsic metavolcanic tuff at Bar Lake has an age of 2897 Ma (*see* Table 2, No. 35).

Tomlinson et al. (1999) studied the geochemistry of the western Lumby Lake greenstone belt and documented a wide variety of rock compositions ranging from basaltic komatiite through tholeiitic and calc-alkalic basalt, andesite, dacite and rhyolite. On the basis of primitive mantle-normalized trace element profiles, the basaltic komatiites were subdivided into enriched and unfractionated varieties. The enriched basaltic komatiites are enriched in LREE as well as Th, Nb and Ta (Th < Nb < Ta) and are depleted in HREE. The unfractionated basaltic komatiites have generally flat trace element profiles at 2× to 5× primitive mantle values with slight depletion in Th, Nb and Eu. Five types of basalts were recognized ranging from those with relatively flat profiles and Th < Nb < Ta (depleted) through slightly and moderately LREE-enriched varieties to strongly LREE-enriched varieties. The multi-element profiles of the more LREE-enriched basalts have Nb and Ta troughs and are somewhat variably depleted in Zr, Hf and Ti. Tomlinson et al. (1999) noted that although all of the lava types occur widely in the Lumby belt, the Lumby North assemblage is composed of mainly unfractionated basaltic komatiite and depleted basalt.

Eight samples of metavolcanic rock and associated gabbro were collected from the eastern Lumby Lake greenstone belt and chemically analyzed to compliment the analyses of Tomlinson et al. (1999) from the western part of the belt. The samples ranged from komatiite through basalt and andesite to rhyolite (Figure 6a). The komatiite from the Lumby South assemblage (sample 02DS54 of Figure 7a) has primitive mantle-normalized trace element profiles similar to the enriched basaltic komatiite of Tomlinson et al. (1999). The basalts collected from the Lumby South, Pinecone and Lumby North assemblages have generally flat trace-element profiles with Th < Nb < Ta and are comparable to the depleted basalts of Tomlinson et al. (1999). One sample of a thin rhyolite tuff unit within pillowed basalts of the Lumby North assemblage is strongly LREE-enriched with Nb and Ti troughs (*see* Figure 7a: sample 02DS41).

Graham Greenstone Belt

The curved and bifurcated Graham greenstone belt ("Graham belt") attains a width of up to a few kilometres and extends over a distance of 30 km in the northern central Wabigoon Subprovince area. The Graham belt is composed of mainly mafic metavolcanic rocks metamorphosed to black amphibole gneiss and is intruded by a small oval pluton of monzodiorite (*see* Figure 5). A few outcrops of dull, dense, brown weathered komatiite are identified within the mafic metavolcanic sequences at the northwest end of the belt. Metavolcanic rocks of the Graham belt unconformably overlie biotite tonalite at the northwest end of the belt. The unconformity is marked by a unit of poorly bedded quartzo-feldspathic sandstone that locally attains a thickness of several tens of metres along the banks of the English River.

The biotite tonalite beneath the unconformity has an age of 2814 Ma (*see* Table 2, No. 151), which implies that metavolcanic sequences overlying the unconformity along the west to northwest side of the Graham belt are younger than 2814 Ma. In contrast, biotite tonalite north and east of the Graham belt has an age of 2929 Ma (*see* Table 2, No. 24) and although the contact between the 2929 Ma biotite tonalite and the east to northeast side of the belt is not well exposed, the high metamorphism of mafic metavolcanic flows near the contact suggests that the Graham belt is intruded by the 2929 Ma biotite tonalite pluton. These data can be reconciled if the Graham belt is composed of 2 assemblages including an assemblage older than 2929 Ma along the east to northeast side and an assemblage younger than 2814 Ma along the west to northwest side. Possibly, these strata represent extensions of the Pinecone and Lumby North assemblages of the Pinecone–Savoy and Whitton domains, respectively (*see* Figure 3).

Four samples of metavolcanic material from the Graham belt were chemically analyzed and range compositionally from komatiite through basaltic komatiite to basalt (Figure 6b). All samples have relatively flat mantle-normalized trace element profiles (Figure 7b). The multi-element profiles of the komatiite (sample 00DS33), basaltic komatiite and basalts are chemically comparable to the unfractionated basaltic komatiite and depleted basalts recorded by Tomlinson et al. (1999) for the Lumby Lake greenstone belt.

Phyllis Lake Greenstone Belt

The Phyllis Lake greenstone belt ("Phyllis belt") attains a width of a few kilometres and extends northeasterly over a distance of about 30 km in the northern central Wabigoon Subprovince area (*see* Figure 5). The Phyllis belt is composed of mafic metavolcanic rocks that show pillows in less deformed areas and widespread amphibolite-facies metamorphism. The metamorphism has transformed the metavolcanic rocks to amphibole gneisses at many localities in the belt. Mafic metavolcanic rocks of the Phyllis belt unconformably overlie biotite tonalite along the northwest side of the belt. The unconformity is marked by a garnetiferous quartzo-feldspathic sandstone unit that attains a thickness of up to a few tens of metres.

A thin felsic tuff within mafic metavolcanic flows in the centre of the Phyllis belt has an age of 2955 Ma (*see* Table 2, No. 143). Tonalite gneisses of the Raven gneiss complex on the northwest side of the Phyllis belt is dated at 2989 Ma (*see* Table 2, No. 16) and probably represent a basement complex on which lavas of the Phyllis belt were deposited. In contrast, biotite tonalite on the southeast side of the Phyllis belt has a U/Pb zircon age of 2817 Ma (*see* Table 2, No. 152) and represents part of the Pinecone–Savoy domain. The Phyllis belt is included with the Whitton domain.

Six samples of mafic metavolcanic rocks from the Phyllis belt and a small greenstone sliver south of the belt were determined to be high-iron tholeiitic basalt (Figure 6c). The mantle-normalized multi-element patterns (Figure 7c) show mainly flat profiles with Th < Nb < La characteristic of the depleted basalts of the Lumby North assemblage (Tomlinson et al. 1999). Two samples including 96DS109 representing gneissic amphibolite at the northeast end of the belt are slightly LREE enriched.

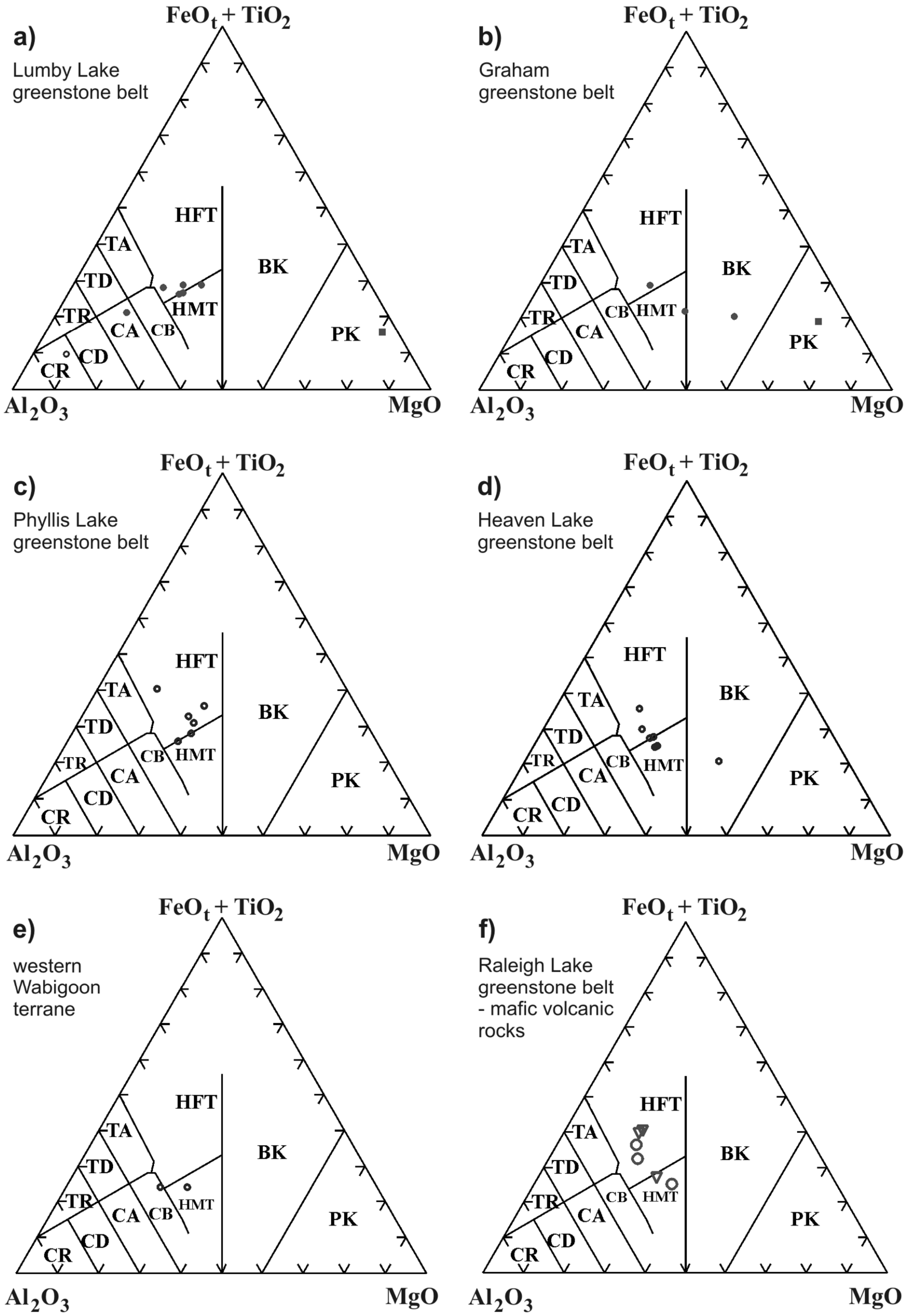

Figure 6. Jensen cation plots of rock samples from major greenstone belts (*after* Jensen 1976).

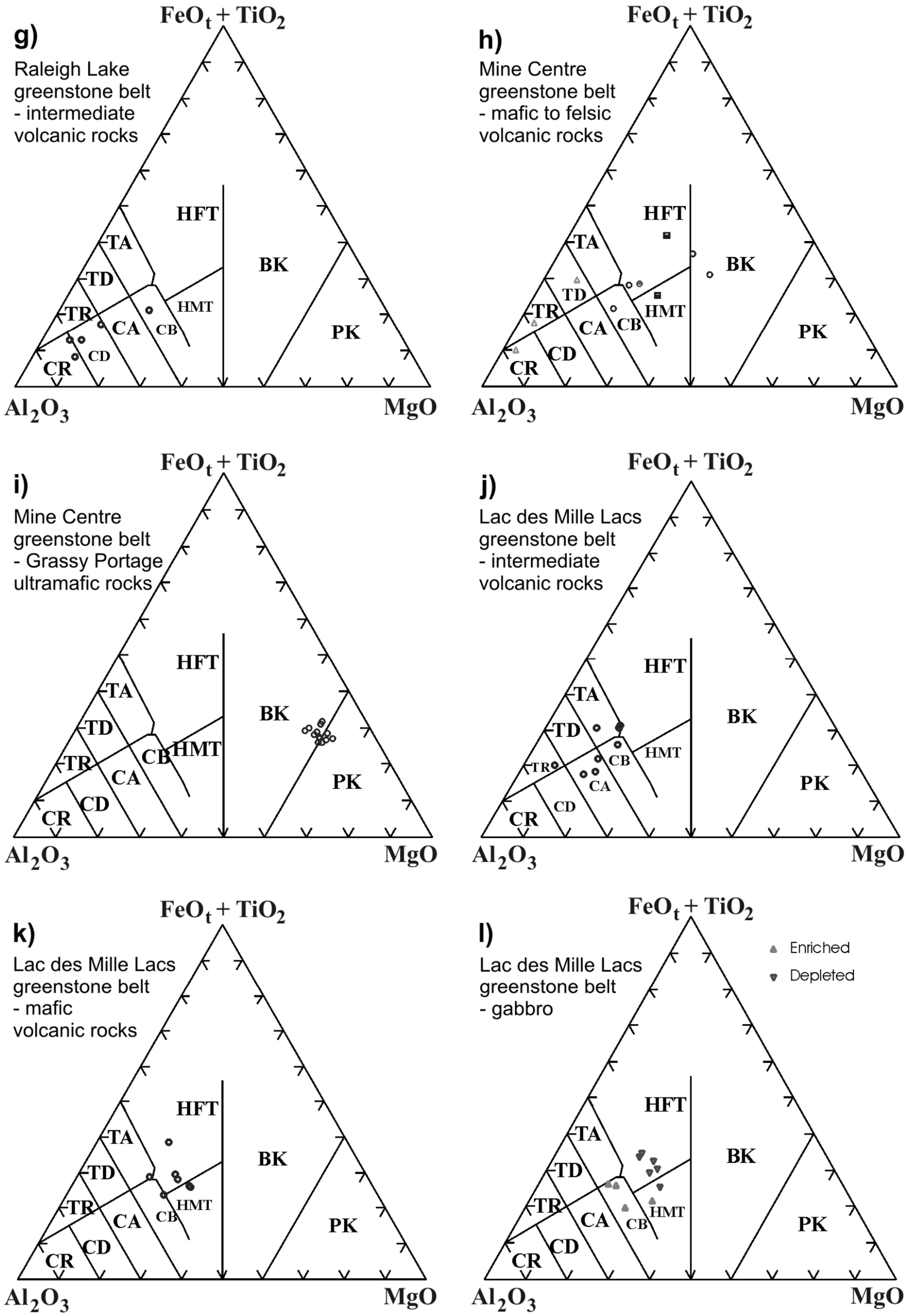

Figure 6. continued.

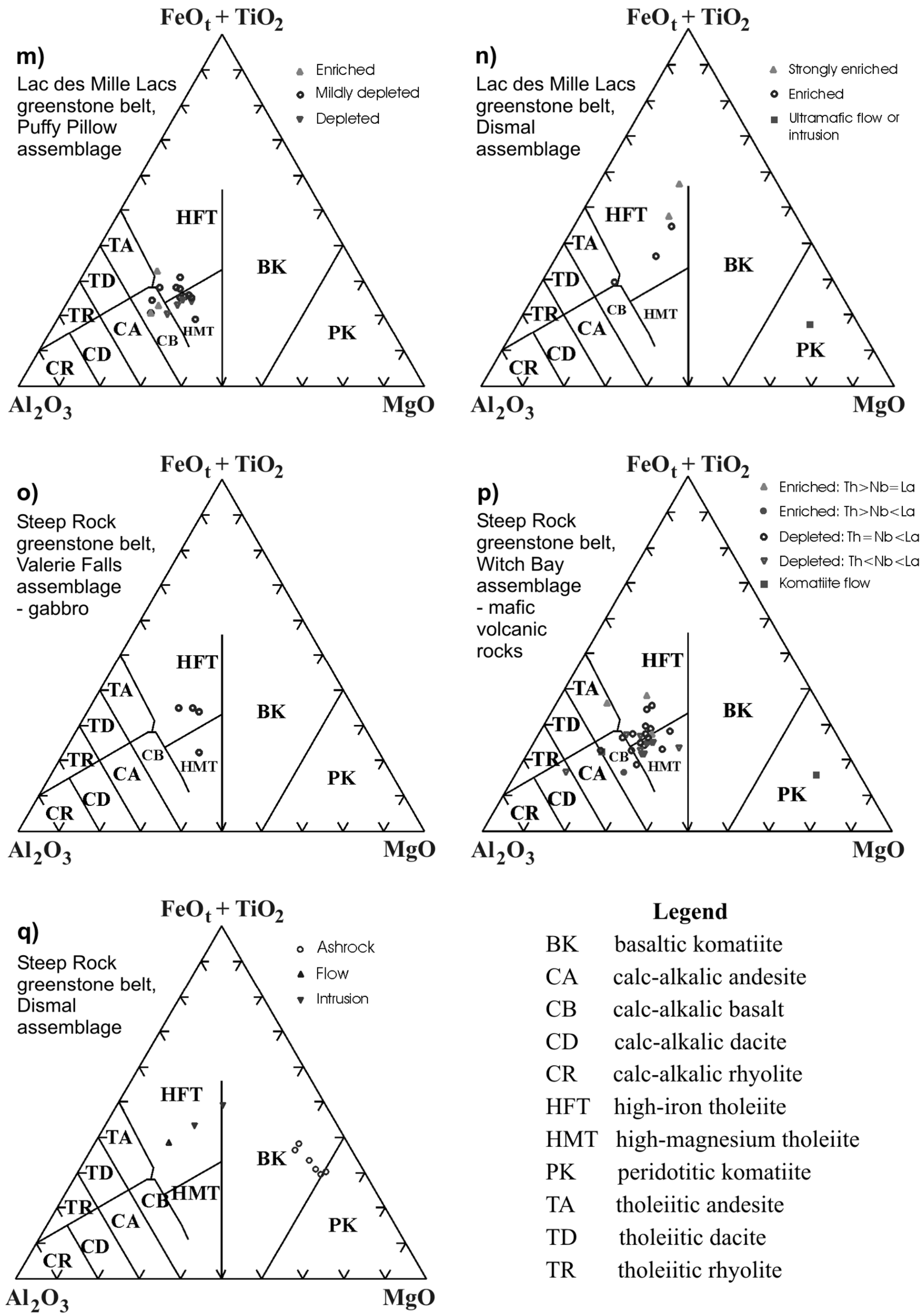

Figure 6. continued

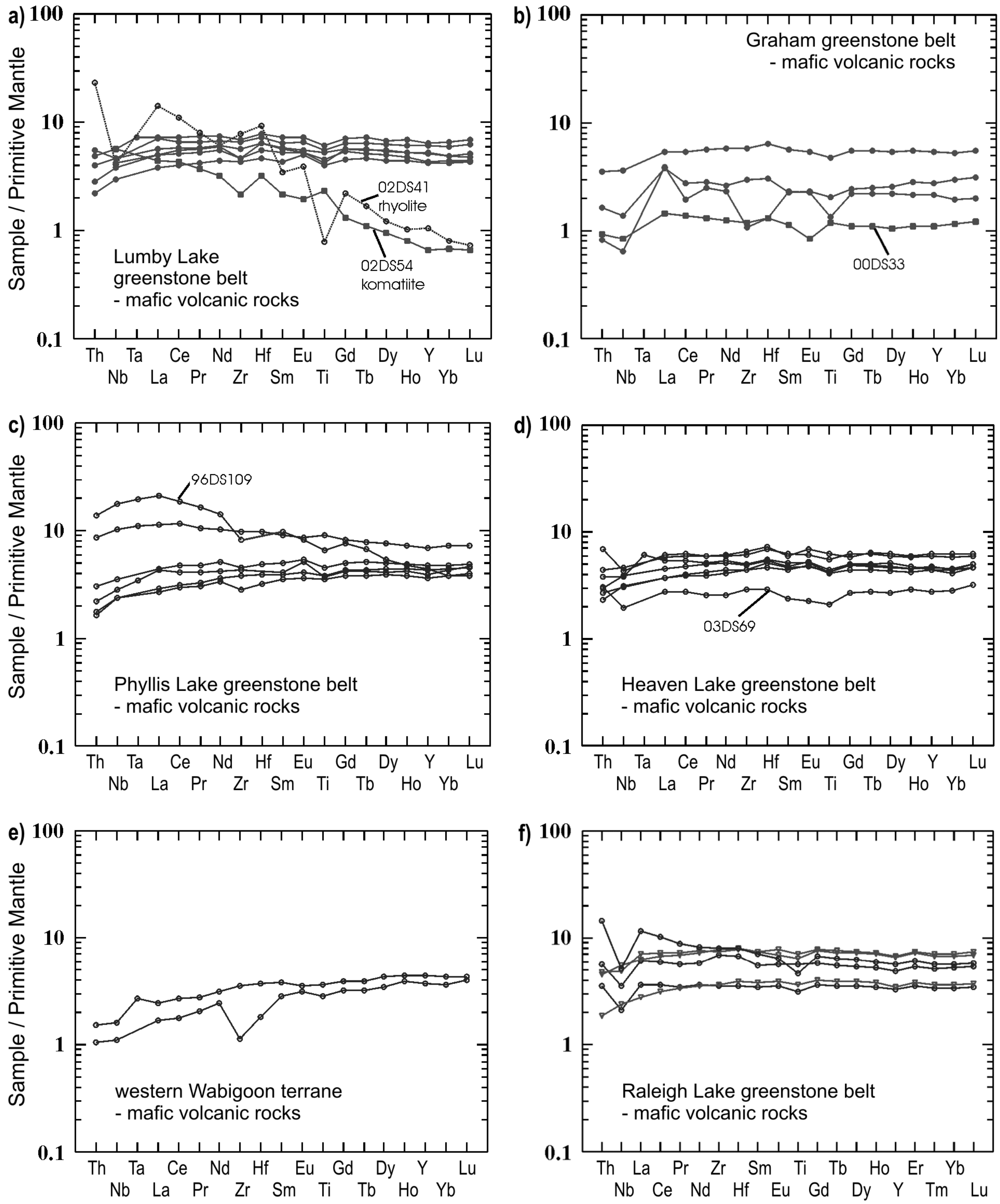

Figure 7. Primitive mantle-normalized trace element plots of samples from major greenstone belts (using the normalizing values of Sun and McDonough 1989).

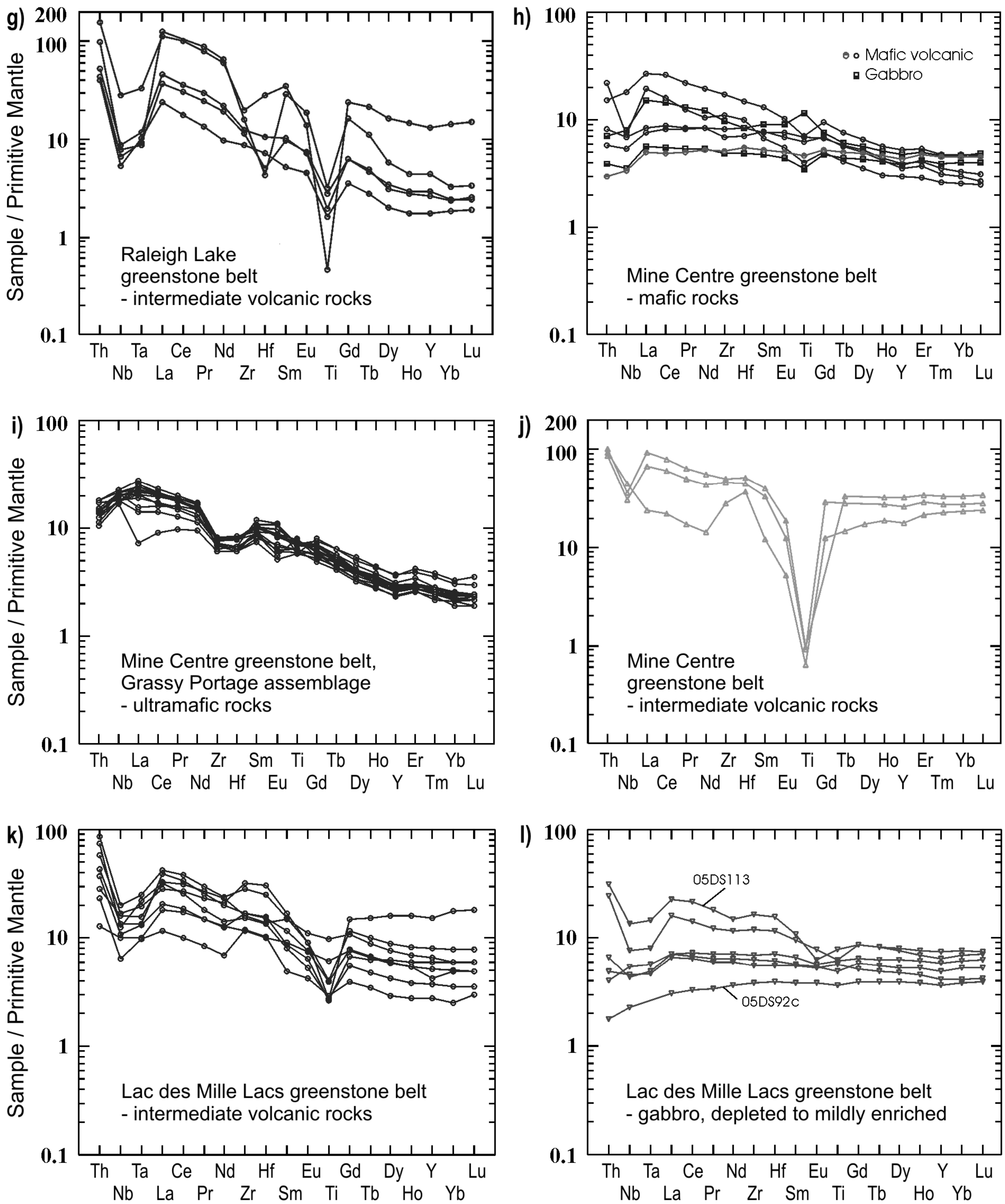

Figure 7. continued.

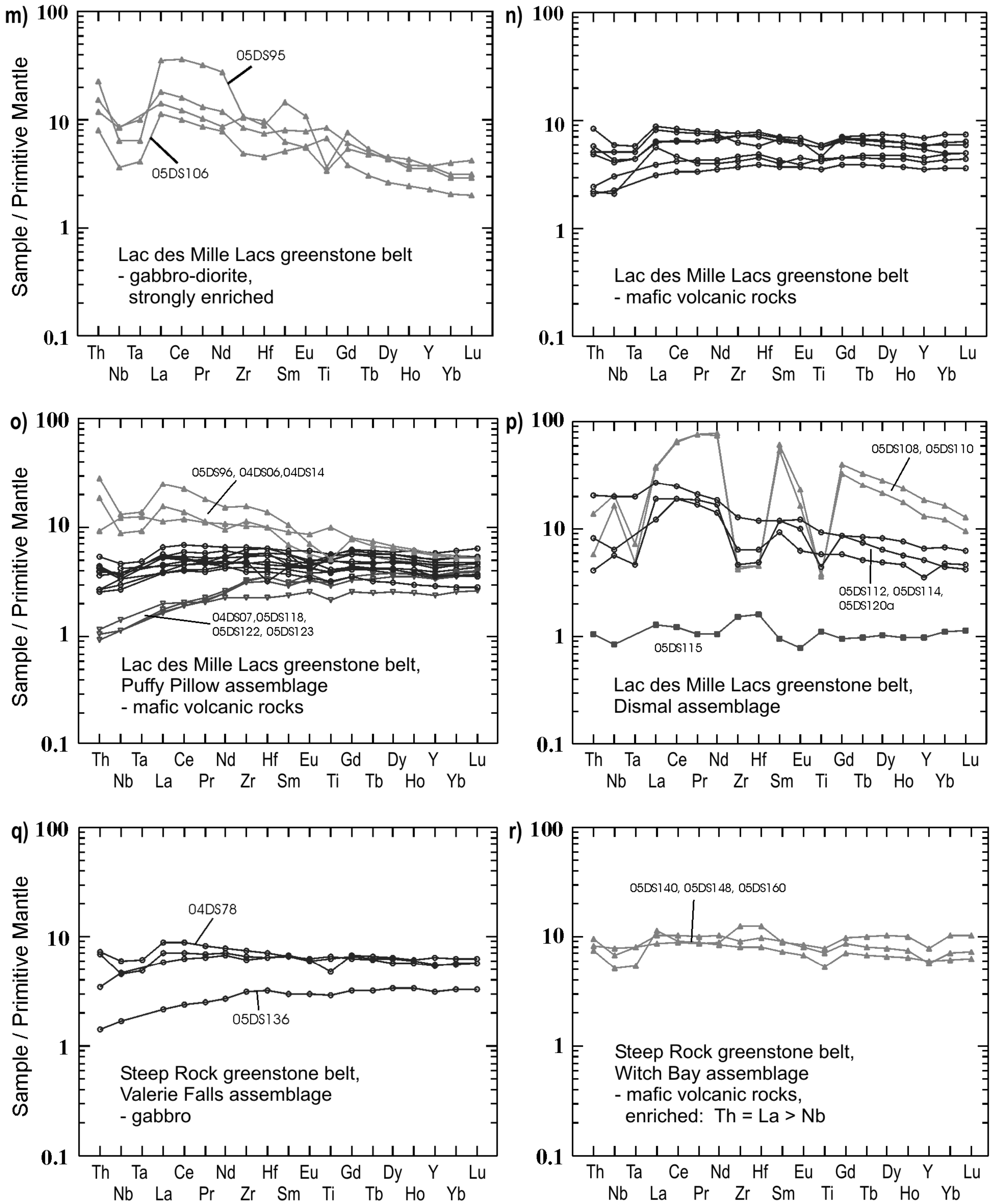

Figure 7. continued

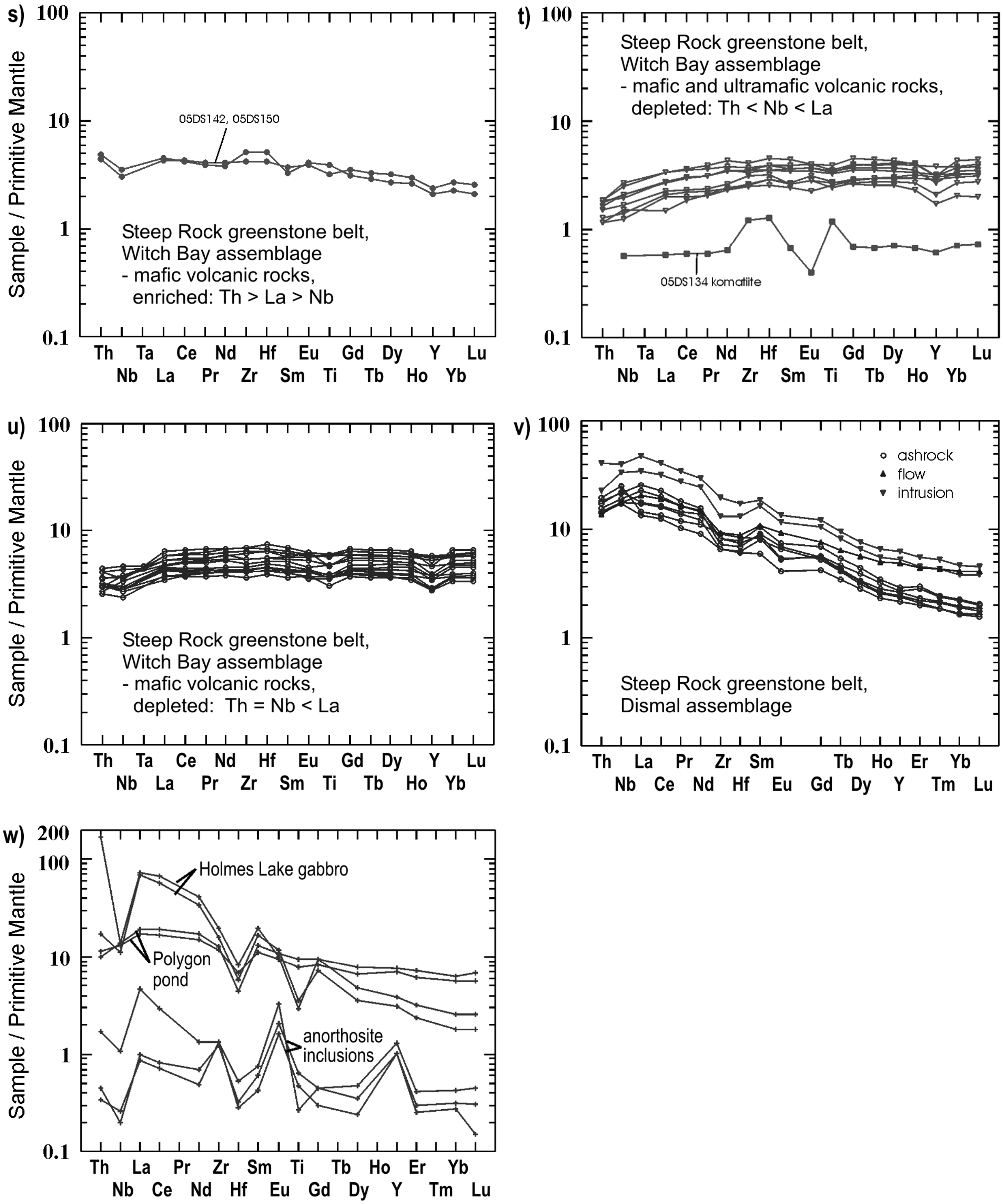

Figure 7. continued

Heaven Lake Greenstone Belt

The Heaven Lake greenstone belt ("Heaven Lake belt") attains a width of up to 5 km and extends easterly over a distance of 60 km at the east side of the central Wabigoon Subprovince area. Eastward, the Heaven Lake belt broadens and is extensively blanketed by Proterozoic diabase sills of the Nipigon sill complex (Hart, MacDonald and Lepine 2001a). Within the present area, the greenstone sequences of the Heaven Lake belt are composed entirely of mafic metavolcanic flows with rare thin interflow iron formation. The mafic rocks are a mix of massive and pillowed flows with associated gabbro intrusions. Metamorphism is generally at amphibolite facies and the combination of metamorphism and deformation has promoted development of amphibole gneisses through most of the belt.

The westernmost section of the Heaven Lake belt (approximately 20 km long) is unexposed and the geometry of the belt in this area is interpreted from aeromagnetic data. The aeromagnetic map (Geological Survey of Canada–Ontario Department of Mines 1965d) shows prominent magnetic anomalies extending southeasterly from the Heaven Lake belt in the areas of Herbert and Muise lakes. Diamond drilling of the Muise Lake anomaly indicates that the anomaly is likely associated with gabbro; the diamond-drill core is stored at the Thunder Bay Resident Geologist Office Conmee diamond-drill core storage site west of Thunder Bay.

Two assemblages including the Whitton and Whistle assemblages are defined by mapping and geochronology within the Heaven Lake belt east of the present area (Tomlinson et al. 2003). Of these, the Whitton assemblage extends westerly and includes the Heaven Lake belt within the present area, whereas the Whistle assemblage terminates to the east. A felsic tuff and a quartz porphyritic dike from the Whitton assemblage have identical ages of 2953 Ma (*see* Table 2, No. 141 and 140). A felsic tuff and a quartz + feldspar porphyry intrusion of the Whistle assemblage have ages of 2729 and 2728 Ma (*see* Table 2, Nos. 60 and 61) and the intrusion contains inherited zircon as old as 2779 Ma. The Whitton assemblage, together with the Lumby North assemblage and Phyllis Lake greenstone belt, are the main components of the Whitton domain—a discontinuous strip of 2953 to 2963 Ma crust in the central Wabigoon Subprovince area (*see* Figure 3). The Whistle assemblage of the Heaven Lake belt represents part of the eastern extension of the Whistle domain.

Mafic metavolcanic rocks of the Whitton assemblage range compositionally from basaltic komatiite to basalt (Figure 6d). The mantle-normalized trace element profiles of these samples are flat (Figure 7d) although 2 groups of samples, including those such as 03DS69 with Th > Nb and those with Th < Nb, are recognized. Three samples of drill core representing the Muise Lake gabbro have primitive mantle-normalized trace element profiles that are enriched in incompatible elements with deep troughs for Ti, Hf–Zr and Nb (not shown). The Muise Lake gabbro is compositionally different from the mafic metavolcanic rocks of the Whitton assemblage and possibly represents a younger intrusion.

Garden Lake Greenstone Belt

The Garden Lake greenstone belt ("Garden Lake belt") extends easterly over a distance of about 50 km with only a small part of the belt occurring within the northeast corner of the central Wabigoon Subprovince area. The Garden Lake belt is dominated by south-facing mafic metavolcanic flows with thin units of felsic metavolcanic rocks and iron formation. A thin unit of conglomerate extends along the central axis of the belt (Hart et al. 2000). Tomlinson et al. (1997, 2004) reported an age of 2726 Ma and an ε_{Nd} value of –2.35 for a thin rhyolite unit on the south side of the belt and an age of 2710 Ma for a quartz + feldspar porphyry intrusion in the eastern part of the Garden Lake belt. The negative ε_{Nd} value provides evidence that the mafic magma interacted with much older crustal material such as occurs in the Winnipeg River terrane; Tomlinson et al. (2004) interpreted the Garden Lake belt to be part of the Winnipeg River terrane.

According to Tomlinson et al. (1997), the mafic metavolcanic rocks of the Garden Lake belt are tholeiitic basalt with 2 distinct trace element patterns. The lavas from Garden Lake in the northeastern part of the belt have flat mantle-normalized trace element profiles with slight depletion in Th and Nb. In contrast, the majority of lavas from Kerns Lake in the southwestern part of the belt are enriched in Th and LREE with negative Nb anomalies. The contrasting geochemical signatures may indicate 2 assemblages of metavolcanic rocks in the Garden Lake belt although extent and boundaries of the assemblages is not well known on the basis of current information.

Lac des Iles Greenstone Belt

The Lac des Iles greenstone belt ("Lac des Iles belt") represents several greenstone slivers at the east margin of the central Wabigoon Subprovince area. These greenstone slivers coalesce east of the present area and have been traced 100 km eastward beneath Lake Nipigon on the basis of aeromagnetic patterns to join with greenstone sequences in the Beardmore–Geraldton area (Thurston and Davis 1985). In the present area, a thin panel of the Lac des Iles belt extends from south of Legris Lake to south of Wakinoo Lake (*see* Map P.2229). Greenstone sequences situated up to 3 km north of Legris Lake are partly overlain by Proterozoic diabase sills. These extend east of the present area and comprise part of a south-younging northern panel of the Lac des Iles belt (Hart, MacDonald and Lepine 2001b). A dacite tuff from the northern panel is dated at 2728 Ma (Map No. 62 of *see* Table 2). Kaye (1969) recognized a third panel of mafic metavolcanic rocks within metasedimentary sequences south of the present area and grouped these with the Lac des Iles greenstone belt.

The thin panel of the Lac des Iles greenstone belt is dominated by well-foliated to gneissic mafic metavolcanic rocks although south-younging pillows are recognized south of Wakinoo Lake. Narrow units of green, intermediate to felsic, commonly fragmental metavolcanic rocks as well as chert units are mingled with the mafic sequences. Also present within the greenstone belt as well as in metasedimentary rocks to the south are elongate units of grey, feldspar-phyric and typically fragmental felsic metavolcanic rocks. The grey siliceous metavolcanic units are possibly alkalic as indicated by feldspar phenocrysts.

Tomlinson et al. (1996, 1997) noted that the mafic metavolcanic rocks in the Beardmore–Geraldton area and in the Lac des Iles belt are tholeiitic to calc-alkalic basalts and andesites. The primitive mantle-normalized trace element profiles show that the basalts and andesites are enriched in LREE and Th with negative Nb and Ti anomalies characteristic of lavas erupted in an arc setting.

Western Wabigoon Terrane (Otukamamoan Lake and Raleigh Lake Greenstone Belts)

The western Wabigoon terrane, also known as the Kakagi Lake–Savant Lake greenstone belt, is a broad anastomosed belt of metavolcanic and metasedimentary rocks interspersed with oval to irregular felsic plutons extending from Minnesota northeastward through Sioux Lookout to Savant Lake (*see* Figure 1). Parts of this major greenstone sequence extend into the present area and are represented by greenstone belts at Otukamamoan Lake and Raleigh Lake (*see* Figure 5; *see* Map P.2229). Greenstone sequences of the western Wabigoon terrane are interpreted to have developed in a simatic environment at about 2745 to 2712 Ma and to have been tectonically emplaced onto the Winnipeg River and Marmion terranes at 2703 to 2695 Ma (Davis, Sutcliffe and Trowell 1988; Sanborn-Barrie and Skulski 2006).

The Otukamamoan Lake greenstone belt ("Otukamamoan belt") is dominated by mafic metavolcanic flows and a lesser component of intermediate to felsic flows, tuffs and fragmental metavolcanic rocks (Blackburn 1973; Berger 1991; Smith 1993). Within the present area, mafic metavolcanic rocks of the

Otukamamoan belt taper eastward within several narrow curved and branched arms and are transitional to mafic tonalite gneisses of the Eltrut gneiss complex (*see* Figure 5). Several narrow, curved greenstone slivers within the gneisses are considered to represent detached segments of the Otukamamoan belt. Mafic metavolcanic rocks within the greenstone arms are highly strained and metamorphosed, typically consisting of foliated to lineated amphibole gneisses. Zones of shallow west-dipping mylonite occur at the east margin of the Otukamamoan belt at Eagle Rock Lake (*see* Map P.2229) and possibly represent the thrust fault on which the greenstone sequences were displaced onto the central Wabigoon Subprovince basement.

Two samples of mafic metavolcanic rock from the Otukamamoan belt are classified as calc-alkalic basalt and high-Mg tholeiitic basalt (Figure 6e) and have mantle-normalized trace element profiles that are sloped to the left due to depletion in incompatible elements with Th < Nb < La (Figure 7e). Smith (1993) reported a larger set of chemical analyses for metavolcanic rocks at Vista Lake (northwest of Otukamamoan Lake) the compositions of which ranged from high-Mg basalt through calc-alkalic basalt, andesite, dacite and rhyolite.

Near Kenora at Lake of the Woods, Ayer and Davis (1997) reported 3 assemblages of metavolcanic rocks, the oldest of which includes Mg- and Fe-tholeiitic basalts with flat to slightly incompatible-element-enriched trace element profiles belonging to the Lower Keewatin assemblage. The older basalts are overlain by diverse mafic to felsic sequences of the Upper Keewatin assemblage that are dominantly calc-alkalic basalt to rhyolite with incompatible-element-enriched trace element profiles. Also in the Upper Keewatin assemblage is a suite of Mg-tholeiitic basalt, basaltic komatiite and komatiite with trace element profiles showing depletion in incompatible elements compared to compatible elements. The Electrum assemblage overlies the Upper Keewatin assemblage at Lake of the Woods and consists of conglomerate, sandstone and alkalic metavolcanic rocks. In comparison, the diverse metavolcanic sequences west of Otukamamoan Lake studied by Smith (1993) appear to be chemically comparable to the Upper Keewatin assemblage at Lake of the Woods. The 2 samples of basalt from the eastern arms of the Otukamamoan belt (*see* Figures 6e and 7e) are comparable to either the Lower Keewatin assemblage or the more primitive mafic suite within the Upper Keewatin assemblage.

The Raleigh Lake greenstone belt (“Raleigh Lake belt”) extends southeasterly over a distance of 50 km into the central Wabigoon Subprovince area and is intruded internally by oval granite plutons and externally by the White Otter and Indian batholiths (*see* Figure 5). Although dominated by mafic metavolcanic flows, the Raleigh Lake belt contains about 30% intermediate to felsic fragmental metavolcanic rocks, one sample of which has an age of 2730 Ma (*see* Table 2, No. 82). A thin unit of migmatized wacke occurs at the south side in contact with the White Otter batholith. A mineral assemblage of hornblende + garnet + plagioclase is common in the metavolcanic sequences and the metasedimentary migmatites locally show an assemblage of garnet + biotite + sillimanite + plagioclase. Although large undeformed pillows are observed in central parts of the belt, strata at margins and in the thin tapered extensions of the belt are strongly foliated.

Mafic metavolcanic rocks of the Raleigh Lake belt are basaltic in composition (Figure 6f). The mantle-normalized trace element profiles (Figure 7f) show 2 types of basalt. Three samples are slightly enriched in LREE and Th with Nb and Ti troughs. The trace element profiles of another 3 samples are flat for compatible elements and are depleted in incompatible elements with Th < Nb < La. Sample 06DS07 (not shown), representing a gabbro dike in intermediate metavolcanic rocks, is also depleted in incompatible elements.

The intermediate to felsic metavolcanic rocks range compositionally from calc-alkalic basalt to rhyolite (Figure 6g) and have trace element profiles strongly enriched in incompatible elements and deep troughs for Nb, Ta and Ti (Figure 7g). The intermediate to felsic metavolcanic rocks mainly have Zr/Y >10 with high Sr and are classified as FI type (Lesher et al. 1986), although sample 96DS58 from the centre of the belt is an FII type with Zr/Y = 4 at Y = 60 ppm, low Sr and a negative Eu anomaly. Broadly, the metavolcanic rocks of the Raleigh Lake belt are chemically comparable to those of the Upper Keewatin assemblage of Ayer and Davis (1997) at Lake of the Woods. The incompatible-element-enriched basalt to rhyolite sequences at Raleigh Lake are similar to those of the calc-alkalic part of the Upper Keewatin assemblage at Lake of the Woods and the depleted basalts at Raleigh Lake are chemically akin to the depleted mafic metavolcanic rocks of the Upper Keewatin assemblage.

Mine Centre Greenstone Belt

The Mine Centre greenstone belt ("Mine Centre belt") extends west from Calm Lake over a distance of more than 60 km through Mine Centre to Fort Frances. Metavolcanic rocks of the Mine Centre belt are in contact with tonalite and gneisses of the Wabigoon Subprovince to the north and sandstone–siltstone sequences of the Quetico Subprovince to the south. The east- and east-northeast-trending Quetico and Seine River faults merge within the Mine Centre belt (*see* Figure 5). The part of the Mine Centre belt situated between the 2 faults has been described as a fault graben or wrench zone characterized by lenticular lithostratigraphic domains separated by steep boundary faults and a preponderance of Z-style asymmetric folds (Poulsen 1986).

The eastern Mine Centre belt situated north of the Quetico fault is dominated by massive to pillowed mafic metavolcanic rocks. The mafic metavolcanic rocks are intruded by crescentic to oval gabbro intrusions in the area between the Little Turtle and Seine rivers. Well-bedded sandstone–siltstone sequences with rare thin iron formation beds occur north of the mafic metavolcanic rocks at Calm Lake and are intruded and contact metamorphosed by the oval Joe and Hillyer biotite granite plutons. Mafic metavolcanic rocks of the eastern Mine Centre belt are overlain by coarse conglomerate and sandstone of the Seine sequence or Seine metasediments (Poulsen 2000b). The conglomerate and thin lenses of metavolcanic rocks occupy the wedge between the Quetico and Seine River faults (*see* Figure 3, Shoal Basin).

The western Mine Centre belt is lithologically diverse. A lenticular domain of intermediate to felsic metavolcanic flows and fragmental rocks extends west from Mine Centre curving south of the Ottertail stock. Numerous gabbro sills of variable thickness cut the intermediate to felsic metavolcanic rocks and are possibly responsible for local elongate zones of residual high magnetic field in the metavolcanic unit (Ontario Geological Survey 2009d). The largest sills at Grassy Portage (1 to 10 km west of the present area) and Bad Vermillion Lake (*see* Figure 5, southwest of Mine Centre) are layered and compositionally variable from melagabbro to anorthosite. At Redgut Bay (5 km west of the Ottertail stock), metavolcanic and gabbroic rocks are interleaved with sandstone and siltstone. This sedimentary unit represents part of the Coutchiching Series of Lawson (1913) whose stratigraphic relation with adjacent metavolcanic rocks was hotly debated in the early part of the twentieth century (*see* "Previous Mapping and Exploration"). Also a few kilometres west of Redgut Bay, a curved unit of possible ultramafic lapilli tuff (Grassy Portage ultramafic unit) occurs within mafic metavolcanic sequences.

Supracrustal rocks of the western Mine Centre belt are intruded by felsic plutonic rocks. These include lenticular units of biotite tonalite that occur on both flanks of the Bad Vermilion gabbro complex and host many of the historic gold mines of the Mine Centre area. Oval to irregular plutons of biotite granite have intruded the Coutchiching sedimentary unit at Redgut Bay. The Ottertail stock of the sanukitoid suite is compositionally variable from monzodiorite to granite and represents the youngest known intrusion in the Mine Centre belt.

U/Pb age determinations of 2728 and 2727 Ma were obtained for the intermediate to felsic metavolcanic unit at Mine Centre (*see* Table 2, Nos. 72 and 70). The mineralized tonalite unit at Bad Vermilion Lake has an age of 2728 Ma and the Ottertail stock has an age of 2686 Ma (*see* Table 2, Nos. 71 and 119). The Coutchiching metasedimentary unit at Redgut Bay has detrital zircon grains ranging in age from 3059 to 2704 Ma (*see* Table 2, No. 105) and was evidently deposited after 2704 Ma postdating volcanism. Likewise, a tonalite cobble in the Seine metasediments east of Mine Centre has an age of 2696 Ma (*see* Table 2, No. 104) representing a maximum depositional age for sedimentary sequences in the Shoal Basin.

The comparatively young supracrustal rocks of the Mine Centre belt overlie or have been displaced onto older felsic plutonic and gneissic rocks of the Hillyer and Central Wabigoon domains (*see* Figure 3). Felsic plutonic rocks of the Hillyer domain contain large anorthositic inclusions such as at Manion Lake and are intruded by a tapered pluton of gabbro to hornblendite at Holmes Lake (5 km northwest of Calm Lake).

Metavolcanic and mafic intrusive rocks of the Mine Centre belt are chemically variable. Mafic metavolcanic rocks and gabbro sills are basaltic komatiite to basalt in composition and 3 samples of the intermediate to felsic metavolcanic unit extending west from Mine Centre are dacitic to rhyolitic in composition (Figure 6h). The mantle-normalized trace element profiles of mafic metavolcanic and gabbro samples are slightly sloped to the right due to variable enrichment in LREEs (Figure 7h). Most samples have Th ≈ Nb < La and slight positive or negative Ti anomalies. Samples of the Grassy Portage ultramafic unit are rich in MgO and $Fe_2O_3^T$ (MgO = 17 to 23 wt %; $Fe_2O_3^T$ = 17 to 19 wt %) and low in Al_2O_3 (Al_2O_3 = 4 to 7 wt %). Although samples plot within the fields of komatiite to basaltic komatiite using the diagram of Jensen (1976; Figure 6i), these rocks have been shown to be chemically distinct from normal Archean komatiites largely on the basis of high Fe-content and are called ferropicrites (Goldstein and Francis 2008). Samples of the Grassy Portage ultramafic unit are consistently LREE enriched with Th < Nb < La and a trough for Zr and Hf (Figure 7i). Trace element characteristics of the Grassy Portage ultramafic unit are similar to those of the ashrock of the Dismal assemblage in the Steep Rock belt (discussed below). Mantle-normalized trace element profiles of the intermediate to felsic metavolcanic rocks are relatively flat with deep troughs for Eu and Ti (*see* Figure 7j). These samples have Zr/Y ≈ 4 and Sr <50 ppm and are classified as FIII rhyolite (Lesher et al. 1986).

Three chemically distinct groups of anorthositic to gabbroic intrusions are recognized north of the Mine Centre belt within plutonic rocks (Figure 7w). For example, the anorthositic inclusions within plutonic rocks of the Hillyer domain have comparatively flat trace element profiles with spikes for Y, Eu and LREE. In contrast, the Holmes Lake gabbro is more enriched in all trace elements, particularly the incompatible elements with troughs for Ti, Hf and Nb. Finally, the crescentic gabbro body intruding mafic metavolcanic rocks at Polygon Pond has a relatively flat trace element profile not much different from that of enclosing mafic metavolcanic rocks (compare Figures 7h and 7w). These intrusions probably reflect 3 different magmas emplaced episodically in the Hillyer domain and Mine Centre belt.

Lac des Mille Lacs Greenstone Belt

The Lac des Mille Lacs greenstone belt (“Lac des Mille Lacs belt”) extends along the southern boundary of the Wabigoon Subprovince over a distance of 100 km from Atikokan in the west to Highway 17 in the east. The Lac des Mille Lacs belt attains a maximum width of 6 km and is bounded on the north by felsic plutonic rocks including biotite tonalite of the Marmion batholith. Southward, greenstone sequences of the Lac des Mille Lacs belt are juxtaposed with sandstone–siltstone sequences of the Quetico Subprovince across the Quetico and Shelby faults. At Lac des Mille Lacs, a narrow band of amphibole gneiss, probably representing an arm of the Lac des Mille Lacs belt, extends eastward into the Quetico Subprovince adjacent to the easterly trending Quetico fault (*see* Map P.2229).

Gabbro is the dominant type of rock within the Lac des Mille Lacs belt between Atikokan and Lac des Mille Lacs. This major gabbroic unit appears to be made up of a network of deformed crosscutting dikes and a minor component of mafic flows. At Crooked Pine Lake, crosscutting relations show that the earliest dikes are coarse-grained leucogabbro superseded by dikes of gabbro and fine-grained melagabbro.

Locally, large outcrops of fine- to medium-grained mafic material are observed and may represent large dikes or mafic metavolcanic flows. The northern contact with the Marmion batholith is gradational over a distance of up to 1 km and is typically marked by a gradual southward-increasing abundance of gabbro dikes. Blocks of deformed tonalite, probably representing fragments of the Marmion batholith occur widely within the gabbro unit. The gabbro is metamorphosed to an assemblage of hornblende and plagioclase.

The gabbro has proven difficult to date due to lack of zircon except for one sample of leucogabbro from north of Crooked Pine Lake that has an age of 3005 Ma (*see* Table 2, No. 11) indicating either that the gabbro is only slightly younger than the Marmion batholith or that it has inherited zircon belonging to the Marmion batholith. Crosscutting relations imply that there can be several generations of gabbro dikes. The gabbro and associated mafic flows range compositionally from tholeiitic to calc-alkalic basalt (Figures 6k and 6l) and have variable enrichment in incompatible elements. The mantle-normalized trace element profiles of the early leucogabbro dikes are strongly sloped to the right due to enrichment in incompatible elements with troughs for Nb and Ta (Figure 7m). Later gabbro dikes and mafic flows range from mild enrichment to depletion in incompatible elements (compare samples 05DS113 and 05DS92c of Figures 7l).

Also present in the gabbro between Atikokan and Crooked Pine Lake are narrow, disrupted units of fine-grained, hard grey intermediate metavolcanic fragmental rocks. These have ages of 3008 and 3006 Ma (*see* Table 2, Nos. 13 and 17) and together with metavolcanic sequences of similar age in the southern Lumby Lake and eastern Finlayson Lake greenstone belts appear to represent rocks into which the Marmion batholith was emplaced. Eastward at Lac des Mille Lacs, intermediate metavolcanic rocks broaden to underlie the northern half of the belt and are made up of massive flows and subsidiary tuffs and breccias. These are generally andesitic to dacitic in composition (Figure 6j) and are enriched in incompatible elements with troughs for Nb, Ta and Ti (Figure 7k). A sample of schistose quartz-porphyritic material from the eastern shore of Lac des Mille Lacs has an age of 2830 Ma (*see* Table 2, No. 149).

The geochronology indicates at least 2 generations of intermediate metavolcanic rocks (~3008 and 2830 Ma) in the Lac des Mille Lacs belt. The extent of the 2830 Ma sequence is not well known on the basis of the present geochronology, but is possibly restricted to the Lac des Mille Lacs area. This poorly defined domain of 2830 Ma rocks (*see* Figure 3, Marmion South Margin domain) possibly represents an extension of the Pinecone–Savoy domain, which could wrap around the eastern end of the Marmion domain.

West of Lac des Mille Lacs, thin discontinuous units of conglomerate occur at the north margin of the Lac des Mille Lacs belt in contact with biotite tonalite of the Marmion batholith. Although highly strained, the clasts in the conglomerate appear to be mainly tonalite, quartz, rare dark material possibly representing gabbro and fine-grained felsic detritus of possible volcanic origin all of which are set in a foliated gritty matrix. A sample of the conglomerate from south of Bedivere Lake yielded detrital zircon grains ranging in age from 3007 to 2927 Ma (*see* Table 2, No. 46). Evidently, the conglomerate is derived by erosion of the Marmion batholith and material representative of the central Wabigoon domain and was deposited after 2927 Ma.

An abrupt change in lithology occurs in the southern part of the Lac des Mille Lacs belt where gabbro and felsic metavolcanic rocks give way to pillowed mafic metavolcanic sequences of the Puffy Pillow assemblage. Zoisite is a major mineral in these metavolcanic sequences and imparts a pale green colour to the rocks. In the area of Sapawe and Crooked Pine Lake, the Puffy Pillow assemblage can be as thin as a few tens of metres or absent at the south margin of the Lac des Mille Lacs belt. The assemblage has been strongly deformed by the Quetico fault in this area and rocks are schistose with stretched pillows. At Lac des Mille Lacs, the Puffy Pillow assemblage broadens to a width of 2 km and large oval pillows indicate that the strata face southward. A felsic metavolcanic unit within the mafic flows at Lac des Mille Lacs has an age of 2725 Ma (*see* Table 2, No. 64). Mafic lavas of the Puffy Pillow assemblage have a narrow range of major element compositions overlapping fields of tholeiitic and calc-alkalic basalt (Figure 6m), but have variable enrichment in incompatible elements (Figure 7o). Three subgroups of mafic metavolcanic flows are distinguished in Figure 7o including samples such as 05DS96 that are mildly enriched in incompatible elements, a large group of samples with more-or-less flat trace element profiles and a group including 04DS07 that are strongly depleted in incompatible elements.

Extending west from Crooked Pine Lake through Sapawe Lake to the Steep Rock Lake greenstone belt at Atikokan is a narrow highly deformed and locally discontinuous unit of mafic rocks comprising the Dismal assemblage. Rocks of the Dismal assemblage are typically dark fine-grained chloritic or tremolitic schists possibly representing metavolcanic flows although medium grained and somewhat more massive rocks suggestive of mafic to ultramafic intrusive material are locally observed. The Dismal assemblage has a characteristically high residual magnetic field (Ontario Geological Survey 2009b). Although the age of crystallization is not known, the ashrock unit of the Dismal assemblage in the Steep Rock Lake greenstone belt contains inherited zircon of 2780 Ma age indicating that it is much younger than the Marmion batholith.

Rocks of the Dismal assemblage in the Lac des Mille Lacs belt are mainly high-iron tholeiitic basalt with single samples of andesite and komatiite (Figure 6n). The trace element profiles for rocks such as sample 05DS112 of the Dismal assemblage are sloped to the right due to enrichment in incompatible elements including LREE (Figure 7p). Two samples of strongly Fe-enriched tholeiite (05DS108 and 05DS110 of *see* Figure 7p) have unusual concave-upward profiles due to strong enrichment in LREE and MREE relative to which the most incompatible elements including Th, Nb, Ta and La are depleted with Th < Nb < La. Deep troughs for Zr, Hf, Eu and Ti possibly indicate fractionation of zircon, feldspar and a titanium-bearing phase from these magmas. Sample 05DS115, from a small ultramafic mass on the north shore of Sapawe Lake, has a trace element profile not unlike that of primitive mantle (*see* Figure 7p); the relation of this small ultramafic mass to the Dismal assemblage is unclear.

Goldstein and Francis (2008) proposed the term “ferropicrite” for the ashrock unit of the Dismal assemblage in the Steep Rock Lake greenstone belt. Rocks of the Dismal assemblage in the Lac des Mille Lacs belt lack the high MgO of the ashrock (compare Figures 6n and 6q), but are nonetheless very high in iron and LREE and are perhaps best described as high-iron tholeiite. The Dismal assemblage has a strong magnetic anomaly (Ontario Geological Survey 2003) and hosts the magnetite deposits of the Atikokan Iron Mine as well as several iron-cobalt occurrences between Sapawe and Atikokan (Atikokan River occurrences of MacTavish 1999). Possibly, these iron deposits have originated due to the extremely high iron content of their host rock.

In summary, the ~3008 Ma intermediate metavolcanic rocks of the linear Lac des Mille Lacs belt and similar ancient volcanic components of the Lumby Lake and Finlayson Lake greenstone belts (discussed below) evidently represent remnants of the host rock into which the Marmion batholith was emplaced a few million years later. The Marmion batholith appears to have been rifted almost immediately after emplacement as indicated by the linear zone of gabbro dikes although it remains unclear whether the zircon in the gabbro (age of 3005 Ma) crystallized from the dike magma at the time that the dikes were emplaced or is inherited from the batholith. In the latter case, the gabbro dikes are known only to be younger than the batholith. Episodes of volcanism occurred much later at 2830 and 2725 Ma. The extent of the 2830 million-years-old volcanic sequences (*see* Figure 3, Marmion South Margin domain) is poorly known, but possibly represents an extension of the Pinecone–Savoy domain around the eastern Marmion domain. Finally, the Puffy Pillow assemblage may represent oceanic material accreted to the Marmion terrane after 2725 Ma.

Finlayson Lake Greenstone Belt

The Finlayson Lake greenstone belt ("Finlayson Lake belt") extends northeasterly over a distance of 30 km joining the Steep Rock Lake greenstone belt at Atikokan with the Lumby Lake greenstone belt in the central Wabigoon Subprovince area (*see* Figure 5). The Finlayson Lake belt is underlain mainly by mafic metavolcanic rocks with thin units of intermediate metavolcanic tuff and metasedimentary units variably composed of conglomerate, sandstone and siltstone. Biotite tonalite of the central Wabigoon domain intrudes the west side of the Finlayson Lake greenstone belt, whereas 2999 Ma hornblende tonalite to granodiorite (*see* Table 2, No. 18) intrudes the east margin and separates the Finlayson Lake greenstone belt from the Marmion batholith. The Marmion fault cuts the narrow northeastern end of the Finlayson Lake greenstone belt near the junction with the Lumby Lake greenstone belt.

Three assemblages of supracrustal rocks including the Finlayson East, Finlayson West and Witch Bay are interpreted within the Finlayson Lake greenstone belt. The Finlayson East assemblage extends along the eastern side of the Finlayson Lake belt and lies mainly east of the eastern shoreline of Finlayson Lake. This assemblage is composed of mafic metavolcanic flows with subsidiary units of intermediate metavolcanic tuffs or flows and clastic metasedimentary rocks. Rocks within this assemblage are generally metamorphosed to amphibolite facies and are well foliated. Conglomerate and sandstone yield detrital zircon grains ranging in age from 3002 to 2997 Ma (*see* Table 2, No. 4) suggesting that the sediments were derived exclusively from the Marmion domain. This and an age of 3003 Ma on a quartz-porphyritic intermediate metavolcanic rock (*see* Table 2, No. 10) are the basis for including the Finlayson East assemblage with the Marmion domain. A prominent linear northeast-trending magnetic anomaly in the Finlayson East assemblage (Ontario Geological Survey 2009c) marks the trace of a late magnetite-bearing mafic dike in the greenstone belt.

The Finlayson West assemblage occupies the northwestern half of the belt and is made up of massive to pillowed mafic metavolcanic rocks with thin units of felsic metavolcanic rocks and iron formation. The grade of metamorphism is generally at amphibolite facies. A felsic tuff on the northwest shore of Finlayson Lake and an adjacent tonalite have ages of 2931 and 2936 Ma, respectively (*see* Table 2, Nos. 25 and 26), demonstrating a similarity in age between rocks of the Finlayson West assemblage and those of the central Wabigoon domain. Narrow amphibolite units, probably representing parts of the the Finlayson West assemblage extend northwesterly and are complexly interfolded with the tonalitic Dashwa gneisses (*see* Figure 5; *see* Map P.2229). Available geochronology (*see* Table 2, No. 146) shows that the Dashwa gneisses are older than the Finlayson West assemblage and implies that the complex late-stage deformation may have interfolded these rocks of divergent age.

Mafic metavolcanic flows exposed on islands and along the shoreline of central Finlayson Lake are characterized by large oval pillows and a weak foliation. The mineral assemblage of these pillow lavas (chlorite + carbonate + albite) is typical of greenschist metamorphic grade and is markedly lower than that of the Finlayson East and Finlayson West assemblages. Although the pillow lavas in the central Finlayson Lake belt have proven difficult to date, these rocks are interpreted to represent a klippe of the Neoarchean Witch Bay assemblage of the Steep Rock Lake greenstone belt (*see* "Steep Rock Lake Greenstone Belt"). This interpretation is supported by LA-ICP–MS geochronology on detrital zircon grains in a small wacke unit within the assemblage (*see* Table 2, No. 78). This geochronology shows zircon grains ranging in age from 2678 to 3031 Ma and indicates the sediments were deposited during the Neoarchean with the majority of detritus probably originating from the Mesoarchean Marmion, Whitton and central Wabigoon domains as well as Neoarchean sources coeval with the Witch Bay assemblage (*see* Figure 3).

Tomlinson et al. (1999) reported that mafic metavolcanic rocks of the Finlayson Lake belt are chemically comparable to those of the Lumby Lake and Steep Rock Lake greenstone belts, although it is unclear what assemblages of the Finlayson Lake belt were sampled. No chemical analyses were made of rocks from the Finlayson Lake belt as part of this study.

Steep Rock Lake Greenstone Belt

The Steep Rock Lake greenstone belt (“Steep Rock belt”) extends easterly over a distance of 40 km from Calm Lake to Atikokan (*see* Figure 5). Southward, the Steep Rock Lake greenstone belt is in contact with metasedimentary sequences of the Quetico Subprovince. The Quetico fault is developed locally at this contact. Three major arms of the Steep Rock belt extend northerly over distances of 10 to 20 km into felsic plutonic rocks of the central Wabigoon Subprovince area. These include a curved and tapered arm extending north of Calm Lake, a central arm extending north through Nevison Lake and a part of the belt that merges with the Finlayson Lake belt north of Steep Rock Lake (*see* Figure 5).

The Steep Rock Lake greenstone belt has been extensively studied over the past century (*see* “Previous Mapping and Exploration”). The belt contains a rare and famous example of an Archean unconformity where the Steep Rock assemblage overlies the Marmion batholith. Iron was mined extensively from the Jolliffe Ore Zone Formation of the Steep Rock assemblage from 1944 to 1979.

The Steep Rock Lake greenstone belt is subdivided into several assemblages (Figure 8), some of which are well defined in extent and age, whereas others are less well known. The Perch assemblage is represented by quartz and feldspar porphyritic felsic flows, tuffs and fragmental metavolcanic rocks in the area of Modred Lake. The felsic metavolcanic unit includes local zones of homogeneous quartz and feldspar porphyry probably representing a subvolcanic intrusion and contains thin massive to pillowed mafic metavolcanic flows cut by gabbro dikes at Modred Lake. These volcanic and intrusive sequences are strongly deformed and altered to an assemblage of chlorite + carbonate + albite + quartz within the Ear Lake fault and, together with deformed tonalitic rocks to the north, are locally mineralized with gold at the Zephyr, Harold and Elizabeth showings (*see* Stone and Lennox 2007).

The felsic metavolcanic unit at Modred Lake is transitional to coarse sandstone with beds of felsic volcanic clasts at Little Falls and Barr lakes. The metasedimentary unit tapers northeasterly at the west margin of the Finlayson Lake belt. Fralick and King (1995) and Fralick, Hollings and King (2008) interpreted the metasedimentary sequence (their Little Falls Group) to represent a clastic debris apron formed adjacent to and deriving sediment from the felsic volcanic centre. This interpretation is supported by an age of 2999 Ma for the felsic metavolcanic unit at Modred Lake, 2998 Ma for felsic metavolcanic rocks at Little Dragon Lake in the central, northerly trending arm of the belt and a single population of 2997 Ma detrital zircon in the metasedimentary rocks at Little Falls Lake (*see* Table 2, Nos. 5, 12 and 3, respectively).

Irregular, bifurcated units of grey intermediate to felsic metavolcanic flows, tuffs and breccias and lesser units of mafic metavolcanic flows occur though the western part of the Steep Rock belt in the area north of Perch Lake and westward to Calm Lake (*see* Map P.2229). These supracrustal units are intruded by voluminous masses and sills of medium- to coarse-grained gabbro. The gabbro is normally well foliated and metamorphosed to amphibolite facies, although chlorite + actinolite schists are locally developed. A sample of felsic metavolcanic fragmental rock from north of Perch Lake has an age of 2926 Ma (*see* Table 2, No. 28) and indicates that the metavolcanic rocks north of Perch Lake are similar in age to those of the Finlayson West assemblage (2931 Ma) and belong to the central Wabigoon domain (*see* Figure 3). The gabbroic masses (age not determined) and 2926 Ma felsic volcanic inclusions are intruded by 2883 Ma hornblende tonalite at Miranda Lake (*see* Table 2, No. 40); therefore, the gabbro is constrained in age between 2883 and 2926 Ma. The Perch assemblage is partly flanked by mafic rocks possibly comprising distinct assemblages. Gabbro, at the west side of the Steep Rock belt (west of Barr Lake, *see* Figure 8) cuts felsic metavolcanic rocks of the Perch assemblage and extends northeasterly, eventually pinching out at the west side of the Perch assemblage. These mafic sequences (age not determined) can represent any of a late stage of 3.0 Ga Perch magmatism, 2.96 Ga Whitton magmatism, part of the 2.93 Ga Finlayson West magmatism or an unknown event.

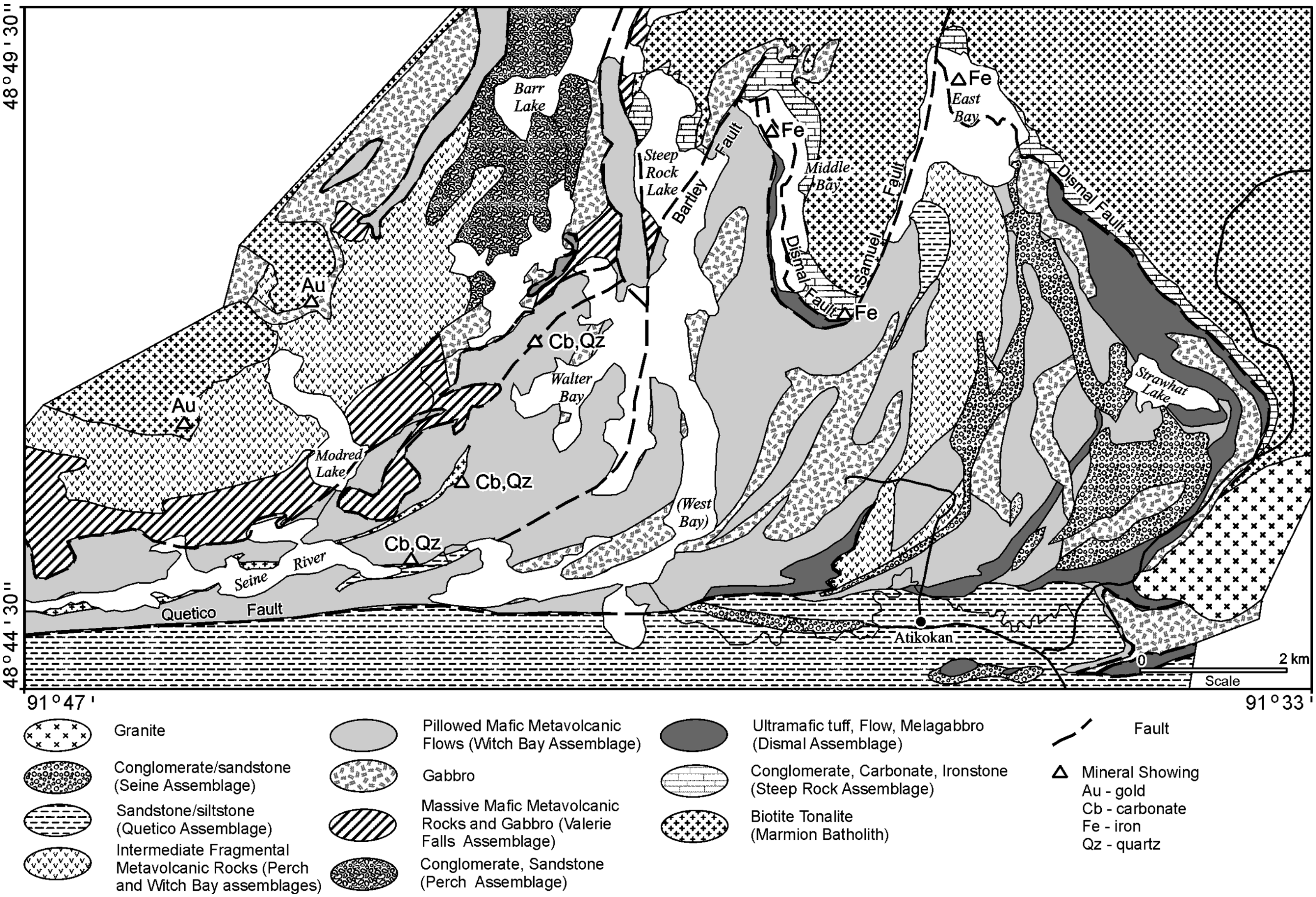

Figure 8. Geology and assemblages of the Steep Rock Lake greenstone belt.

The Valerie Falls assemblage represents a band of gabbro and massive mafic metavolcanic flows extending along the east side of the Perch assemblage from Barr Lake through Modred Lake and west to the Eye River (*see* Figure 8). The dominantly mafic lithology and amphibolite grade of metamorphism are characteristic of the Valerie Falls assemblage, although the tectonic relation of this undated assemblage to the Perch assemblage is unknown due to poor exposure of their mutual contact. Four samples of the gabbroic rocks north of Perch Lake are classified as tholeiitic basalt (*see* Figure 6o) and have trace element profiles that are variably depleted or mildly enriched in incompatible elements (compare the profiles of samples 05DS136 and 04DS78 of Figure 7q).

The Steep Rock assemblage attains a width of several hundred metres and extends discontinuously along the northeast margin of the Steep Rock belt (*see* Figure 8). The Steep Rock assemblage represents the 3 lower formations of the Steep Rock Group of Wilks and Nisbet (1988). In ascending stratigraphic order, the formations include the Wagita Formation (tonalite cobble conglomerate and sandstone overlying the Marmion batholith), Mosher Formation (grey dolomitic limestone) and the Jolliffe Formation (goethite-hematite deposits probably representing altered iron formation). The Dismal formation (ultramafic lapilli tuff), which was previously considered to represent the upper formation of the Steep Rock Group is now interpreted as part of the Dismal assemblage.

Gabbro dikes are common in the Marmion batholith adjacent to the eastern Steep Rock belt and zircon from one dike has an age of 3002 Ma (*see* Table 2, No. 15). As is the case for gabbro of the Lac des Mille Lacs greenstone belt, this age indicates either that numerous dikes were emplaced shortly after intrusion of the Marmion batholith or else that the dikes intruded at a later date and inherited zircon from the batholith. In either case, the Marmion batholith and most dikes are unconformably overlain by the Steep Rock assemblage. The basal conglomerate of the Steep Rock assemblage contains mainly 3.0 Ga detritus probably eroded from the Marmion batholith, but also includes a 2779 Ma zircon grain (*see* Table 2, Nos. 47, 50 and 52). The young zircon grain constrains the deposition of the Steep Rock assemblage and development of the iron ore deposits to have occurred after 2779 Ma.

The Witch Bay assemblage is composed of pillowed to massive mafic metavolcanic flows and associated gabbro sills and intermediate metavolcanic fragmental rocks. Mafic metavolcanic rocks of the Witch Bay assemblage have a mineral assemblage of actinolite + zoisite + plagioclase and are characterized by a pale green colour. This, together with a low level of strain (except at margins of units and in the Quetico fault) makes mafic rocks of the Witch Bay assemblage distinct from other mafic units in the Steep Rock belt.

The Witch Bay assemblage attains its greatest extent in the area of the west arm of Steep Rock Lake where a 5 km thick section of east-facing pillow flows is probably capped to the east by a 1 km thick unit of intermediate metavolcanic fragmental rocks (*see* Figure 8). Other dominantly mafic metavolcanic units of the Witch Bay assemblage occur in the eastern lobe of the Steep Rock belt, at scattered localities in the western Steep Rock belt and as a narrow unit extending 30 km northeasterly along the central Finlayson Lake greenstone belt. Contacts of the scattered units of Witch Bay assemblage are typically schistose and can represent deformed thrust faults. Accordingly, the units of Witch Bay assemblage are probably klippe that have been down-folded into the basement of older assemblages.

A leucogabbro sill within the Witch Bay assemblage at the Atikokan airport has an age of 2735 Ma and pillowed mafic metavolcanic sequences west of Steep Rock Lake are cut by a 2730 Ma tonalite dike (*see* Table 2, Nos. 65 and 66). Witch Bay metavolcanic flows have compositions of high-iron to high-magnesium basalt with a lesser component of calc-alkalic basalt, andesite and dacite (Figure 6p). Trace element profiles of Witch Bay assemblage lavas are mainly flat although 2 subgroups of mildly enriched (Figures 7r and 7s) and depleted basalts (Figures 7t and 7u) are defined. Two small units of komatiite with overall flat but irregular trace element profiles (*see* Figure 7t) are identified west of Steep Rock Lake. The Witch Bay assemblage is a few million years older and lacks strongly enriched basalts, but is otherwise very similar to the Puffy Pillow assemblage of the Lac des Mille Lacs belt.

The Dismal assemblage extends discontinuously around the northeast margin of the Steep Rock belt where it attains a width of up to 2 km and can be traced in outcrop eastward along the boundary of the Wabigoon and Quetico subprovinces to Sapawe. The Dismal assemblage includes the Dismal Ashrock Formation (upper formation of the Steep Rock Group) dominantly pillowed mafic lavas and associated gabbro and melagabbro. The ashrock is a dull, dense dark soft ultramafic lapilli tuff characterized by a mineral assemblage of ferrostilpnomelane + chlorite + magnetite + calcite + tremolite (Shklanka 1972). Mafic metavolcanic flows of the Dismal assemblage are well developed in the Strawhat Lake area and are dark green with accessory magnetite and typically display small pillows. Jolliffe (1955) observed pillow lavas and large pyritic lenses, in the ashrock. Locally in the eastern Steep Rock belt, cherty banded iron formation is identified at close proximity to outcrops of ashrock.

Sample of ashrock contain inherited zircon grains ranging in age from 2997 to 2779 Ma (*see* Table 2, Nos. 49 and 51) indicating that ashrock magmas interacted with older crustal material representative of the Marmion and Dog domains and were emplaced at some time after 2779 Ma. In the Steep Rock Lake greenstone belt, the ashrock of the Dismal assemblage occurs east of a thick east-facing section of pillow lavas of the Witch Bay assemblage and, hence, is probably situated stratigraphically above the Witch Bay assemblage. This stratigraphic relation is supported by dikes of melagabbro, which are correlated with the Dismal assemblage and cut metavolcanic rocks of the Witch Bay assemblage in the eastern Steep Rock belt (Stone 2008a). The dikes provide an indication that the Dismal assemblage is younger than 2735 Ma.

Samples of the ashrock exhibit compositions ranging from komatiite through basaltic komatiite (*see* Figure 6q), whereas the associated mafic flows and intrusions of the Dismal assemblage are high-iron tholeiites. Although irregular, the trace element profiles of the ashrock are typically sloped to the right due to enrichment in incompatible elements, particularly LREE with Th and Nb < La (Figure 7v). These trace element characteristics are broadly comparable to those of samples of the Dismal assemblage in the Lac des Mille Lacs greenstone belt (*see* Figure 7p) and the Grassy Portage ultramafic unit in the Mine Centre belt (*see* Figure 7i). Tomlinson et al. (1999) and Goldstein and Francis (2008) studied the chemistry of ashrock from the Steep Rock Mine area and noted characteristically low-Al, high Ti and high Fe in samples. Goldstein and Francis (2008) found that the high Fe and enriched LREEs are unusual among Archean ultramafic rocks and proposed that these rocks should be called ferropicrite rather than komatiite.

Quetico Metasedimentary Belt

The Quetico metasedimentary belt ("Quetico belt")represents a narrow strip of predominantly metasedimentary rock extending easterly along the south side of the central Wabigoon Subprovince area (*see* Figure 3; *see* Map P.2229). These metasedimentary sequences are part of the Quetico Subprovince, which is traced eastward for nearly 1000 km from Minnesota through the present area to Quebec (*see* Figure 1; Williams 1991).

The metasedimentary rocks of the Quetico belt are overwhelmingly composed of grey to brown, centimetre- to decimetre-bedded wacke and siltstone. Sedimentary beds typically are made up of medium- to coarse-grained (0.5 to 2 mm grain size) massive largely feldspathic material. Graded beds are observed where the massive brown sandy basal material grades upward to fine-grained variably laminated to foliated dark silt at the tops of beds. The sedimentary bedding strikes mainly east with subvertical dip and the grain-size gradation indicates a northerly direction of younging at most localities.

Conglomerate and iron formation are minor components of the Quetico belt. Beds of conglomerate comprising rounded cobbles of plutonic, volcanic and sedimentary material are observed at a few localities along Highway 11 south of Atikokan. The conglomerate is intimately interbedded with wacke and appears to represent part of the Quetico metasedimentary sequence rather than conglomerate of the Seine sequence (discussed below).

Iron formation comprising thin layers of black granular quartz-magnetite interbedded with wacke occurs along the northeast-trending, northern boundary of the Quetico belt in the area of Wakinoo Lake (*see* Map P.2229). Although the beds are only a few centimetres in width, they extend continuously along most of the east-northeast-trending segment of the Quetico Subprovince boundary adjacent to the Shelby fault and are recognized as magnetic "highs" on aeromagnetic maps (Ontario Geological Survey 2002).

Quetico metasedimentary rocks are in contact with metavolcanic rocks of the Mine Centre, Steep Rock Lake and Lac des Mille Lacs greenstone belts at the north margin of the Quetico Subprovince. The sharp east- to east-northeast-trending contact is sheared by the Quetico and Shelby Lake faults at most localities. Wacke–siltstone sequences are strongly foliated within the faults; mylonites are developed where coarse granular rocks such as peraluminous granite dikes are present.

Metamorphic facies increases progressively southward within the Quetico Subprovince. Whereas a mineral assemblage of chlorite + plagioclase + quartz is common in wacke at the north boundary, more-or-less east-trending mineral isograds marking the first appearance of biotite, amphibole, garnet and staurolite are mapped at 1 to 2 km intervals to the south by Stone and Kamineni (1989). South of the present area, the Quetico metasedimentary sequences are migmatized. Kamineni, Stone and Johnson (1991) obtained a pressure/temperature estimate of 3.5 kilobars/550°C for the garnet/staurolite zone using the quartz-garnet-plagioclase-aluminosilicate geobarometer and garnet-biotite geothermometer.

Extensive geochronological analyses have been done on Quetico metasedimentary rocks in the Mine Centre–Atikokan areas (*see* Map P.2229 for locations of geochronology samples). Davis, Poulsen and Kamo (1989) and Davis, Pezzutto and Ojakangas (1990) obtained detrital zircon grains ranging in age from 3059 to 2698 Ma from the metasediments at the north margin of the Quetico Subprovince (*see* Table 2). These data are interpreted to indicate that detritus of quite variable age was derived from the Marmion terrane and deposited in the adjacent Quetico basin shortly after 2698 Ma. After deposition, the Quetico sedimentary rocks were intruded by gneissic biotite granodiorite, biotite granite, peraluminous granite, hornblende granodiorite (sanukitoid suite) and gabbro stocks of the Lac des Iles suite from 2688 to 2653 Ma (*see* Map P.2229; Williams 1991). Large peraluminous granite plutons are developed south of the present area at the central axis of the Quetico Subprovince (Percival 1989).

Chemical analyses of Quetico metasediments are provided by Sawyer (1986), Kamineni, Stone and Johnson (1991) and Stone, Kamineni and Jackson (1992). These have been interpreted by Kamineni, Stone and Johnson (1991) to indicate a mixed protolith of largely tonalitic and felsic volcanic material for the wacke augmented by mafic igneous rocks with elevated Ca, Cr, V and Ni for certain Ca-amphibole-bearing beds.

Seine Metasedimentary Sequence

Metasedimentary rocks of the Seine sequence occur in a series of units scattered along the north margin of the Quetico Subprovince. The largest unit occupies the Shoal Basin, which represents a wedge between the Quetico and Seine River faults southeast of Mine Centre (*see* Map P.2229). The Seine sequence also occurs as irregular units in the eastern Steep Rock Lake greenstone belt and as small units throughout the Marmion terrane.

Metasedimentary rocks of the Seine sequence include clast-supported conglomerate, interbedded conglomerate and sandstone, sandstone and minor siltstone. The conglomerate unconformably overlies wacke and siltstone of the Quetico sequence as well as metavolcanic rocks of the Mine Centre and Steep Rock Lake greenstone belts. Many outcrops of conglomerate lack bedding, whereas finer clastic metasedimentary components including sandstone and conglomerate-sandstone sequences are well

Table 5. Field and age characteristics of plutonic suites, central Wabigoon Subprovince area.

Suite (Map Unit)	Rock Type	Colour	Grain Size	Fabric	Form and Occurrence (% of area[1])	Inclusion Type	Mineral Assemblage[2]	Age (Ma)[3]
Biotite tonalite (12)	Biotite tonalite to granodiorite	White to grey	Fine to coarse	Foliated to weakly gneissic; weakly quartz and feldspar megacrystic	Irregular to crescentic and lobate bodies; scattered (30%)	Amphibolite, supracrustal xenoliths	pl + qtz + bt + Kfeld + mag ± ttn ± ep ± ap ± aln ± ilm ± zrn	2994, 2952, 2936, 2929, 2924, 2917, 2866, 2817, 2814, 2773, 2725, 2722, 2719,2717, 2716, 2709, 2697, 2688
Tonalite gneiss (11)	Biotite + hornblende tonalite to granodiorite	Dark grey to white	Variable between layers	Foliated; layered; folded	Belts, masses; scattered in and near biotite tonalite (8%)	Supracrustal xenoliths	pl + qtz + hbl + bt + Kfeld ± mag ± ep ± ttn ± ap ± aln ± ilm ± zrn	3009, 2957, 2952, 2937, 2928, 2924, 2887, 2880, 2777, 2775, 2729, 2732, 2713, 2673
Hornblende (16)	Hornblende + biotite tonalite to granite	Grey to white and pink	Coarse	Foliated; granular; feldspar megacrystic	Irregular to elongate bodies of variable size; scattered (4%)	Lensoid dioritic inclusions	pl + bt + hbl + qtz + Kfeld + mag ± ep ± ttn ± ap ± aln ± ilm ± zrn	3002, 2999, 2883, 2869, 2781, 2728, 2721
Biotite granite (15)	Biotite granodiorite to granite	White to pink	Medium to coarse	Massive to weakly foliated	Dikes, irregular masses, oval batholiths; scattered (28%)	Biotite tonalite, amphibolite	pl + qtz + Kfeld + bt + ttn ± mag ± ep ± ap ± ilm ± aln ± zrn	2686, 2686, 2685
Peraluminous (S-type) granite (13)	Biotite + muscovite granodiorite to granite	White	Coarse to pegmatitic	Massive; locally mylonitic	Elongate to irregular bodies (<1%)	sediment	pl + Kfeld + qtz + bt + ms + grt ± ap ± sil ± crd ± mnz	post 2695
Sanukitoid (14)	Biotite + hornblende + pyroxene quartz diorite, tonalite, quartz monzodiorite, granodiorite, quartz monzonite, quartz syenite, granite	Grey to pink and red	Medium to coarse	Massive to weakly foliated	Oval plutons (5%)	Hornblendite, amphibolite	pl + bt + hbl + cpx + Kfeld + qtz ± mag ± ttn ± ap ± ep ± ilm ± py ± zrn	2697, 2690, 2688, 2686, 2685, 2684

Notes: [1] greenstone and sedimentary belts comprise 16% of area;
[2] Kfeld = potassium feldspar; other mineral abbreviations *from* Kretz (1983): aln, allanite; ap, apatite; bt, biotite; cpx, clinopyroxene; crd, cordierite; ep, epidote; grt, garnet; hbl, hornblende; ilm, ilmenite; mag, magnetite; mnz, monazite; ms, muscovite; pl, plagioclase; py, pyrite; qtz, quartz; sil, sillimanite; ttn, titanite; zrn, zircon;
[3] for sources of age determinations, *see* references accompanying Table 2.

bedded and commonly show graded beds and cross beds. Wood (1980) noted a preponderance of intermediate volcanic clasts among a clast-population that includes mafic to felsic volcanic, tonalite and sedimentary material. Wood (1980) also noted that poorly sorted conglomerate at lower stratigraphic levels of the Shoal Basin is succeeded up-section by sorted conglomerate with sandstone interbeds, pebbly sandstone and trough cross-bedded sandstone.

The maximum age of deposition of the Seine sequence is constrained by geochronology. Davis, Poulsen and Kamo (1989) obtained an age of 2696 Ma for a tonalite clast in the Seine conglomerate of the Shoal Basin and samples from east of Legris Lake and in a small outlier north of Irene Lake produced populations of detrital zircon grains the youngest members of which have ages of 2700 and 2696 Ma (*see* Table 2, Nos. 104, 106 and 90). These data indicate that the Seine sequence was deposited after 2696 Ma and concur with crosscutting relations whereby the Seine sequence unconformably overlies most other supracrustal rocks.

Poulsen (1986) noted that units of Seine conglomerate are spatially associated with the Quetico fault and proposed that the sediments had been deposited in pull-apart basins developed along the late transcurrent fault. Scattered small units mapped far from the fault suggest however that the Seine sequence could have been regionally extensive and is perhaps best preserved where it has been deeply down-folded into basement rocks at localities such as along the Quetico fault.

Intrusive Rocks

Intrusive rocks are broadly subdivided into felsic to intermediate and mafic varieties. Felsic to intermediate intrusive rocks are, in turn, subdivided into 6 plutonic suites following Stone (1998, 2005b). Mafic intrusive rocks include Archean gabbro associated with various volcanic assemblages, distinct mafic plutonic suites such as the Lac des Iles suite, Proterozoic dikes and the Nipigon Sill Complex.

The field and age characteristics of the 6 suites of felsic to intermediate plutonic rocks are summarized in Table 5 and average chemical analyses of samples from each suite are provided in Table 6. The characteristics and geochemistry of the suites are described below.

BIOTITE TONALITE SUITE

The biotite tonalite suite represents the most common type of rock underlying 30% of the central Wabigoon Subprovince area. White to grey, generally medium grained and variably massive to foliated and weakly gneissic biotite tonalite to granodiorite is the principal type of rock within this suite. Bodies of the biotite tonalite suite occur in a wide range of sizes and shapes including small dikes and inclusions in other types of rock through irregular to crescentic masses to large oval batholiths such as the Marmion batholith (*see* Map P.2229). The biotite tonalite suite is gradational to the tonalite gneiss suite largely through progressive development of a gneissic texture and is otherwise cut by intrusions representative of most other felsic plutonic suites.

The biotite tonalite suite contains an average of 11% mafic minerals including mainly biotite with accessory magnetite, titanite, ilmenite and zircon (Figure 9a; *see* Table 5). Amphibole occurs rarely within biotite tonalite typically in association with amphibolite inclusions or with the more mafic phases of tonalite. Within local zones of altered biotite tonalite, the biotite is chloritized and feldspars are altered to epidote.

Intrusions of the biotite tonalite suite show considerable variation in age (2994 to 2688 Ma; *see* Tables 2 and 5) and occur in most age domains of the central Wabigoon Subprovince area (*see* Figure 3). Geochemically, samples of the biotite tonalite suite are calc-alkalic, of the low- to medium-potassium type of LeMaitre (1989; Figure 10a), mildly metaluminous to peraluminous (ACNK = 0.9 to 1.1; Table 7)

Table 6. Average chemical analyses of plutonic suites in the central Wabigoon Subprovince area.

Suite	Gneissic	Biotite Tonalite (Al-rich)	Biotite Tonalite (Al-poor)	Peraluminous S-Type Granite	Sanukitoid	Biotite Granite	Hornblende Tonalite
No. samples	**11**	**49**	**7**	**6**	**39**	**51**	**16**
SiO_2	67.85	68.52	73.08	71.11	57.96	70.72	63.06
TiO_2	0.36	0.39	0.18	0.06	0.62	0.22	0.57
Al_2O_3	15.95	15.91	13.97	14.78	15.99	15.12	16.05
$Fe_2O_3^T$	3.20	3.30	1.77	0.84	6.34	1.79	5.20
FeO^T	2.75	2.97	1.60	0.75	5.73	1.61	4.68
Fe_2O_3	1.17	1.20	0.74	0.31	2.45	0.78	2.13
MnO	0.04	0.04	0.03	0.07	0.09	0.03	0.08
MgO	1.19	1.01	0.38	0.17	4.06	0.57	2.24
CaO	3.80	3.61	2.20	0.71	5.76	2.04	4.79
Na_2O	4.85	4.82	4.41	4.84	4.16	4.65	4.29
K_2O	1.23	1.27	2.35	4.47	3.09	3.42	2.05
P_2O_5	0.08	0.11	0.06	0.05	0.31	0.08	0.15
LOI	0.74	0.78	0.60	0.59	1.05	0.69	0.79
Total	99.32	99.76	99.00	97.59	99.42	99.27	99.27
CO_2	0.13	0.12	0.12	0.11	0.18	0.10	0.10
S	0.03			0.01	0.11	0.03	0.03
FeO	1.85	2.02	0.91	0.47	3.53	0.95	2.41
H_2O_m	0.16	0.16	0.15	0.12	0.20	0.16	0.18
H_2O_p	0.50	0.56	0.35	0.30	0.91	0.34	0.66
Mg #	42	38	30	27	54	36	46
ACNK	1.0	1.0	1.0	1.1	0.8	1.0	0.9
Li	41	34	41	34	28	26	30
Be	1	1		1	1	1	1
K	5124	5271	9754	18561	12831	14177	8525
Rb	46	45	93	135	84	116	68
Ba	529	407	677	482	1161	763	722
Sr	433	371	241	143	1081	457	666
Cs	2.1	1.2	2.2	4.5	2.6	2.4	2.1
Sc	5	4	2	2	11	3	10
Ti	2167	2336	1071	360	3700	1335	3436
Y	6	7	8	21	15	7	14
Zr	126	147	113	40	131	136	148
Nb	4.8	5.6	6.5	5.7	7.3	5.9	7.9
Hf	2.8	3.4	3.5	2.1	3.0	3.8	3.3
Ta	0.3	0.5	1.0	0.8	0.5	0.5	0.5
Mo	16	14	18	23	7	9	11
W	9			4			
Ga	19	19	16	20	19	18	20
Sn		6		10	17	8	
Co	9	10		3	23	8	17
Cr	27	23	22	14	155	20	64
Cu	16	17	10	10	42	12	36
Ni	16	11	7	5	69	13	35
V	43	40	14	5	108	21	88
Zn	61	64	45	28	79	46	72
Pb	9	9	15	29	15	20	13
Th	5.2	4.1	10.1	3.4	6.2	13.8	5.0
U	0.9	0.7	2.4	4.3	1.2	1.7	1.0
P	185	233	127	109	687	173	332

Table 6. continued.

Suite	Gneissic	Biotite Tonalite (Al-rich)	Biotite Tonalite (Al-poor)	Peraluminous	Sanukitoid	Biotite Granite	Hornblende Tonalite
No. samples	**11**	**49**	**7**	**6**	**39**	**51**	**16**
La	25.1	22.1	25.2	4.8	47.2	29.3	30.5
Ce	45.6	42.4	48.6	10.6	105.4	56.9	66.2
Pr	5.0	4.5	5.2	1.3	12.9	6.0	7.6
Nd	16.7	15.4	17.1	4.7	50.3	20.1	28.5
Sm	2.5	2.5	2.7	1.3	8.5	3.1	4.8
Eu	0.8	0.8	0.6	0.2	2.2	0.7	1.2
Gd	1.9	1.9	2.2	1.6	6.0	2.1	3.5
Tb	0.2	0.3	0.3	0.4	0.7	0.3	0.5
Dy	1.1	1.3	1.6	3.0	3.2	1.3	2.5
Ho	0.2	0.2	0.3	0.7	0.6	0.2	0.5
Er	0.5	0.7	0.8	2.0	1.4	0.7	1.4
Tm	0.1	0.1	0.1	0.3	0.2	0.1	0.2
Yb	0.5	0.6	0.8	2.0	1.2	0.6	1.3
Lu	0.1	0.1	0.1	0.3	0.2	0.1	0.2
$(La/Yb)_N$	46.4	34.9	38.6	4.9	28.4	41.6	25.8
Eu/Eu*	1.2	1.3	1.0	0.5	0.9	1.0	0.9

and have 54 to 74% SiO_2 with magnesium numbers (Mg#) in the range of 27 to 45. In comparison with other plutonic suites the biotite tonalite suite has low abundances of large ion lithophile elements (LILE) and moderately fractionated rare earth elements (REEs) with negligible to weakly positive or negative Eu anomalies. Rare earth element profiles (Figure 11a) are steeply sloped through light rare earth elements (LREEs: La to Sm) and weakly sloped to the right through heavy rare earth elements (HREEs: Gd to Lu). Where normalized to primitive mantle, trace element profiles are sloped from incompatible to compatible elements with deep troughs for Nb and Ti (Figure 12a). The profiles of individual samples tend to criss-cross each other for incompatible elements Cs through Zr suggesting that these elements could be affected by alteration.

Early studies (e.g., Barker and Arth 1976) subdivided rocks herein assigned to the biotite tonalite suite into high-Al (Al_2O_3 > 15 weight %) and low-Al varieties (Al_2O_3 < 15 weight %). High-Al tonalites are characterized by moderately fractionated REEs with low HREEs and negligible Eu anomalies whereas low-Al tonalites have high abundances of HFSE (especially Zr and Nb) and REE profiles have low La/Yb_N with negative Eu anomalies (Beakhouse 2007). Samples from the central Wabigoon Subprovince are of the high-Al type and, although a few have low-Al_2O_3, these lack many geochemical characteristics such as abundant Zr and Nb and REEs with low La/Yb_N and negative Eu anomalies for published low-Al tonalites (*see* Table 7; *see* Figure 11a). Accordingly, "true" low-Al tonalites appear to be rare in the central Wabigoon Subprovince area and those samples with Al_2O_3 somewhat less than 15 weight % are probably members of the high-Al suite.

Magmas of the high-Al tonalite-trondhjemite-granodiorite (TTG) suite, which is probably equivalent to the biotite tonalite suite of this study, have been historically interpreted as the products of partial melting of basalt at lower crustal depths (Arth and Hanson 1975; Arth and Barker 1976). A residue of garnet and/or amphibole possibly accounts for HREE-depletion in tonalite. Drummond and Defant (1990) subsequently placed the petrogenesis of tonalitic magmas in a plate-tectonic setting by proposing that they originated through partial melting of subducted oceanic crust or slab-melting. Recently, the origin of tonalite by slab-melting has been questioned by Smithies (2000) who noted that Archean tonalites lack the high Mg numbers and lower SiO_2 of Cenozoic adakites, which have originated by slab-melting and interaction with the mantle. Smithies and Champion (2000) went on to propose that since Archean tonalites show no clear evidence of having interacted with a mantle wedge, they likely represent partial melts of basaltic lower crust rather than of subducted oceanic crust.

Table 7. Geochemical characteristics of plutonic suites, central Wabigoon Subprovince area.

Suite (Map Unit)	Rock Type		SiO_2 min–max (average) (wt %)	Al_2O_3 min–max (average) (wt %)	K_2O min–max (average) (wt %)	Mg# [1] min–max (average)	ACNK [2] min–max (average)	LILE Rb (ppm)	LILE (averages) Ba (ppm)	LILE Sr (ppm)	HFSE Y (ppm)	HFSE (averages) Zr (ppm)	HFSE Nb (ppm)	$(La/Yb)_N$ [3] min–max (average)	Eu/Eu* [4] min–max (average)	Age [5] (Ma)
Biotite tonalite (12)	biotite tonalite to granodiorite	High-Al	53.5–74.1 (68.5)	14.6–19.7 (15.9)	0.5–2.0 (1.3)	28.9–44.5 (37.5)	0.9–1.1 (1.0)	45	407	371	7	147	5.6	3–155 (35)	0.6–8.7 (1.3)	2994, 2952, 2936, 2929, 2924, 2917, 2866, 2817, 2814, 2773, 2725, 2722, 2719, 2717, 2716, 2709, 2697, 2688
		Low-Al	71.4–74.2 (73.1)	13.3–14.5 (14.0)	0.9–4.9 (2.4)	26.5–36.5 (30.0)	1.0–1.0 (1.0)	93	677	241	8	113	6.5	11–129 (39)	0.5–1.4 (1.0)	
Tonalite gneiss (11)	biotite + hornblende tonalite to granodiorite		60.5–71.3 (67.9)	14.8–18.1 (16.0)	0.6–2.3 (1.2)	29.6–55.2 (42.2)	0.9–1.0 (1.0)	46	529	433	6	126	4.8	11–155 (46)	0.7–2.5 (1.2)	3009, 2957, 2952, 2937, 2928, 2924, 2886, 2880, 2777, 2775, 2729, 2732, 2713, 2673
Hornblende (16)	hornblende + biotite tonalite to granite		51.7–71.4 (63.1)	14.6–18.4 (16.1)	0.5–4.1 (2.1)	30.5–57.8 (45.9)	0.7–1.1 (0.9)	68	722	666	14	148	7.9	2–74 (26)	0.6–1.2 (0.9)	3002, 2999, 2883, 2869, 2781, 2728, 2721
Biotite granite (15)	biotite granodiorite to granite		60.1–76.3 (70.7)	12.7–17.8 (15.1)	0.6–6.7 (3.4)	3.7–52.2 (35.7)	0.9–1.09 (1.0)	116	763	457	7	136	5.9	2–136 (42)	0.1–2.8 (1.0)	2686, 2686, 2685
Peraluminous (S-type) granite (13)	biotite + muscovite granodiorite to granite		63.5–75.4 (71.1)	12.9–19.0 (14.8)	2.9–7.0 (4.5)	11.7–39.9 (27.2)	0.9–1.2 (1.1)	135	482	143	21	40	5.7	0–9 (5)	0.06–0.88 (0.45)	post 2698
Sanukitoid (14)	biotite + hornblende + pyroxene quartz diorite, tonalite, quartz monzodiorite, granodiorite, quartz monzonite, quartz syenite, granite		44.9–68.3 (58.0)	12.9–22.9 (16.0)	0.8–7.7 (3.1)	35.0–71.5 (54.0)	0.6–1.0 (0.8)	84	1161	1081	15	131	7.3	5–51 (28)	0.65–2.04 (0.9)	2697, 2690, 2688, 2686, 2685, 2684

[1] Mg# = (MgO/molecular weight MgO)/((MgO/molecular weight MgO)+(FeO total/molecular weight FeO))* 100
[2] ACNK = (Al_2O_3/molecular weight Al_2O_3)/((CaO/molecular weight CaO)+(Na_2O/molecular weight Na_2O)+(K_2O/molecular weight K_2O))
[3] $(La/Yb)_N$ = (La(ppm)/chondrite normalizing value for La)/(Yb(ppm)/chondrite normalizing value for Yb)
[4] Eu/Eu* = (Eu (ppm)/chondrite normalizing value for Eu)/((Sm (ppm)/chondrite normalizing value for Sm)+(Gd (ppm)/chondrite normalizing value for Gd))/2
[5] sources of age determinations are references accompanying Table 2

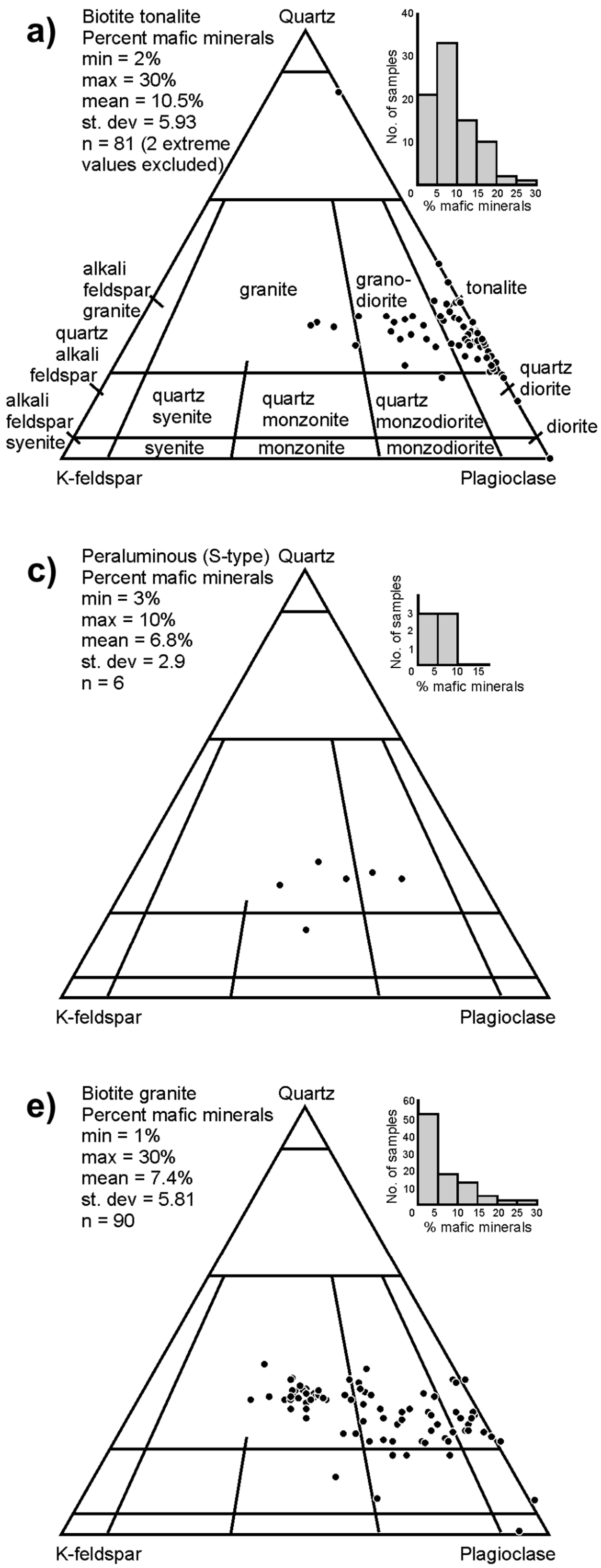

Figure 9. Quartz–alkali feldspar–plagioclase diagram for intermediate to felsic plutonic suites.

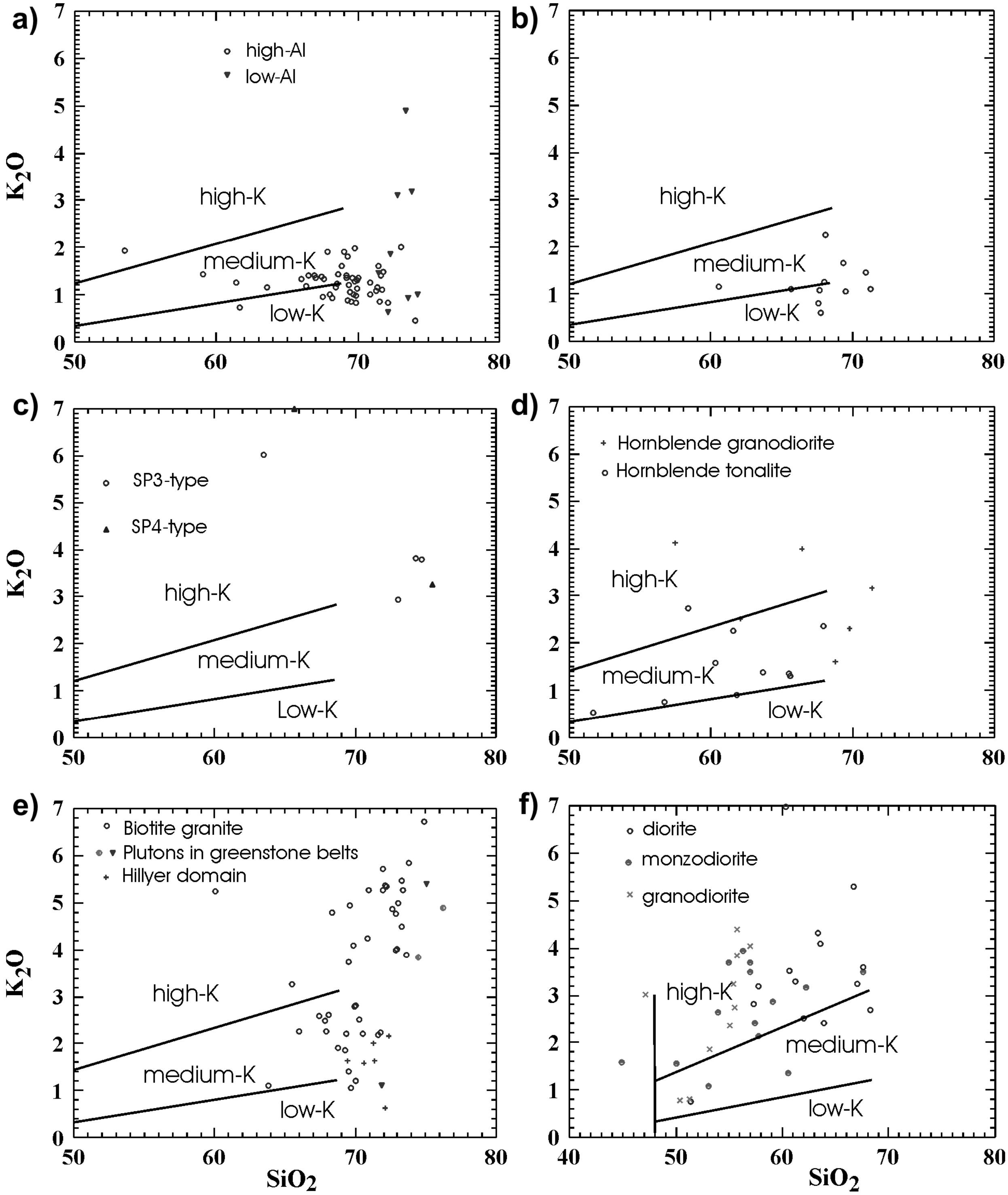

Figure 10. SiO_2 versus K_2O plots (Le Maitre 1989) for plutonic suites: a) biotite tonalite, b) tonalite gneiss, c) peraluminous granite, d) hornblende tonalite, e) biotite granite and f) sanukitoid.

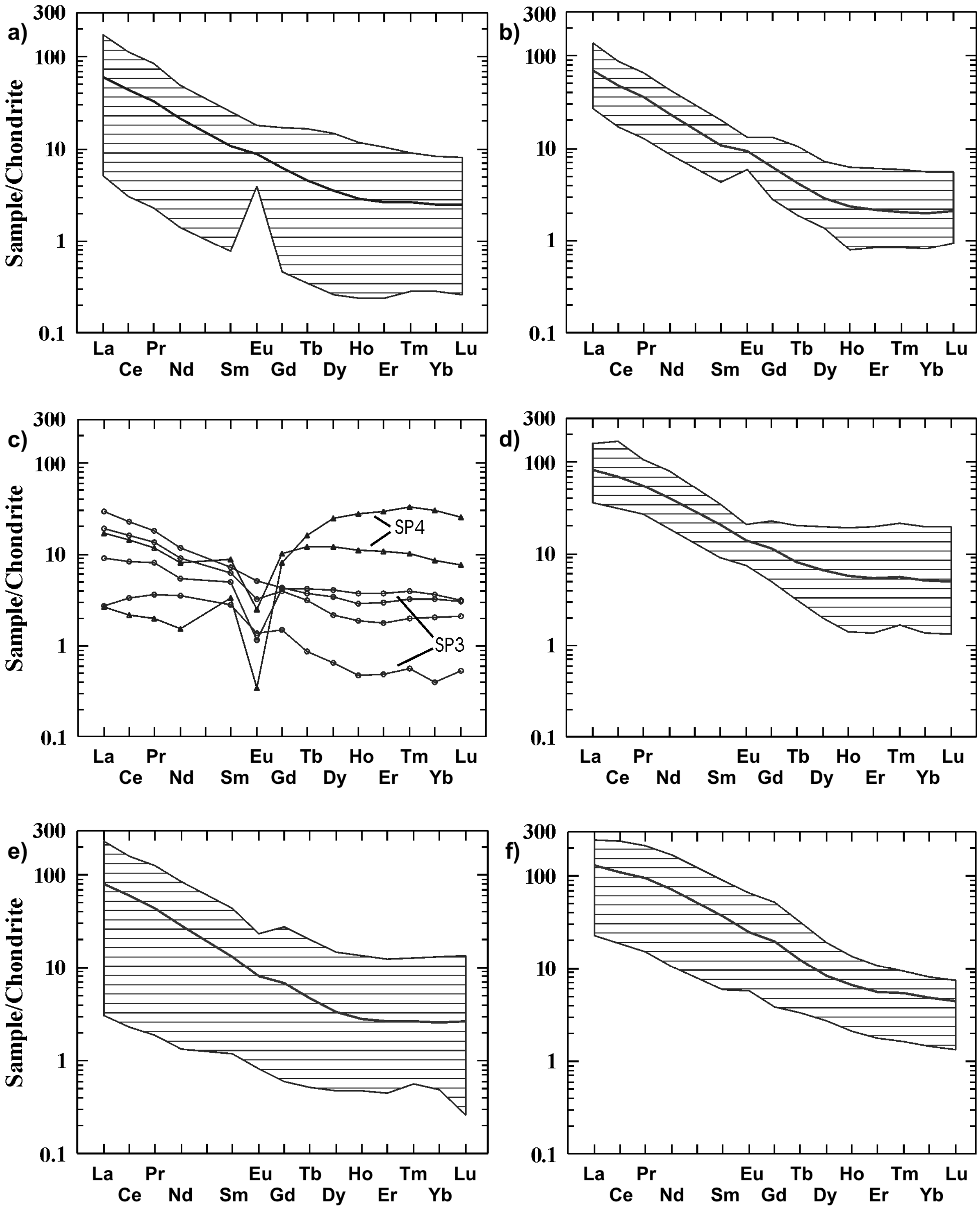

Figure 11. Rare earth element plots for plutonic suites for a) biotite tonalite, b) tonalite gneiss, c) peraluminous granite, d) hornblende tonalite, e) biotite granite and f) sanukitoid; solid line represents average.

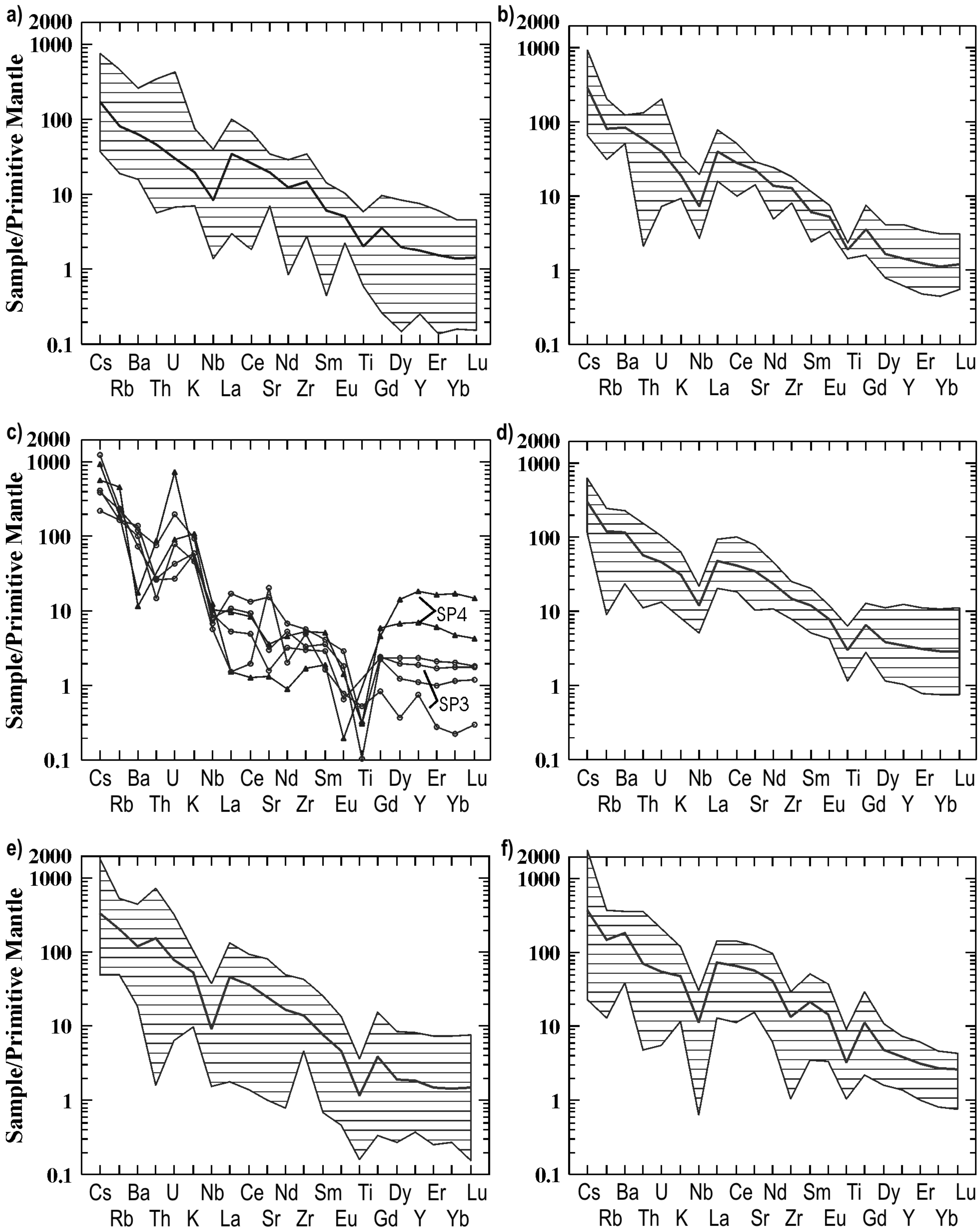

Figure 12. Primitive mantle-normalized trace element plots for plutonic suites. Solid line represents average for a) biotite tonalite, b) tonalite gneiss, c) peraluminous granite, d) hornblende tonalite, e) biotite granite and f) sanukitoid.

Several lines of evidence support the interpretation of Smithies and Champion (2000) for the origin of biotite tonalite from the central Wabigoon Subprovince area. Firstly, geologic mapping shows a lack of ultramafic inclusions within this suite such as are widely found in intrusions of the mantle-derived sanukitoid suite. Secondly, the biotite tonalite suite lacks high Mg numbers and high concentrations of transition metals (Co, Cr, Cu, Ni, V and Zn) typical of the sanukitoid suite (*see* Tables 6 and 7). The geologic and geochemical evidence implies that although the biotite tonalite suite has originated probably by partial melting of basaltic material at lower crustal depths, the magmas have not passed through or interacted significantly with the ultramafic mantle.

TONALITE GNEISS SUITE

The tonalite gneiss or gneissic suite is texturally and compositionally heterogeneous comprising 8% of the central Wabigoon Subprovince area. Tonalite gneisses are layered with individual layers varying compositionally from leucocratic tonalite and granodiorite through mesocratic tonalite and granodiorite to diorite and amphibolite. The layers are typically several centimetres in thickness and may be continuous on the scale of an outcrop or gradually dissipate over distances of a few metres. The gneisses can show a wide variety of textures including boudinage of certain stiff layers, folding and development of strongly foliated to mylonitic zones. Gneisses are cut by variably deformed dikes of tonalite, granodiorite and amphibolite and may contain xenoliths of amphibolite representing either deformed dikes or remnants of volcanic rocks. Locally, mafic tonalite gneisses grade to amphibole gneisses of volcanic origin or migmatites of sedimentary origin. The more felsic tonalite gneisses are gradational to, and spatially associated with, biotite tonalite and are cut by dikes of most other plutonic suites. Units of gneisses are irregular to belt-like in shape (*see* Map P.2229) and are concentrated in the boundary zone between the western and central Wabigoon terranes.

Gneisses are composed of variable amounts of plagioclase, quartz and potassium feldspar such that they range compositionally from tonalite through granodiorite to granite (Figure 9b). The proportion of combined hornblende and biotite is 19% on average, but this varies substantially between layers as well as various outcrops of gneisses. Although some mafic gneisses are diorite to quartz diorite (*see* Figure 9b) in composition, many of the mafic gneisses appear to be amphibolite of probable volcanic origin. Accessory minerals include magnetite, titanite, apatite, allanite, ilmenite and zircon.

Like biotite tonalite, the gneisses range substantially in age from 3009 to 2673 Ma (*see* Table 5) and appear to have developed episodically through the Meso- to Neoarchean. Chemically, the gneisses are comparable to the more mafic varieties of the biotite tonalite suite. Tonalite gneisses are calc-alkaline, of the low- to medium-potassium variety (Figure 10b) and are mildly metaluminous to peraluminous (ACNK = 0.9 to 1.0; *see* Table 7). Gneisses have 61 to 71 weight % SiO_2, greater than 14.4 weight % Al_2O_3 and low levels of K_2O (0.6 to 2.3 weight %; *see* Table 7). Gneisses are distinguished from the average biotite tonalite by somewhat higher Mg numbers and concentrations of transition metals. These chemical distinctions reflect that gneisses have, on average, more mafic minerals than biotite tonalite.

The LREE profiles of tonalite gneisses are steeply sloped, whereas HREE profiles are weakly sloped to flat and variably depleted with an overall average $(La/Yb)_N$ of 46. Eu anomalies are negligible to weakly positive or negative (Figure 11b). Where normalized to primitive mantle, the trace element profiles of gneisses are steeply sloped from incompatible to compatible elements with deep troughs for Nb and Ti (Figure 12b). The more incompatible elements, particularly Th and U are variable, perhaps due to alteration.

The origin of gneisses appears to be complex and at least partly tectonic. On the one hand, geochemical evidence including REE profiles suggest that tonalite gneisses have originated in a similar manner to high-Al biotite tonalite and, hence, the bulk of tonalitic material within gneisses is probably derived by partial melting of basalt at lower crustal depths. On the other hand, gneisses are at least locally gradational to supracrustal rocks including amphibole gneisses of probable mafic volcanic origin, biotite gneisses of sedimentary origin and quartzofeldspathic, layered gneisses of obscure but probably sedimentary or mixed volcano-sedimentary origin. Gneisses commonly show strongly foliated to mylonitic textures and belts of gneisses are spatially associated with zones of high strain such as the margins of large batholiths and tectonic boundary zones. These observations imply that gneisses are a mixture of high-Al tonalite and a smaller proportion of supracrustal rocks. The high-Al tonalite evidently intruded and was tectonically interlayered with supracrustal rocks in zones of high strain.

PERALUMINOUS (S-TYPE) GRANITE SUITE

The peraluminous or S-type granite suite is a minor component of the central Wabigoon Subprovince area occurring in association with wacke–siltstone sedimentary sequences and greenstone belts. Although comprising only 1% of the area, peraluminous granite is extensive southward at the central axis of the Quetico Subprovince (e.g., Percival 1989).

Peraluminous granite occurs in a wide range of forms varying from large oval to elongated batholiths south of the present area to thin dike-like concordant bands alternating with wacke in metasedimentary units. The latter alternating "*lit-par-lit*" arrangement of peraluminous granite neosome and wacke paleosome is characteristic of metasedimentary migmatites (e.g., Mehnert 1968). Large bodies of peraluminous granite typically contain inclusions of sedimentary rocks.

Peraluminous granite is coarse grained to pegmatitic, white in colour and can be variably massive or strongly foliated to mylonitic. It is made up of quartz, plagioclase and potassium feldspar with a few percent of biotite and/or muscovite and ranges compositionally from granodiorite through granite (Figure 9c). Accessory minerals are unique among the plutonic suites and include garnet, apatite, sillimanite, cordierite and monazite (*see* Table 5).

Although ages have not been determined for these rocks in the study area, peraluminous granite is interpreted to postdate the sedimentary rocks from which it is melted. Accordingly, peraluminous granite bodies within the Quetico Subprovince are younger than 2698 Ma in age (Davis, Pezzutto and Ojakangas 1990), although small bodies associated with sedimentary components of greenstone sequences can be different in age and probably older.

Peraluminous granite is calc-alkaline, of the medium- to high-potassium variety (Figure 10c) and mildly metaluminous to strongly peraluminous (ACNK = 0.9 to 1.2; *see* Table 7). In comparison with other plutonic suites, peraluminous granite has a low Mg#, low concentrations of transition metals and fairly high concentrations of LILE elements with the exception of Sr (*see* Table 7). Two principal types of peraluminous granite are distinguished in terms of rare earth elements including those that are mildly LREE-enriched and those that have higher Na and more-or-less flat chondrite-normalized REE profiles with deep Eu troughs (Figure 11c). Where normalized to primitive mantle, trace element profiles for peraluminous granite are irregular but steeply sloped for incompatible elements and more-or-less flat for compatible elements with deep Ti troughs (Figure 12c).

The strong spatial association with sedimentary migmatites provides compelling evidence that peraluminous granite originated by partial melting of sedimentary rocks. Certain geochemical characteristics such as depletion in Ca, Na and K relative to Al and low Sr are typical of the sedimentary source rocks from which these elements were dissolved by water (Chappell and White 1974; White et al. 1986). The 2 types of peraluminous granite (mildly LREE-enriched and more-or-less flat REE profiles) are representative of the SP3- and SP4-type peraluminous granite of Sylvester (1994). Although Sylvester (1994) attributed these types to have originated from partial melts of a wacke-pelite source and a sodiumrich source such as wacke+felsic volcanic rocks, field relations in the central Wabigoon Subprovince show that the SP3-type represents fairly normal peraluminous granite presumably derived by partial melting of sediments whereas the SP4-type samples are derived from late and highly evolved pegmatite dikes. Breaks and Moore (1992) noted SP4-type REE profiles for late pegmatites and suggested that these developed due to fractionation of monazite and apatite that caused LREE depletion and due to partitioning of divalent Eu into aqueous fluids that led to negative Eu anomalies.

HORNBLENDE TONALITE SUITE

Rocks of the hornblende tonalite suite represent a few percent of the central Wabigoon Subprovince area and occur in irregular to oval and highly elongate forms of variable size. Large examples of the latter include two intrusions at margins of the Marmion batholith (*see* Map P.2229).

Rocks of the hornblende tonalite suite range compositionally from tonalite through granodiorite to granite and also include significant proportions of quartz diorite and quartz monzodiorite (Figure 9d). Hornblende tonalite is typically a coarse-grained grey to white rock characterized by a granular texture due to large blocky feldspars in a darker matrix of quartz and mafic minerals. The rock is massive to weakly foliated and has distinct lensoid dioritic inclusions. Grey hornblende tonalite locally grades to pink hornblende granodiorite and granite through development of pink potassium feldspar phenocrysts. Mafic minerals include amphibole and biotite as well as accessory magnetite, titanite, apatite, allanite, ilmenite and zircon. Amphibole and biotite are more abundant in tonalitic than in granitic phases and overall comprise an average of 19% of the hornblende tonalite suite.

Like biotite tonalite, the hornblende tonalite suite shows considerable variation in age (3002 to 2721 Ma; *see* Table 5) and appears to have developed episodically through the Meso- and Neoarchean. The hornblende tonalite suite is calc-alkaline, of the medium- to high-potassium type (Figure 10d) and most samples are metaluminous (ACNK = 0.7 to 1.1; *see* Table 7). The hornblende tonalite suite has a fairly high Mg# of 46, high concentrations of transition metals and LILE as well as high field-strength elements including Zr and Nb (*see* Tables 6 and 7). REE profiles are steeply sloped for LREE and relatively flat for HREE with negligible Eu anomalies (Figure 11d). Primitive mantle-normalized trace elements are steeply sloped with deep troughs for Nb and Ti (Figure 12d). The profiles for individual samples cross each other for incompatible elements more so than for compatible elements.

The hornblende tonalite suite is not recognized by other workers who presumably include this rock with biotite tonalite of the TTG suite or with the sanukitoid suite. Still, hornblende tonalite to granodiorite and granite are common in the western Superior Province (Stone 1998, 2005b) and represent a unique suite recognizable in the field on the basis of its dominant mafic mineral, fabric and compositional variation.

The hornblende tonalite suite has chemical similarities to, as well as the dual combination of primitive and evolved compositional characteristics of the sanukitoid suite. For example, both suites have high Mg numbers and high concentrations of transition metals and yet also have high concentrations of LILE elements including potassium. Shirey and Hanson (1984) attributed the dual primitive-evolved character of monzodioritic sanukitoid plutons to melting of primitive ultramafic mantle that had been metasomatically enriched in LILE and fractionated REEs prior to melting. Stern and Hanson (1991) explained petrogenesis of the evolved granodiorite to granite cores of sanukitoid plutons by crystallization differentiation of the monzodioritic magmas. Possibly, magmas of the hornblende tonalite suite originated in a similar manner although several important differences are apparent.

Firstly, the hornblende tonalite suite lacks the extreme concentrations of transition metals and high Mg# as well as ultramafic inclusions of the sanukitoid suite. These allude to a mafic rather than ultramafic source from which hornblende tonalite magmas originated. Secondly, the concentrations of potassium, LILE and HFSE is more variable and somewhat lower in the hornblende tonalite suite than in the sanukitoid suite, which implies that enrichment in these elements was less efficient for the hornblende tonalite suite than for the sanukitoid suite. Thirdly, geochronology shows that hornblende tonalite magmas were generated at least episodically through the Meso- to Neoarchean, whereas sanukitoid magmas came only in the late Neoarchean (*see* Table 7) as melting switched from the subducted slab to the mantle wedge. These suggest that hornblende tonalite magmas could have been produced more-or-less continuously by ongoing subduction rather than in a discrete episode when subduction ceased due to convergence of plates as is though to be the case for sanukitoid magmas.

Hildreth and Moorbath (1988) proposed a mechanism of melting, assimilation, storage and homogenization (MASH) that may account for the genesis of hornblende tonalite magmas. These authors proposed that basaltic melts would be produced in the mantle wedge as oceanic crust was subducted beneath a continent. The basaltic melts would rise into the lower crust where they would stall due to lack of buoyancy and produce partial melts of lower crustal rocks. The basaltic and crustal melts would mix, homogenize and rise producing a spectrum of magma compositions ranging from diorite to hornblende granite depending on the relative proportions of the two parent magmas. This mechanism could have been active continuously or discontinuously in the Archean giving rise to diverse rocks of the hornblende tonalite suite.

BIOTITE GRANITE SUITE

Rocks of the biotite granite suite are voluminous underlying 28% of the central Wabigoon Subprovince area. These occur in forms ranging from small dikes and masses intruding other rocks to large oval batholiths. Examples of the large biotite granite intrusions include the White Otter, Muskeg and Cecil batholiths (*see* Map P.2229).

Rocks of the biotite granite suite are typically coarse grained and massive to weakly foliated and white to pink in colour. Although generally equigranular, potassium feldspar megacrystic varieties are common. The biotite granite suite ranges compositionally from granite through granodiorite to tonalite (Figure 9e) and, in many areas, is gradational to rocks of the biotite tonalite suite. The biotite granite suite is fairly leucocratic with an average of 7% biotite and accessory titanite, magnetite, epidote, apatite, ilmenite, allanite and zircon. Rocks of the biotite granite suite contain a wide variety of xenoliths representing many supracrustal and plutonic country rocks. Among the more widespread and common inclusions are those of partly assimilated biotite tonalite possibly representing the source rock from which biotite granite was melted. Also present are amphibolite inclusions of uncertain (volcanic or intrusive) origin.

Rocks of the biotite granite suite crosscut most other Archean lithologies. These crosscutting relations are confirmed by U/Pb geochronology that shows ages with a narrow range from 2685 to 2686 Ma characteristic of late, posttectonic plutons (*see* Table 5). The biotite granite suite is calc-alkaline, of the medium- to high-potassium type (Figure 10e) and mildly metaluminous to peraluminous (ACNK = 0.9 to 1.09; *see* Table 7). The biotite granite suite has generally low Mg# (4 to 52) and low concentrations of transition metals, whereas LILE particularly Rb and Ba and also Th are high (*see* Table 6). REEs are strongly LREE-enriched with weak positive or negative Eu anomalies (Figure 11e) and have considerable variation in the ΣREE. Samples with high ΣREE tend to have negative Eu anomalies whereas those with low ΣREE have positive Eu anomalies. Similar to most other plutonic suites, primitive mantle-normalized trace element profiles are steeply sloped from incompatible to compatible elements with deep troughs for Nb and Ti (Figure 12e).

Rocks of the biotite granite suite are widely recognized in Archean terranes and variously described as the crustal I-type suite (Beakhouse 2007) or granodiorite-granite-monzogranite (GGM) suite (Whalen et al. 2004). On the basis of field and laboratory studies, there is almost complete agreement that biotite granite originates by partial melting of tonalitic crust and associated sediments at mid to lower crustal levels (Beakhouse and McNutt 1991; de Wit 1998 and references therein). The widespread occurrence of variably digested inclusions of biotite tonalite within biotite granite of the central Wabigoon Subprovince area provides local field evidence in support of this interpretation. High SiO_2 and K_2O combined with high LILE and LREEs are geochemical indications of an evolved and largely crustal source for biotite granite magmas. Low Mg numbers and low transition metals are inconsistent with a mafic or metasomatized mantle source such as is interpreted for the biotite tonalite and sanukitoid suites.

Comparison of REE patterns provides supporting evidence that the biotite granite suite could have been derived by partial melting of the biotite tonalite suite. For example, melting experiments (e.g., Skjerlie and Johnson 1993) show that tonalite breaks down to a residue of orthopyroxene + two oxides + granitic melt at 6 kilobars and orthopyroxene + garnet + granitic melt at 10 kilobars. Both residual silicate minerals would tend to concentrate HREEs in the restite and reject LREEs into the melt (Rollinson 1993) and, indeed, the average biotite granite has HREEs about equal to those of biotite tonalite and LREEs somewhat higher (*see* Table 6). The experimental work supports the interpretation that the biotite granite suite could have originated at mid-crustal depths by partial melting of older crustal material.

SANUKITOID SUITE

The sanukitoid suite represents a range of quartz-undersaturated to saturated intermediate to felsic plutonic rocks with remarkable variation in composition. These include diorite to tonalite, monzodiorite to granodiorite and monzonite to granite as well as syenite (Figure 9f). Their compositional heterogeneity is brought about by highly variable proportions of quartz, plagioclase and potassium feldspar as well as mafic minerals including biotite, hornblende and clinopyroxene. Mafic minerals tend to be concentrated in quartz-undersaturated phases and comprise an average of 29% of the rock. Magnetite, titanite, apatite, epidote, ilmenite, pyrite and zircon constitute a few percent of sanukitoid rocks.

Intrusions of the sanukitoid suite are mainly oval to elongate plutons underlying 5% of the central Wabigoon Subprovince area. These show east-northeasterly alignment (*see* Map P.2229) and most are visible as distinct magnetic “highs” on aeromagnetic maps. Sanukitoid plutons tend to be zoned with dioritic to monzodioritic rims grading to granodioritic to granite cores. Whereas the rims of sanukitoid plutons are made up of grey medium-grained, foliated, quartz-undersaturated intermediate rock, the cores tend to be massive red coarse-grained to potassium-feldspar megacrystic granodiorite to granite. Ultramafic (typically hornblendite) inclusions occur in all phases of sanukitoid plutons, but are larger and more common in the primitive rim phases.

Sanukitoid plutons crosscut most other lithologies and show a narrow range of U/Pb ages from 2697 to 2684 Ma (*see* Table 5). The wide compositional spectrum of sanukitoid plutons is manifest in geochemical variation diagrams. Sanukitoid plutons are calc-alkaline and mainly of the high-potassium type (Figure 10f). Many are strongly enriched in K_2O+Na_2O to the extent that they overlap the fields of alkaline and subalkaline compositions for volcanic rocks (Irvine and Baragar 1971). Sanukitoid plutons are strongly metaluminous with ACNK values of 0.6 to 1.0 and have high Mg numbers of 35 to 72 (*see* Table 7). In comparison with other plutonic suites, sanukitoid plutons have high concentrations of transition metals, certain LILE elements such as Ba and Sr and elements including Ti and P (*see* Table 6). The REE profiles of sanukitoid rocks are LREE-enriched with $(La/Yb)_N$ ranging from 5 to 51 (Figure 11f) and have negligible Eu anomalies. Primitive mantle-normalized trace elements are strongly sloped from incompatible to compatible elements with deep troughs for Nb, Zr and Ti (Figure 12f).

Four samples of mafic to ultramafic inclusions within the Entwine and Osprey Lake intrusions of the sanukitoid suite (*see* MRD 242: Table 8, samples 98DS10, 98DS14, 98DS114 and 98DS202) were determined to be subalkaline tholeiites with high Mg numbers of 62 to 84. The REE profiles and primitive-mantle normalized trace element profiles of the inclusions are similar to those of the host rocks (compare Figures 12f and 13a). These data imply that the ultramafic inclusions could represent the primitive magma from which the sanukitoid magmas were partially melted more so than exotic xenoliths.

Shirey and Hanson (1984) were the first to recognize a distinct suite of plutonic rocks in the western Superior Province having chemical similarities to high-Mg andesites of the Setouchi belt of Japan. These authors applied the term sanukitoid to this monzodioritic plutonic suite and interpreted the high FeO^T, high Mg# and high abundance of transition metals to indicate that the suite originated by partial melting of a peridotitic mantle source. The high LILE and REE abundances were explained as evidence that the mantle source had been metasomatically enriched in these elements prior to partial melting. Stern and Hanson (1991) showed that high-Mg granodiorite to granite, typically found in cores of sanukitoid plutons could have been derived by crystallization differentiation of the monzodioritic phase although Stevenson, Henry and Gariépy (1999) proposed that crustal assimilation played a role in generating the quartz-rich phases. Lassen, Hattori and Percival (2000) noted a compositional change in mantle-derived magmas from early calc-alkalic sanukitoid magmas through late alkalic syenite, nepheline syenite and carbonatite compositions.

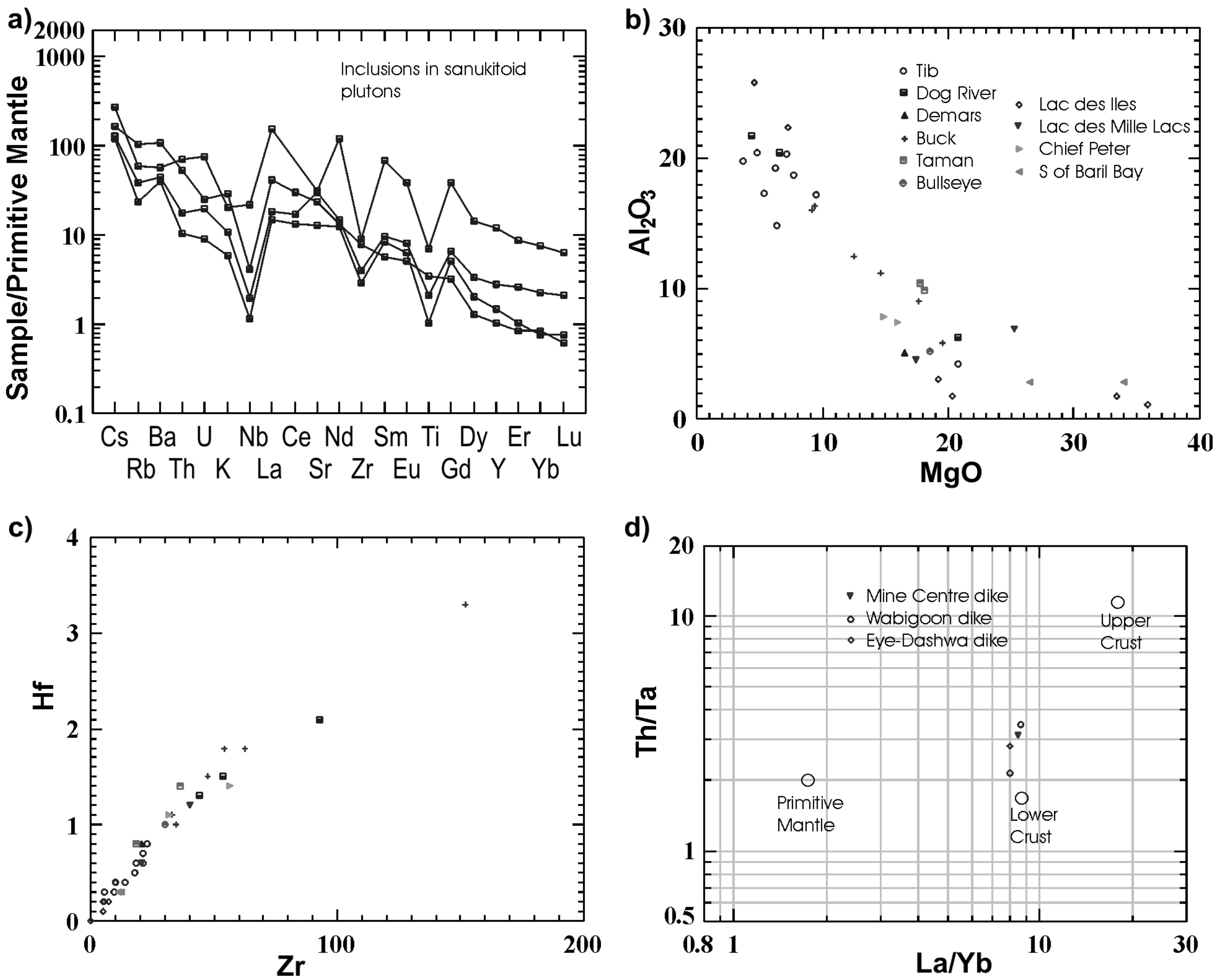

Figure 13. Plots of a) primitive mantle-normalized trace element profiles for mafic inclusions in sanukitoid plutons, b) MgO–Al_2O_3 for intrusions of the Lac des Iles Suite, c) Zr–Hf for intrusions of the Lac des Iles Suite, and d) La/Yb versus Th/Ta plot for diabase dikes.

Sanukitoid plutons were emplaced within a narrow time interval overlapping intrusion of the youngest biotite tonalite plutons and voluminous biotite granite intrusions (*see* Table 7). The interval of sanukitoid magmatism corresponds with the cessation of subprovince-scale subduction and related magmatism and the onset of a brief but pervasive episode of late intracrustal melting. Several mechanisms have been proposed, which may account for the generation of sanukitoid magmas in this tectonic environment. Among these are the introduction of hot mantle at the base of the crust (e.g., Corfu, Stott and Breaks 1995) or slab breakoff due to attempted subduction of continental crust (e.g., Atherton and Ghani 2002).

Mafic Intrusive Rocks

LAC DES ILES–QUETICO SUITES

The Lac des Iles suite represents Neoarchean mafic to ultramafic intrusions in the eastern central Wabigoon Subprovince area. The Lac des Iles suite includes several intrusive bodies: the Lac des Iles, Tib, Buck, Taman and Demars stocks and the Dog River intrusion (*see* Map P.2229). The Quetico suite includes a similar array of Neoarchean mafic to ultramafic dikes and small stocks scattered more-or-less along the northern margin of the Quetico Subprovince. Examples include small stocks south of Chief Peter and Crooked Pine lakes and southwest of the study area at Samuels and Redhorse lakes. Three new intrusions probably belonging to one or another of these suites were identified in the Lac des Mille Lacs area and are located 3 km north of Reserve Bay, 3 km south of Baril Bay and on Case Island of Lac des Mille Lacs.

Various intrusions of the Lac des Iles suite were mapped by Kaye (1966, 1969), Pye (1968), Sage, Breaks, Stott, McWilliams and Robertson (1974), Smith and Sutcliffe (1987), Sutcliffe and Sweeny (1986) and Stone, Fell, Daley et al. (2003). The Lac des Iles Complex is the largest and most economically important member of the suite comprising the 4 km diameter northern ultramafic intrusion centred at Lac des Iles and the elliptical Mine Block intrusion situated immediately to the south. The northern ultramafic intrusion is composed of a layered to massive sequence of altered pyroxenite and peridotite cut by hornblende gabbro and diorite. In the field, clinopyroxene can be recognized by 2 nearly perpendicular cleavages and a greenish colour, whereas orthopyroxene is brownish. Olivine is typically altered to a soft, dull, dark green aggregate of serpentine, chlorite and magnetite although fresh grains of olivine are preserved. The proportions of these mafic minerals and plagioclase vary between magmatic layers.

The Mine Block intrusion is zoned with a gabbro rim and largely gabbronorite core and is mineralized with Pd and lesser Pt, Au, Cu and Ni within 3 main zones characterized by brecciated and pegmatitic gabbro. The Mine Block intrusion has been recently mined by North American Palladium Ltd. The geology and mineralization of the Mine Block intrusion has been extensively described (Watkinson and Dunning 1979; Talkington and Watkinson 1984; MacDonald 1985; Sutcliffe, Sweeney and Edgar 1989; Lavigne and Michaud 2001). The latest work by Hinchey, Hattori and Lavigne (2005) concluded that, although the Lac des Iles deposit is texturally similar to contact-type PGE deposits, the platinum group elements were likely introduced by pulses of primitive magma similar to layered intrusion-hosted deposits. Energetic intrusion of the magmas formed the complex breccia and magma-mingling textures observed in the mine area.

The Tib Lake intrusion is an oval pluton of 7 km diameter composed mainly of layered gabbronorite with local compositional variations from anorthosite to orthopyroxenite and several large xenoliths of mafic volcanic rocks (Smith and Sutcliffe 1987; Smith 1991). Gabbro of variably fine- to coarse-grain size with pegmatitic patches forms the southern border zone of the intrusion and is locally mineralized with Pd and minor Pt, Cu and Ni.

The Buck Lake intrusion, a 7 km long elongate intrusion trending northeasterly, is composed primarily of hornblende gabbro with minor hornblendite, pyroxenite and gabbro breccia. The breccia consists of angular blocks of hornblendite and hornblende melagabbro in a matrix of hornblende gabbro and leucogabbro. Smaller generally lobate bodies of mainly hornblende gabbro to hornblendite composition occur in a cluster on the Dog River, and at Taman and Demars lakes (*see* Map P.2229). Gravity modelling by Gupta and Sutcliffe (1990) suggests that the Dog River intrusions represent apophyses of a larger mafic intrusion beneath a felsic plutonic veneer.

Previous regional mapping by Sutcliffe and Smith (1988) and exploration companies identified hornblende gabbro at Camp Lake (south of the Lac des Iles intrusion) and at Wakinoo, Legris and Towel lakes. Although many authors include these gabbro bodies with the Lac des Iles suite, this study shows that they are dispersed along the rim of the Shelby Lake batholith of the sanukitoid suite. Stone, Lavigne et al. (2003) suggested that these gabbro bodies could represent early mafic phases of the sanukitoid intrusion that have been compressed against the rim of the batholith by emplacement of later core phase. Alternatively, the gabbro bodies represent one or more Lac des Iles intrusions disrupted by the Shelby Lake batholith.

Intrusions of the Quetico suite were not extensively studied during this survey, but were mapped by Irvine (1963) and Pirie (1978) and subsequently studied by Watkinson and Irvine (1964), MacTavish (1999) and Pettigrew and Hattori (2006). Most are small (typically up to several hundreds of metres in size) in comparison with intrusions of the Lac des Iles suite and are dike-like or oval in form. MacTavish (1999) noted that most are composed of hornblende-rich lithologies ranging from diorite through gabbro to hornblendite although larger oval intrusions such as the Samuels Lake intrusion have a rim phase of olivine-bearing clinopyroxenite (Pettigrew and Hattori 2006).

Geochronologic studies show a narrow range of ages from 2699 to 2671 Ma for mafic intrusions of the Lac des Iles area (*see* Table 2, Nos. 130 to 136) although it is unclear if all of these ages represent the Lac des Iles suite. Samples representing the Tib and Lac des Iles intrusions show a narrower age range of 2693 to 2685 Ma and overlap the 2690 Ma age for the nearby Shelby Lake batholith of the sanukitoid suite (Davis 2003; Heaman and Easton 2006). Pettigrew and Hattori (2006) reported an age of 2688+6/–5 Ma for the Samuels Lake intrusion of the Quetico suite, which is identical to the 2688 Ma age for the Blalock intrusion of the sanukitoid suite (*see* Table 2, No. 117) lying within the array of Quetico intrusions.

Samples of various intrusions of the Lac des Iles suite show wide compositional variation depending largely on the proportion of plagioclase in the rock. Samples with a high proportion of plagioclase have high Al_2O_3, Na_2O and Sr, whereas those with a high proportion of mafic minerals have high FeO, MnO and MgO (e.g., Figure 13b). In this context, samples of the Tib Lake intrusion are generally more plagioclase rich than other members of the Lac des Iles suite. Hinchey, Hattori and Lavigne (2005) interpreted a positive correlation between Hf and Zr as evidence for a co-genetic origin for various rocks of the Mine Block intrusion. This same correlation is noted at a larger scale for samples representing intrusions of the Lac des Iles suite as well as 3 stocks in the Lac des Mille Lacs area belonging to one or another of the Lac des Iles or Quetico suites (Figure 13c).

Rare earth element profiles for all lithologic rock types of the Lac des Iles suite are relatively unfractionated (Figures 14a, 14b, 14c and 14d). The average values of $(La/Yb)_N$ for samples of the various intrusions are in the range of 1 to 10 with those of the Buck Lake intrusion showing the highest value of $(La/Yb)_N$ and also overall higher concentrations of REEs. Hinchey, Hattori and Lavigne (2005) determined that most rocks of the Mine Block intrusion are cumulates from a primary komatiitic magma derived by high degrees of partial melting of a moderately depleted, refractory mantle source. A somewhat lower degree of partial melting may account for the LREE enrichment in the Buck Lake intrusion.

Pettigrew, Hattori and Percival (2000) noted that the Quetico and Lac des Iles intrusions fall within a broad east-northeasterly trending swath extending from the central Quetico Subprovince in the west of the

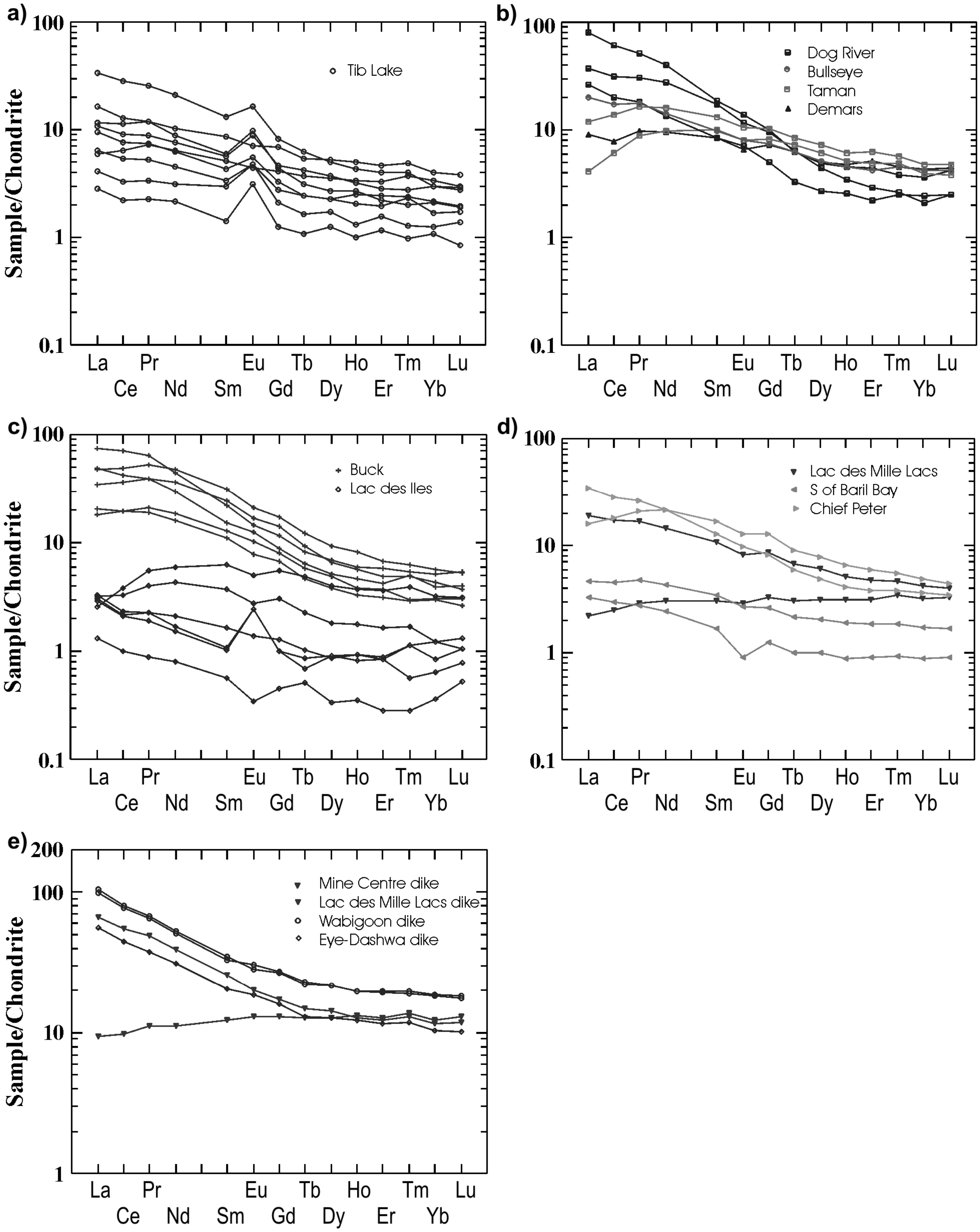

Figure 14. Chondrite-normalized rare earth element plots for (a-d) intrusions of the Lac des Iles Suite and e) diabase dikes.

central Wabigoon Subprovince in the east. Given the similar age of the suites, it is likely that the Quetico and Lac des Iles intrusions developed in the same magmatic event, which probably involved partial melting of the upwelled mantle wedge during accretion of the Wawa arc to the Wabigoon Subprovince (Pettigrew, Hattori and Percival 2000). Several larger intrusions including the Blalock and North and South Elbow intrusions also occur within the swath of Quetico–Lac des Iles intrusions. Pettigrew, Hattori and Percival (2000) and Lassen, Hattori and Percival (2000) have included these with somewhat younger 2680 Ma strongly alkaline intrusions of the southern Quetico Subprovince (cf. Hattori and Percival 1999) although the present mapping has grouped the intrusions at Elbow Lake with the sanukitoid suite.

DIABASE DIKES

A few widely spaced diabase dikes cut Archean rocks of the central Wabigoon Subprovince area. The dikes are typically a few metres to tens of metres wide with sharp, fine-grained and apparently chilled contacts. Although poorly exposed, dike rocks are massive, black weathering brown gabbroic material with diabasic texture. The larger diabase dikes are visible as linear anomalies in aeromagnetic maps (e.g., Geological Survey of Canada–Ontario Department of Mines 1965a-e) and are divisible into west-northwest- and north-trending sets.

Large west-northwest-trending diabase dikes or *en échelon* sets of dikes include the Wabigoon dike that extends more-or-less parallel to Highway 17 from Lac des Mille Lacs to the south of Ignace (*see* Map P.2229) and the Eye–Dashwa dike that crosses Highway 622 north of Atikokan in the Grey Trout Lake area. North-trending dikes include the Mine Centre dike that crosses Highway 11 a few kilometres east of Mine Centre, the Lac des Mille Lacs dikes that extend through western Lac des Mille Lacs and the Empey Lake dike that crosses Highway 17 a few kilometres west of Raith. Other diabase dikes of north-northeast to unknown trend are noted east of Crooked Pine Lake and in the eastern Steep Rock Lake greenstone belt (Stone 2008b, 2008c).

Samples of the Wabigoon, Eye–Dashwa, Mine Centre, Lac des Mille Lacs and Empey Lake dikes were analyzed (*see* MRD 242: Tables 8 and 9). Although FeO was not resolved for the Lac des Mille Lacs dike, other diabase dikes are tholeiitic with Mg numbers of 30 to 44 and K_2O varying from 0.16 to 1.88 weight %. In terms of Th/Ta and La/Yb systematics (Condie 1997), the dikes plot close to the position of lower crust (Figure 13d) and are similar to 1139 Ma dikes of the Sudbury swarm. All are LREE enriched with the exception of the Lac des Mille Lacs dike that has a more-or-less flat REE pattern (Figure 14e).

Fahrig and West (1986), Osmani (1991) and Buchan and Ernst (2004) reported a K/Ar age of 1900 Ma for the Wabigoon dike and Kamineni and Stone (1983) listed K/Ar ages of 1132±27 and 1143±27 Ma for dikes of the Eye–Dashwa swarm. Recent work by Ernst et al. (2006) indicates U/Pb zircon ages of 1137±20 Ma for the Mine Centre dike, 1084.7±9.3 Ma for the Lac des Mille Lacs dike and 1178±31 Ma for the Empey Lake dike. Ernst et al. (2006) noted that the ages and paleomagnetic direction of the Empey Lake and Mine Centre dikes are reliable, whereas the age of the Lac des Mille Lacs dike is less reliable because it was obtained from zircon of presumed hydrothermal origin. Although the orientation is different, the K/Ar ages of the Eye–Dashwa dike are comparable to those of the Mine Centre and Empey Lake dikes, whereas the age of the Wabigoon dike is not easily correlated with the ages of known dikes.

Most dikes of the central Wabigoon Subprovince area were emplaced prior to the main phase of rifting and intrusion of Logan and Nipigon sills at 1115 to 1087 Ma and correlate with a proposed early stage of Midcontinent Rift magmatism at 1150 to 1130 Ma (Heaman et al. 2007). The early stage of Midcontinent Rift magmatism included intrusion of ultramafic lamprophyre dikes (Queen et al. 1996), the Abitibi dikes (Krogh et al. 1987) and possibly felsic volcanism.

NIPIGON SILLS

Irregular, mainly sill-like units of Mesoproterozoic gabbroic rocks of the Lake Nipigon Embayment occur in the northeast corner of the central Wabigoon Subprovince area (*see* Map P.2229). These represent the western limits of gabbroic sills that are extensive in the Lake Nipigon–Thunder Bay area and together with mafic to ultramafic intrusions comprise the Logan igneous suite (Hart and MacDonald 2007). Gabbroic sills of the Logan igneous suite have been subdivided into 5 largely geographic groups including the Nipigon sills (geographically centred on Lake Nipigon), which occur in the present area (Hart and MacDonald 2007; Hollings et al. 2007).

Gabbroic rock of the Nipigon sills is massive, black weathering brown medium- to coarse-grained material with diabasic texture made up of plagioclase, clinopyroxene and a few percent of olivine and magnetite. Although sills attain a thickness of up to 200 m near Lake Nipigon, thicknesses observed in vertical escarpments of the present area are upwards of only a few tens of metres. Regionally, mafic rocks of the Lake Nipigon Embayment intrude flat-lying epicontinental clastic sediments of the Sibley Group, which overlie the Archean basement although in the present area, the sills appear to intrude the basement rocks. At sill contacts, gabbro is chilled to a fine grain size and Archean country rocks can be bleached white.

The distribution of gabbroic rock of the Nipigon sills is strikingly portrayed in aeromagnetic maps of the present area (Ontario Geological Survey 2002). In the aeromagnetic maps, the gabbroic sills appear as magnetic “lows” in contrast with magnetic “highs” corresponding with Archean country rocks. Several oval magnetic highs of 10 to 15 km diameter are rimmed by magnetic lows in the northeast corner of the present area and appear to represent Archean felsic plutonic rocks mantled by gabbro. Possibly, the gabbro sills have an undulating to locally domical shape and have been eroded at the centers of domes to produce the observed pattern of oval magnetic highs rimmed by magnetic lows.

The diabasic sills are subalkalic tholeiites whose geochemistry is discussed by Sutcliffe (1987) and Hollings et al. (2007). Although ages were not determined in the current study, the age of Nipigon sills is likely in the range of 1115 to 1105 Ma based on recent studies of the timing of magmatism in the Lake Nipigon region (Heaman et al. 2007). These authors proposed several stages of Midcontinental Rift magmatism that included 1) an early 1150 to 1130 Ma stage possibly represented by diabase dikes of the present area; 2) 1115 to 1105 Ma magmatism that included emplacement of Nipigon sills, ultramafic intrusions and Osler volcanism; 3) 1100 to 1094 Ma Mamainse Point volcanism and related gabbroic intrusions including the Duluth Complex; and 4) late gabbro intrusions and dikes at 1087 Ma.

Metamorphism

Metamorphism is the mineralogical adjustment that rocks have undergone as a result of having experienced pressure and temperature conditions different from those under which the rock originally formed. The assemblages and compositions of minerals in volcanic and sedimentary rocks that originally formed at surface are particularly sensitive to change as a result of burial with attendant increased pressure and temperature. For example, metavolcanic rocks of the central Wabigoon Subprovince area show “greenschist facies” mineral assemblages including chlorite + albite + carbonate + quartz indicative of mild metamorphism or “amphibolite facies” mineral assemblages including amphibole + plagioclase + quartz indicative of more advanced metamorphism. Extremely high-grade or low-grade metamorphism as shown by granulite-facies assemblages or zeolite-facies assemblages are largely absent from the area. In some instances, the temperature (T) and pressure (P) of metamorphism can be estimated by applying geothermometers and geobarometers, which are essentially equations relating the compositions of minerals in certain assemblages to T and P. Metamorphism of the various supracrustal belts is reviewed below.

Table 10. Pressure and temperature determinations of metamorphism for volcanic and sedimentary rocks using the method of Berman (1988, 1991).

Sample	UTM		Area (Greenstone Belt)	Rock Type	Mineral Assemblage[1]	*P* (kilobars)	*T* (°C)	Notes
	Easting (m)	Northing (m)						
96DS01	572700	5481800	N of Raleigh L. (Raleigh)	Intermediate volcanic	grt + bt + pl + ms + st + qtz	6.6 4.4	635 596	Garnet core Garnet rim
96DS14	583500	5461650	W of Balmoral L. (Raleigh)	Sedimentary migmatite	grt + bt + and + pl + qtz	3.3	640	Garnet core
96DS19	580550	5461500	W. of Balmoral L. (Raleigh)	Gneissic mafic volcanic	grt + bt + pl + hbl + qtz	4.1 2.4	707 468	Garnet core Garnet rim
96DS103	592850	5464150	Elbow L (Raleigh)	Sedimentary migmatite	grt + bt + and + pl + qtz	3.6 2.7	628 590	Garnet core Garnet rim
97DS113	564890	5414610	W of Dovetail L. (Mine Centre)	Anorthosite inclusion	grt + bt + pl + hbl + qtz	9.0	750	Garnet core

[1] mineral abbreviations *from* Kretz (1983): and, andalusite; bt, biotite; grt, garnet; hbl, hornblende; ms, muscovite; pl, plagioclase; qtz, quartz; st, staurolite.

Jackson (1985b) noted that the western Lumby Lake greenstone belt is metamorphically zoned with greenschist-facies assemblages in the south and centre, whereas amphibolite-facies assemblages occur in the north. He attributed higher metamorphism in the north to intrusion of the Norway Lake pluton of the sanukitoid suite, whose thermal energy contact metamorphosed the north side of the belt. Indeed, Ayres (1978) stated that many large greenstone belts in the Superior Province are metamorphically zoned with greenschist-facies cores, reflecting regional burial metamorphism, and amphibolite-facies rims, reflecting contact metamorphism by surrounding late batholiths. In contrast, small greenstone belts tend to be at amphibolite grade because the thermal aureoles of surrounding batholiths have penetrated the entire belt.

The eastern Lumby Lake greenstone belt also has greenschist-facies mineral assemblages in the centre and amphibolite-facies assemblages at the rims, although Stone and Metsaranta (2002) interpreted rocks at the core of the belt to be part of the Pinecone assemblage, which is younger than assemblages on the sides of the belt. Hence, the core-rim variation in metamorphism of the belt may be due to the core rocks being younger than those at the rim and having escaped an early episode of amphibolite-facies metamorphism.

Mafic metavolcanic rocks in many small greenstone belts including the Graham, Phyllis Lake, Heaven Lake, Garden Lake and Lac des Iles belts have mineral assemblages containing amphibole. Indeed, the mafic metavolcanic rocks that originated mainly as pillowed lava flows have also been deformed to the extent that many are strongly foliated amphibole gneisses. These reflect a regional amphibolite grade of metamorphism in the dominantly plutonic areas.

Likewise, metavolcanic rocks of the Otukamamoan Lake and Raleigh Lake greenstone belts show mainly amphibole-bearing mineral assemblages, although Smith (1993) stated that greenschist-facies mineral assemblages are common in broad parts of the Otukamamoan Lake greenstone belt (Vista Lake area) west of the present area. Four samples of supracrustal rocks from the Raleigh Lake greenstone belt contain garnet-bearing mineral assemblages suitable for the thermobarometric calculations of Berman (1988, 1991). These give pressures of 3.3 to 6.6 kilobars and temperatures of 635 to 707°C using garnet core compositions and pressures of 2.4 to 4.4 kilobars and temperatures of 468 to 596°C using garnet rim compositions (Table 10).

Poulsen (2000a) defined an amphibolite-greenschist-facies transition zone in the Mine Centre greenstone belt where parts of the belt including and lying west of the Ottertail Lake stock are at amphibolite facies giving way to greenschist facies to the east and south. Mapping as part of this study contrasts somewhat with this interpretation by recording assemblages of amphibole + garnet within metavolcanic rocks of the northern Calm Lake area. Going south across the Mine Centre greenstone belt, metamorphic grade initially decreases to greenschist facies more-or-less at the Wabigoon–Quetico subprovinces boundary south of which it increases toward the central axis of the Quetico Subprovince.

The central and southern Steep Rock Lake greenstone belt also shows low-grade metamorphism with, for example, an assemblage of chlorite + albite + quartz + carbonate in pelitic rocks. Metamorphic grade increases south of the Steep Rock Lake belt with the appearance of biotite followed by amphibole and garnet within Quetico sediments as mapped by Stone and Kamineni (1989). Pressure–temperature conditions of 3.5 kilobars and 550°C are recorded for garnet-bearing samples from the Quetico belt (Kamineni, Stone and Johnson 1991).

The Steep Rock Lake greenstone belt provides an interesting contrast in metamorphic mineral assemblages. The iron ore deposits of the Steep Rock assemblage at the northeast margin of the Steep Rock Lake belt show evidence of very low temperature metamorphism or alteration. For example, the iron-ore deposits of the eastern Steep Rock Lake belt have a mineral assemblage of goethite + aragonite + quartz (Smith 1942) that has been interpreted as evidence of post-Archean alteration or weathering (*see* discussions of Shklanka (1972), Kimberley and Sorbara (1976) and Wilks (1986)). This interpretation is somewhat problematic because the very low-grade metamorphism or alteration is restricted to the Jolliffe Ore Zone Formation and nearby metavolcanic rocks have typical greenschist mineral assemblages. It is more likely that the Steep Rock assemblage was uplifted and weathered in the Neoarchean after which it was overthrust by the Dismal assemblage and was never metamorphosed by younger plutons. Indeed, the Steep Rock assemblage at the northeast margin of the belt unconformably overlies the Marmion batholith and has not been intruded by late plutons. In contrast, volcanic strata at the northwest margin of the belt are intruded by a variety of plutons and are at least 200 million years (m.y.) older than the Steep Rock assemblage. Amphibolite-facies mineral assemblages in volcanic strata at the northwest margin of the belt are probably due to this part of the belt having been intruded by late plutons or else due to these older rocks having felt an episode of early amphibolite-facies metamorphism prior to deposition of the Steep Rock assemblage in the Neoarchean.

The Finlayson Lake greenstone belt is also metamorphically variable. Recent mapping (Stone 2008b) shows the 2.73 Ga Witch Bay assemblage at the core of the belt with a greenschist-facies mineral assemblage of chlorite + epidote + albite + quartz, whereas the 2.93 Ga Finlayson West and 3.0 Ga Finlayson East assemblages on opposite flanks of the belt have amphibolite-facies mineral assemblages. Here again, the amphibolite-facies rims of the belt may be due to the rims having been intruded by late hot plutons or else the rims were subjected to amphibolite-facies metamorphism before deposition and overthrusting of the Witch Bay assemblage at or shortly after 2.73 Ga.

The linear Lac des Mille Lacs belt shows mineral assemblages indicative of amphibolite-facies metamorphism. A narrow and possibly discontinuous zone of greenschist-facies mineral assemblages occurs approximately centred at the south margin of this belt and the adjacent Quetico belt. Southward, metamorphic grade increases within the Quetico belt and sedimentary rocks are migmatized.

THE AGE OF METAMORPHISM

The age of metamorphism can be established in some instances through geochronology or can be inferred from geologic information. Minerals such as titanite and monazite with a low blocking temperature for Pb-loss can be dated by the U/Pb method and the resulting ages used to interpret the ages of post-crystallization tectonic and metamorphic events. In some instances, zircon grains can grow during metamorphism and develop either outer rims or new crystals characterized by low thorium (Davis 2008). The U/Pb ages of these overgrowths or new crystals can be used to infer the timing of metamorphic events.

Table 2 lists U/Pb ages of metamorphism (*see* “Metamorphic Age”) established by one or another of the above methods for a few samples from the central Wabigoon Subprovince area. These ages range from 2722 to 2657 Ma and generally indicate late Neoarchean metamorphism for rocks, some of which crystallized 200 m.y. earlier. Valli, Guillot and Hattori (2004) observed a similar trend of late metamorphism in Quetico sediments based, in part, on a monazite age of 2667 Ma. General inspection of the data in Table 2 indicates, however, that there could have been older metamorphic events such as contact metamorphism associated with intrusion of tonalite plutons at about 2.87 Ga in the Hillyer domain, 2.93 Ga in the central Wabigoon domain, 2.96 Ga in the Whitton domain or at 3.0 Ga in the Marmion domain. Possibly, the evidence for these early metamorphic events including minerals with low blocking temperatures or zircon overgrowths have been obliterated by the latest metamorphism.

Geologic evidence supports early metamorphic events. For example in the Finlayson Lake greenstone belt, the 2.93 Ga Finlayson West assemblage and the 3.0 Ga Finlayson East assemblage show amphibolite-facies mineral assemblages and are either overlain by the upper greenschist-facies Witch Bay assemblage at 2.73 Ga or have been overthrust by the Witch Bay assemblage after 2.73 Ga. Regardless of their structural relation to the Witch Bay assemblage, the older assemblages could have been metamorphosed to amphibolite facies in the 200 m.y. interval prior to emplacement of the Witch Bay assemblage. The Steep Rock Lake and Lumby Lake greenstone belts show a similar pattern of amphibolite-facies metamorphism in old assemblages and greenschist-facies minerals in young assemblages. In places, the amphibolite-greenschist facies transition can be attributed to a high-grade metamorphic aureole associated with intrusive batholiths, but elsewhere, the transition is sharp and localized at assemblage boundaries. In the latter case, the variation in metamorphic grade is more likely related to the ages of the rocks with older assemblages having been strongly metamorphosed prior to development of the younger assemblage.

ALUMINUM-IN-HORNBLENDE BAROMETRY

The aluminum- or Al-in-hornblende barometer has been applied widely to felsic plutonic rocks of the central Wabigoon Subprovince area. The Al-in-hornblende barometer takes account of the linear relation between the aluminum content of hornblende in a rock containing an appropriate mineral assemblage and the pressure or depth at which the rock crystallized. Although the aluminum content of hornblende was originally shown to be pressure dependent on an empirical basis (Hammarstrom and Zen 1986; Hollister et al. 1987; Schmidt 1992), Holland and Blundy (1994) showed that temperature also affects the aluminum content of hornblende and these authors produced a hornblende thermometer. Anderson and Smith (1995) subsequently produced a temperature-dependent barometer that is used here and in other regional tectonic studies (Stone 1998, 2005b).

The compositions of coexisting amphibole+plagioclase pairs were determined by microprobe analyses (*see* MRD 242: Tables 11 and 12 (amphiboles) and Tables 13 and 14 (feldspars)). Various barometers and thermometers are applied to these mineral compositions and the results (P_{Schmidt} , $T_{\text{Holland and Blundy}}$ and $P_{\text{Anderson and Smith}}$) are listed with the amphibole analyses, where P_{Schmidt} represents the barometer of Schmidt (1992), $T_{\text{Holland and Blundy}}$ represents the thermometer of Holland and Blundy (1994) and $P_{\text{Anderson and Smith}}$ represents the temperature-corrected barometer of Anderson and Smith (1995). The temperature-corrected barometer of Anderson and Smith (1995) contains several chemical tests to determine if amphibole analyses are suitable for barometry and, in approximately one-third of the samples, the amphibole analyses fail these tests ($P_{\text{Anderson and Smith}}$= unsuitable).

RESULTS

Hornblende-bearing plutonic suites including the hornblende tonalite, sanukitoid and gneissic suites are ideal for Al-in-hornblende barometry and have been systematically sampled. A few *P–T* determinations are made from other plutonic suites including the biotite tonalite and biotite granite suites at localities where amphibole occurs as an auxiliary mineral phase. The *P–T* determinations are also made from some of the more felsic (largely quartz diorite) phases of mafic intrusions such as those of the Lac des Iles suite. In contrast, metavolcanic rocks do not normally contain the required mineral assemblage with the result that virtually no Al-in-hornblende data are available for greenstone belts. Quetico metasedimentary rocks contain amphibole-bearing granitic neosome layers that appear to develop during advanced metamorphism and prior to partial melting. A few samples of these amphibole-bearing layers were tested with routine Al-in-hornblende barometry, but yielded highly variable results (*see* locations near Nym Lake in Figure 16 (discussed below)), possibly because the neosome layers do not have the requisite mineral assemblage. Further work is required to determine if Al-in-hornblende barometry can be applied to migmatized sedimentary rocks.

Temperature and pressure data are grouped according to various crustal blocks or plutonic suites in Table 23 and an average is shown in bold for data representative of the blocks or plutonic suites. Some data are excluded from the averages in situations such as where the host rock possibly postdates the crustal block or where the host rock is an inclusion of uncertain origin within the plutonic suite. Also shown are the known or inferred ages of the pressure determination where known age refers to an age determination at the site where the sample was taken for pressure determination. In the case of late homogeneous granite plutons, the pressure and known age determinations can be distant from each other within the same pluton. In these cases, it is reasonable to conclude that the age determination represents the time at which hornblende compositions were “locked in”. Inferred age refers to a situation where the pressure and age determinations are made at separate localities within the same crustal block and for rock types other than late plutonic suites. In this case, the extent to which the age determination marks the timing of the pressure determination is more tenuous. Table 24 summarizes the average pressure and temperature data for various crustal blocks or plutonic suites.

TEMPERATURE

Aluminum-in-hornblende temperatures were calculated using the thermometer of Holland and Blundy (1994) and are summarized in Tables 23 and 24. The average temperatures for hornblende-bearing mineral assemblages in crustal blocks and plutonic suites (*see* Table 24) vary between 676 and 732°C and do not seem to show systematic spatial variation. These temperatures are intermediate between the low melting temperatures for water-saturated granitic melts and the high melting temperatures for dry granite melts (Wyllie 1977; Carroll and Wyllie 1990). The temperatures indicate that magmas were not saturated in water and justify the use of the temperature-corrected barometer of Anderson and Smith (1995).

Figure 15a shows temperature variation through time for various crustal blocks and plutonic suites. Each temperature measurement has an associated error of ±40°C and averaging tends to reduce this error. Nonetheless, average temperatures appear to be approximately constant at about 720°C from 3.0 to 2.7 Ga, after which temperatures decrease for late plutonic suites and intrusions of the Quetico Subprovince. Possibly the late temperature decrease reflects the Neoarchean thickening, stabilization and overall cooling of the crust.

PRESSURE

Aluminum-in-hornblende pressures are determined by the method of Anderson and Smith (1995) in Table 23 and are averaged and arranged in order of increasing pressure for various crustal blocks and plutonic suites (*see* Table 24). Average pressures are approximately constant at about 6 kilobars for most early crustal blocks. In general, these average pressures are somewhat higher than what is found for late plutonic suites of the Berens River area (2 to 4 kilobars; Stone 1998) and for various crustal blocks and plutonic suites in the northern Superior area (3 to 6.6 kilobars; Stone 2005b). The data imply that the present area is more deeply eroded than northerly parts of the Superior Province, although this area also contains a greater proportion of older rocks. High pressures in the central Wabigoon Subprovince area may simply imply that old crust has undergone more uplift than young crust.

Table 23. Aluminum-in-hornblende data for crustal blocks and plutonic suites of the central Wabigoon Subprovince area.

Sample	Rock	Area	$P_{Schmidt}$ (±1 kilobar)	$T_{H\&B}$ (±40°C)	$P_{A\&S}$ (±1 kilobar)	Inferred Age (Ma)	Known Age (Ma)	Crustal Block
95DS20	12b	Companion	7.1	787	4.8	3000		Marmion
95DS15	11b	Mercutio	5.5	697	5.2	3002		Marmion
95DS18	11a	S Lumby	6.1	734	5.2	3009	3009	Marmion
95DS14	11b	Mercutio	6.8	755	5.3	3002		Marmion
95DS21	16b	Companion	6.8	744	5.6			Marmion
05DS62b	11c	W Raven Lake	6.3	723	5.6	2989		Marmion
95DS13	12b	Mercutio	6.9	746	5.6	3000		Marmion
05DS61	12c	W Raven Lake	6.9	745	5.7	2989	2989	Marmion
00DS19	11b/16b ?	W Raven	7.4	760	5.8	2989	2989	Marmion
95DS17a	11a	S Lumby	7.0	730	6.1	3009	3009	Marmion
95DS70	16b	Bedivere	7.3	715	6.6			Marmion
05DS52	11c	Sawmill Bay	7.4	687	7.2	3001		Marmion
95DS215	16b	E Lizard	5.4	681				Marmion
04DS03	10a	W Lac des Mille Lacs	6.8			3001		Marmion
04DS04	11c	N Panache Lake	8.4			3001		Marmion
08DS35a	16b	Hydro Plant	2.2			3001		Marmion
07DS47	16b	S of Lumby	5.0	638	5.3		2999	Diversion / Marmion
02DS03	12g	Lac des Mille Lacs	6.7	711	6.2			Marmion
95DS31	16b	NE Sawbill	6.3	705	5.8	2999		Diversion
02DS28	16b	W Upsala	7.1	716	6.4	2999		Diversion
95DS39	16b	N Wilson	8.2	686	8.0	2999		Diversion
95DS28	16b	Old Man	9.6	679		2999		Diversion
			6.7	**718**	**5.9**	**2999**	**2999**	**Marmion**
95DS66	16b	N Redpaint	6.6	712	6.1			Whitton
96DS41	16a	Phyllis Lake	5.3	688	5.2			Whitton
00DS43	16b	W Graham	6.5	763	4.9	2953		Whitton
03DS02	10g	Muise Lake	6.7	815	3.7			Whitton / central Wabigoon
03DS06	12c	Muise Lake	5.3	749	4.2			Whitton / central Wabigoon
02DS39	16b	Trewartha	5.7	695	5.4			Whitton / late plutonic
08DS01	11b	W of Eye–Dashwa pluton	7.3	737	6.2			Whitton
08DS03	10d-16b	W of Finlayson	6.8	681	6.8			Whitton
08DS16	11c	N of Turtle	6.6	739	5.5			Whitton
08DS19	11a	N of Turtle	6.1	745	4.9		2957	Whitton
08DS28	12m	W Hardtack	6.9	700	6.5			Whitton
08DS32	16b	N Hardtack	7.4	730	6.5			Whitton
08DS30	16b in 11b	NW Hardtack	8.6	695				Whitton Dashwa gneisses
			6.6	**727**	**5.5**	**2953**	**2957**	**Whitton**
00DS10	11b	NL Turtle	6.5	747	5.3			central Wabigoon
97DS33	12a	Trout Lake	0.9	635				central Wabigoon
96DS43	16b	Hawk Lake	5.0	711	4.5			central Wabigoon
00DS40	16c	S Wawang	3.4	702	3.1			central Wabigoon
03DS71	16b	Bonnie	7.6	700	7.2			central Wabigoon
03DS75	16b	NE Bonnie	6.4	695	6.2			central Wabigoon
03DS76	16b	E Bonnie	8.9	725	8.0			central Wabigoon
08DS05	11b	W of Mine Centre	3.4	673				central Wabigoon
00DS52	11b	Mack	5.4	733	4.5	2925		central Wabigoon
05DS56	12c	Oscar	7.6	798	4.9	2929	2929	central Wabigoon

$P_{Schmidt}$ represents the barometer of Schmidt (1992), $T_{H\&B}$ represents the thermometer of Holland and Blundy (1994) and $P_{A\&S}$ represents the barometer of Anderson and Smith (1995).

Table 23. continued

Sample	Rock	Area	$P_{Schmidt}$ (±1 kilobar)	$T_{H\&B}$ (±40°C)	$P_{A\&S}$ (±1 kilobar)	Inferred Age (Ma)	Known Age (Ma)	Crustal Block
00DS09	11b	WL Turtle	5.6	709	5.2	2924	2924	central Wabigoon
05DS69	12g-16b	E Pakashkan	7.1	736	6.1	2925		central Wabigoon
05DS64	12c-16b	S Garden Lake	7.7	753	6.2	2924	2924	central Wabigoon
05DS65a	12c-16b	S Garden Lake	7.7	750	6.3	2924	2924	central Wabigoon
05DS65b	12c-16b	S Garden Lake	7.8	747	6.4	2924	2924	central Wabigoon
02DS73	16b	Mooseland Lake	7.3	724	6.5			central Wabigoon
05DS58	11b	E Trewartha	7.2	737	6.2	2952		Whitton /central Wabigoon
00DS35	11a	N Trewartha	7.2	717	6.6	2952		central Wabigoon
95DS36	12m	S Serpent	7.6	672	7.6	2936		central Wabigoon
05DS141	10g	W Miranda Lake	8.6	679	8.5	2925		central Wabigoon
03DS67a	11b	E Armistice	10.0	714	9.2	2924		central Wabigoon
00DS07	11b	L Turtle	3.9	699		2924		central Wabigoon
00DS08	11b	WL Turtle	1.9	662		2924		central Wabigoon
02DS72	16b	Mooseland Lake	8.4	756				central Wabigoon
			6.4	**716**	**6.2**	**2929**	**2925**	**central Wabigoon**
08DS43	11a	W Manion	1.3	606		2925		Hillyer / central Wabigoon
00DS05	16b	W Miranda	5.0	731	4.2	2883	2883	Hillyer
97DS04	16b	S Manion	7.7	756	6.1	2869	2869	Hillyer
95DS44	16b	W L Muddy	8.3	710	7.7	2883		Hillyer
97DS07	16b	SE of Dovetail Lake	8.5	715	7.8	2875		Hillyer
97DS104	16b	E of Roscoe Lake	4.7	666		2883		Hillyer
06DS52	12c	E Manion	6.4	712	5.8	2866	2866	Hillyer
08DS23	16b	S Turtle	7.8	742	6.6	2877		Hillyer
08DS25	16b	W Nevison	6.6	714	6.1	2877		Hillyer
08DS37	12m	N Wild Potato	5.7	708	5.3	2877		Hillyer
08DS40	10r	N Wild Potato	6.0	748	4.8	2877		Hillyer
08DS42	16b	W Manion	7.4	753	5.9	2877		Hillyer
08DS45	12c	E Manion	7.4	753	6.0	2866	2866	Hillyer
08DS46	10r	E Manion	6.0	704	5.6	2866		Hillyer
08DS27	11b	E Nevison	7.3	698	7.0		2886	Hillyer
97DS107	16c	E of Roscoe Lake	4.2	678	4.2	2883		Hillyer
97DS113	10r	W Dovetail Lake	10.9	718	10.0			Hillyer / inclusion in late plutonic
			6.5	**712**	**6.2**	**2879**	**2874**	**Hillyer**
96DS49	16b	Scotch Lake	4.0	702	3.7			Pinecone / central Wabigoon
05DS60	12c	E Raven Lake	6.3	739	5.3	2817	2817	Pinecone / Savoy
			5.1	**720**	**4.5**	**2817**		**Pinecone / Savoy**
02DS09	16b	Lac des Mille Lacs	7.0	709	6.5			Dog / inclusion in late plutonic
02DS64	16b	NE Sawmill Bay	5.6	730	4.8	2781		Dog
05DS71	10d in 15g	Mug Lake	5.9	734	5.0	2780		Dog
05DS70	11c	Mug Lake	6.7	743	5.5	2780		Dog
05DS85	11c	W Shelby	6.6	728	5.8	2780		Dog
02DS04	16b	Lac des Mille Lacs	6.3	695	6.0	2781		Dog
03DS50	16b	SE Taman	7.5	754	6.0	2780		Dog
05DS84	11c	W Shelby	6.9	716	6.3	2780		Dog
02DS59	16b	Lac des Mille Lacs	8.9	707	8.3	2781		Dog
02DS09	16b	Lac des Mille Lacs	7.0	709	6.5			Dog
			6.8	**723**	**6.1**	**2780**		**Dog**

$P_{Schmidt}$ represents the barometer of Schmidt (1992), $T_{H\&B}$ represents the thermometer of Holland and Blundy (1994) and $P_{A\&S}$ represents the barometer of Anderson and Smith (1995).

Table 23. continued

Sample	Rock	Area	$P_{Schmidt}$ (±1 kilobar)	$T_{H\&B}$ (±40°C)	$P_{A\&S}$ (±1 kilobar)	Inferred Age (Ma)	Known Age (Ma)	Crustal Block
03DS57	16b	W Lac des Iles	6.5	728	5.6	2728		Whistle / Dog
02DS16a	16b	S Upsala	8.4	697	8.0	2717		Whistle / Marmion
02DS19	10g	SE Upsala	8.6	709	8.0	2717		Whistle / Marmion
02DS16b	10a	S Upsala	9.4	707	8.8	2717		Whistle / Marmion
02DS20	16b	Lac des Mille Lacs	6.7	725	5.9			Whistle / Marmion
02DS15	10a	S Upsala	8.6	720	7.8			Whistle / Marmion
03DS18	16b	E Buck Lake	6.4	725	5.7			Whistle
03DS36	16c	S Tib Lake	6.5	732	5.6			Whistle
			7.6	**718**	**6.9**	**2720**		**Whistle**
00DS109	16c	SE Radio	4.6	722	4.0			Winnipeg River
00DS110	16c	S Radio	4.5	709	4.1			Winnipeg River
05DS53a	11c	E Wawang	5.6	739	4.6	2712	2712	Winnipeg River
00DS24	16b	Selwyn	6.1	756	4.7	2721		Winnipeg River
00DS23	16b	W Selwyn	5.9	736	4.9	2721		Winnipeg River
05DS55	11c	E Wawang	5.8	716	5.3	2712	2712	Winnipeg River
00DS112	16b	N Wawang	6.0	738	5.1	2721		Winnipeg River
00DS20	11b	Valora	6.7	741	5.6	2704		Winnipeg River
			5.6	**732**	**4.8**	**2715**	**2712**	**Winnipeg River**
96DS22	16b	Agimack Lake	6.6	739	5.6			western Wabigoon
96DS23	16c	Agimack Lake	7.1	732				western Wabigoon
96DS26	16a	E of Agimack Lake	5.8	719	5.2			western Wabigoon
96DS29	16a	E of Raleigh	6.7	730	5.9			western Wabigoon
98DS20	16c	N Pekagoning	4.3	703	4.0	2721		western Wabigoon
99DS19	11a	N Entwine	5.7	743	4.6	2716	2716	western Wabigoon
00DS13	11b	NW Eltrut	6.5	775	4.6	2732		western Wabigoon
98DS29	16b	N Bernadine	5.1	709	4.7	2721		western Wabigoon
99DS25	5g	S Entwine	5.7	732	4.9			western Wabigoon
98DS22	16b	N of Turtle River bridge	5.5	717	4.9	2721		western Wabigoon
00DS12	11b	N Heron	6.1	745	5.0	2732		western Wabigoon
00DS04	11b	S Eltrut	5.9	733	5.0			western Wabigoon / Hillyer
00DS03	12a	N Eltrut	6.4	757	5.0	2688	2688	western Wabigoon / Winnipeg River
00DS16	11b	W Pekagoning	6.0	734	5.1	2732		western Wabigoon
97DS105	16b	SW of Clearwater	5.8	714	5.2	2697	2697	western Wabigoon
98DS12	11b	SE Entwine	6.1	723	5.3			western Wabigoon
99DS02	11a	Fishtrap	6.4	699	6.1			western Wabigoon
00DS15	11b	N Islets	6.7	714	6.1	2732		western Wabigoon
98DS156	16b	W of Butler	7.6	751	6.2	2721		western Wabigoon
99DS07	16c	N Vedette	4.5	737		2721		western Wabigoon
99DS09	11b	S Bearpaw	2.0	640				western Wabigoon
99DS26	5g	S Entwine	7.7	775				western Wabigoon
99DS108	16c	E of Kawawiag	2.1	697		2721		western Wabigoon
08DS08	12b	W Little Sawbill	6.5	733	5.5	2721		western Wabigoon
08DS12	14 or 12	Vickers	3.7	690		2690		western Wabigoon
08DS15	16c	N Vickers	2.3	682	2.2	2721		western Wabigoon
98DS235	16c	S Asinn Lake	5.1	724	4.4			inclusion in late plutonic
98DS24	16b	N Pekagoning	4.6	696	4.37			rim of sanukitoid
			5.5	**723**	**5.0**	**2718**	**2700**	**western Wabigoon**

$P_{Schmidt}$ represents the barometer of Schmidt (1992), $T_{H\&B}$ represents the thermometer of Holland and Blundy (1994) and $P_{A\&S}$ represents the barometer of Anderson and Smith (1995).

Table 23. continued

Sample	Rock	Area	$P_{Schmidt}$ (±1 kilobar)	$T_{H\&B}$ (±40°C)	$P_{A\&S}$ (±1 kilobar)	Inferred Age (Ma)	Known Age (Ma)	Crustal Block
04DS80b	11a	Raith	4.1	691	4.0	2673	2673	Quetico
04DS80a	11a	Raith	4.5	696	4.3	2673	2673	Quetico
08DS48	7b	SE Atikokan	3.8	642	4.0	2695		Quetico
			4.1	**676**	**4.1**	**2680**	**2673**	**Quetico**
02DS65	15f	Muskeg Lake	4.5	726	3.9	2685		late granite
05DS47	15f	Savanne	3.5	707	3.2	2686	2686	late pluton
05DS46b	15f	Savanne	4.3	737	3.5	2686	2686	late pluton
05DS48b	15f	Savanne	4.0	710	3.6	2686	2686	late pluton
05DS46a	15f	Savanne	4.1	684	4.0	2686	2686	late pluton
05DS48c	15f	Savanne	4.7	706	4.4	2686	2686	late pluton
05DS54	15d	E Wawang	4.1	711	3.7	2685		late plutonic
98DS119	15f	S Pekagoning Lake	4.0	703	3.7	2685		late plutonic
06DS51	15b	NW Joe	4.7	712	4.2	2686	2686	late plutonic / Hillyer
08DS44	16b	S Manion	4.3	669	4.3	2686		late plutonic / Hillyer
99DS54	11b	S. Downhill	3.0	673	3.0			gneissic inclusion in biotite granite
96DS60	16c	N Nora	4.6	694	4.42			inclusion in White Otter batholith
98DS119	15f	S Pekagoning Lake	4.0	703	3.66			inclusion in White Otter batholith
			4.1	**703**	**3.8**	**2686**	**2686**	**mainly biotite granite**
95DS03	14b	S Sapawe	4.4	719	3.8	2688	2688	sanukitoid in Quetico
96DS46	14b	N Spook Lake	4.1	796	2.1	2690		sanukitoid
96DS47	14c	Norway	3.4	736	2.6	2690		sanukitoid
00DS45	14c	NW Graham	2.7	680	2.7	2690		sanukitoid
97DS32	14c	Mine Centre	2.9	701	2.7	2686	2686	sanukitoid
95DS230	14b	Norway	3.9	758	2.7	2690		sanukitoid
98DS102	14c	N of Jones	3.1	687	3.0	2685	2685	sanukitoid / Entwine
95DS223	14a	Elbow	3.1	682	3.1	2690		sanukitoid
99DS13	14c	W Kenozhe	3.9	736	3.2	2685		sanukitoid / W Entwine
95DS37	14b	Eye Dashwa	3.5	689	3.3	2684		sanukitoid
00DS02	14b	E Entwine	3.5	691	3.4	2685	2685	sanukitoid
95DS130	16c	N Van Nostrand Lake	4.4	741	3.5	2690		sanukitoid
03DS41	14e	Coldwater	3.8	700	3.5	2690		sanukitoid
95DS229	14b	Norway	4.6	745	3.6	2690		sanukitoid
97DS136	14a	W Redgut Bay	3.7	677	3.6	2690		sanukitoid
99DS57	14b	E Raven	3.8	683	3.8	2690		sanukitoid
99DS43	14a	N Entwine	5.1	757	3.8	2685	2685	sanukitoid
03DS53	14b	N Towle	4.2	714	3.8	2690		sanukitoid
00DS01	14b	Eye Dashwa	4.2	707	3.8	2688		sanukitoid
98DS25	14b	N Pekagoning	4.1	695	3.9	2690		sanukitoid
99DS47	14c	N Entwine	4.2	703	3.9	2685	2685	sanukitoid
95DS224	14a	Elbow	3.9	668	3.9	2690		sanukitoid
00DS50	14f	Pakashkan	4.5	719	4.0	2690		sanukitoid
03DS43	14e	W Wakinoo	4.6	726	4.0	2690		sanukitoid
00DS47	14f	SW Pakashkan	4.4	708	4.0	2690		sanukitoid
98DS117	14b?	Eric Lake	4.2	691	4.1	2685		sanukitoid / Entwine
00DS41	14f	Ken Lake	4.6	716	4.1	2690		sanukitoid
00DS49	14c	Pakashkan	4.7	715	4.2	2690		sanukitoid
04DS47	14d	Little Athestane Lake	5.2	738	4.3	2690		sanukitoid
04DS41	14c	Linko	5.3	737	4.4	2690		sanukitoid

$P_{Schmidt}$ represents the barometer of Schmidt (1992), $T_{H\&B}$ represents the thermometer of Holland and Blundy (1994) and $P_{A\&S}$ represents the barometer of Anderson and Smith (1995).

Table 23. continued

Sample	Rock	Area	$P_{Schmidt}$ (±1 kilobar)	$T_{H\&B}$ (±40°C)	$P_{A\&S}$ (±1 kilobar)	Inferred Age (Ma)	Known Age (Ma)	Crustal Block
05DS80b	14d	NE Wakinoo	5.2	724	4.5	2690		sanukitoid
99DS49e	14a	ER1 61.0m	5.7	740	4.7	2685	2685	sanukitoid
05DS80a	14f	NE Wakinoo	5.5	723	4.8	2690		sanukitoid
05DS79	14c	NE Wakinoo	6.3	736	5.3	2690		sanukitoid
95DS129	14a	Van Nostrand	2.2	689		2690		sanukitoid
96DS45ab	14b	Norway	3.8	743		2689		sanukitoid
96DS50	14c	Norway	3.1	717		2690		sanukitoid
96DS51	14b	Norway	3.5	725		2690		sanukitoid
98DS111	14b	N Wapageisi R	3.0	697		2690		sanukitoid
99DS17	14b	E Kenozhe	2.4	712		2685		sanukitoid / Entwine
99DS21	14b	N Entwine	2.3	696		2685	2685	sanukitoid
99DS28	14a	Entwine	3.2	744		2685	2685	sanukitoid
99DS40	14b	N Entwine	2.5	716		2685	2685	sanukitoid
99DS49d	14a	ER1 51.3m	2.9	682		2685	2685	sanukitoid
99DS50b	14a	ER10 24.0m	2.4	626		2685	2685	sanukitoid
00DS06	14b	Otter Tail	0.5	628		2686	2686	sanukitoid
03DS45	14b	E Wakinoo	3.2	703		2690		sanukitoid
03DS55	14b	Trumper Lake	9.6	761		2690		sanukitoid
03DS58	14c	E Shelby	3.7	711		2690		sanukitoid
03DS60	14f	SW Legris	2.6	692		2690		sanukitoid
03DS61	14g	Legris	2.3	691		2690		sanukitoid
03DS64	14g	Legris	4.5	729		2690		sanukitoid
03DS91	14b	Camp Lake	1.8	682		2690	2690	sanukitoid
04DS66	14b	Elbow Lake	7.8			2690		sanukitoid
05DS81	14c	NE Wakinoo	10.5	744		2690		sanukitoid
05DS82b	14b	NE Wakinoo	9.2	768		2690		sanukitoid
05DS82c	14b	NE Wakinoo	9.0	760		2690		sanukitoid
05DS86	14c	Wakinoo Lake	3.3	694		2690		sanukitoid
06DS23	14b	N Entwine	4.6	718	4.1	2685	2685	sanukitoid
06DS47	14b	Niven Lake	3.4	675		2690		sanukitoid
95DS137	14a	Miserable	6.3	814	3.4			sanukitoid / Quetico
95DS138	14c	Miserable	3.5	692	3.3			sanukitoid / Quetico
			4.2	**714**	**3.7**	**2689**	**2686**	**sanukitoid**
03DS39	10r	W Tib Lake	6.2	762	4.6			Lac des Iles - Tib
03DS79	10d	Lac des Iles	5.1	792	3.0	2691		Lac des Iles
05DS67	10d	Tib	4.1	742	3.2	2690		Lac des Iles
05DS66	10x	Tib	5.0	774	3.3	2690		Lac des Iles
05DS01c	10d	Lac des Iles Baker zone	4.1	693	3.9	2689	2689	Lac des Iles
05DS68	10d	Tib	5.7	763	4.2	2690		Lac des Iles
03DS09	15d	Buck Lake	6.2	728	5.4	2690		Lac des Iles
03DS19	10g	Taman Lake	4.3	681		2691		Lac des Iles
03DS106	10g	Dog River	5.0	739		2691		Lac des Iles
05DS01a	10d	Lac des Iles Baker zone	3.5	678		2689	2689	Lac des Iles
05DS01b	10d	Lac des Iles Baker zone	2.1	646		2689	2689	Lac des Iles
05DS72a	10g	Buck Lake	4.9	719		2690		Lac des Iles
05DS72b	10g	Buck Lake	3.1	667		2690		Lac des Iles
05DS72c	10g	Buck Lake	2.4	652		2690		Lac des Iles
05DS83a	10g	Demars	5.0	649		2690		Lac des Iles
05DS83b	10g	Demars	4.5	656		2690		Lac des Iles
			4.4	**709**	**4.0**	**2690**	**2689**	**Lac des Iles**

$P_{Schmidt}$ represents the barometer of Schmidt (1992), $T_{H\&B}$ represents the thermometer of Holland and Blundy (1994) and $P_{A\&S}$ represents the barometer of Anderson and Smith (1995).

Table 24. Summary of average aluminum-in-hornblende data for crustal blocks and plutonic suites of the central Wabigoon Subprovince area.

$P_{Schmidt}$	$T_{H\&B}$	$P_{A\&S}$	Age	Crustal Block
6.7	718	5.9	2999	Marmion
6.6	727	5.5	2953	Whitton
6.4	716	6.2	2929	central Wabigoon
6.5	712	6.2	2879	Hillyer
5.1	720	4.5	2817	Pinecone / Savoy
6.8	723	6.1	2780	Dog
7.6	718	6.9	2720	Whistle
5.5	723	5.0	2718	West Wabigoon
5.6	732	4.8	2715	Winnipeg River
4.4	709	4.0	2690	Lac des Iles
4.1	703	3.8	2686	mainly biotite granite
4.2	714	3.7	2689	sanukitoid
4.1	676	4.1	2680	Quetico

$P_{Schmidt}$ represents the barometer of Schmidt (1992), $T_{H\&B}$ represents the thermometer of Holland and Blundy (1994) and $P_{A\&S}$ represents the barometer of Anderson and Smith (1995).

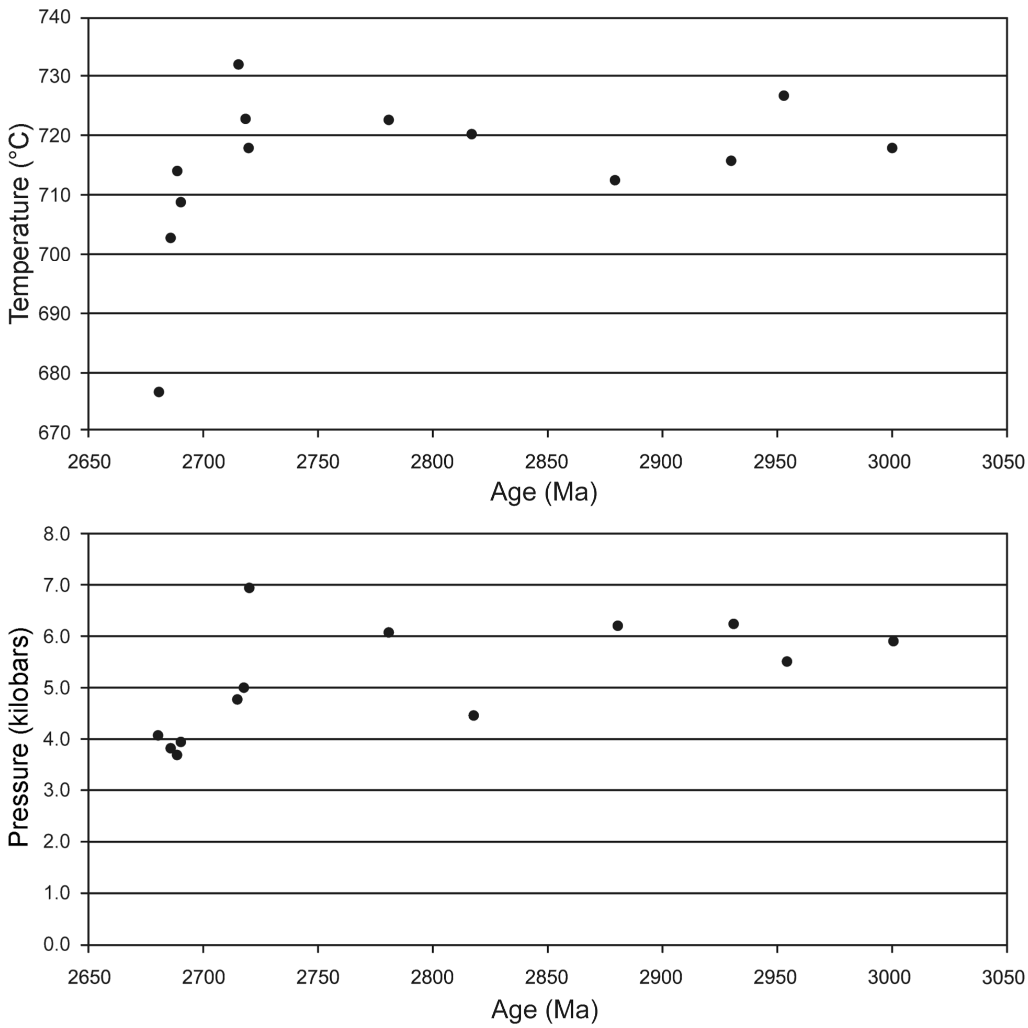

Figure 15. Plots of a) Al-in-hornblende temperature versus time for crustal blocks and late plutonic suites of known age and b) average Al-in-hornblende pressure versus time for crustal blocks and late plutonic suites of known age.

Indeed, Figure 15b shows the overall trend of pressure variation through time. Average crystallization pressures remained approximately constant at about 6 kilobars from 3.0 to about 2.72 Ga after which average pressure decreases rapidly to 4 kilobars for late plutonic suites at about 2690 Ma. Two exceptions are the Pinecone–Savoy and Whistle blocks that have unusually low and high pressures, respectively. The pressure estimate for the Pinecone–Savoy block is based on only 2 measurements and those for the Whistle block are concentrated mainly in the high pressure and possibly uplifted eastern end of the Marmion domain (discussed below). Hence, the exceptions may be due to sampling issues reflecting either a small number of samples or samples concentrated in an anomalous part of the crustal block. Assuming that all plutonic rocks originally crystallized at about the same depth below surface, the data indicate that uplift was minimal from 3000 to about 2750 Ma, after which the present erosional surface was rapidly uplifted from 18 to 12 km within a few tens of million years.

Kamineni, Thivierge and Stone (1988) studied and dated epidote-bearing mineral assemblages in fractures of the Eye–Dashwa pluton (sanukitoid suite) in the mid-central Wabigoon Subprovince area. These authors established on the basis of the stability of mineral assemblages and Rb–Sr systematics of epidote that the fracture fillings developed under pressures of 3 kilobars at about 2300 Ma. In combination with the present data (*see* Table 24), this indicates that the 2700 Ma episode of rapid uplift was short-lived, after which the central Wabigoon Subprovince area resumed very slow uplift that could have continued to the present.

Figure 16 represents a partly contoured map of pressure variation in the central Wabigoon Subprovince area. The map is extremely complex because the area contains many crustal blocks and plutonic suites each of which is of different age (compare Figures 16, 5 and 3). Because each block is of different age, data must be contoured separately within each block, unless it can be shown that the blocks were joined together before pressures were locked in. In many areas such as greenstone belts and biotite-bearing plutonic suites, the pressure data are sparse or absent and no estimation of local crystallization pressures and related contouring is possible with the exception of some posttectonic biotite granite batholiths, which underlie approximately one-third of the area. The batholiths have not been strongly deformed and it is reasonable to extrapolate pressure values from a few widely spaced sites to represent the entire batholith.

In general, Figure 16 shows a similar trend as Figure 15b. The young plutonic areas corresponding with posttectonic batholiths have low pressures (<4 kilobars), whereas areas of older plutonic rocks such as the Marmion, Whitton and Central Wabigoon domains have higher pressures of 5 to 7 kilobars. Three areas of anomalously high pressure can be defined on the basis of at least 2 adjacent measurements of >7 kilobars. These occur in the northern Hillyer domain, eastern Marmion domain and eastern Central Wabigoon domain. High pressures seem to be concentrated at the northeastern margin of the Hillyer domain and imply that this area has been uplifted, possibly due to overthrusting of the Hillyer domain onto the Central Wabigoon domain or displacement of an unseen fault at the margin of the Hillyer domain. Similarly, high pressure in the eastern Central Wabigoon domain (north of Lac des Iles) may be due to displacement on unknown faults at or near the margin with the adjacent Whitton domain. The eastern Marmion domain and part of the Whistle domain (south of Upsala) are enveloped by the crescentic Muskeg Lake batholith and high pressures at this locality are possibly due to the old rocks having been uplifted on top of the intrusive batholith.

The Al-in-hornblende pressure data allude to the importance of vertical tectonics in long-term crustal evolution. The earth’s crust appears to have been progressively uplifted through time with blocks of old crust raised more than adjacent areas of younger crust. The oldest crust may be ultimately eroded away in the event that this process was to continue. The regional uplift is locally accentuated where parts of old crustal blocks have either migrated laterally and overridden other blocks or have been raised by displacements on unseen faults or intrusion of late granites. These local perturbations in the pattern of time-dependent regional uplift are represented by the comparatively small areas of high pressure (>7 kilobars) in Figure 16.

Structure

REGIONAL STRUCTURAL TRENDS AND OVAL STRUCTURES

Regional structural trends defined by lithologic contacts, foliations, gneissosity and faults are aligned mainly easterly to northeasterly in the central Wabigoon Subprovince area (*see* Map P.2229) and indeed in most of the western Superior Province. The easterly trending boundary between the Quetico and Wabigoon subprovinces represents the most regionally extensive structural element in the area. Most structures dip subvertically although local areas of low-dip fabric are observed.

The regional pattern of easterly structural trends is locally perturbed by northerly trending structures and oval structures. For example, the southern Wabigoon Subprovince in the area between Atikokan and Mine Centre has several northerly trending greenstone belts such as the Nevison Lake and Finlayson Lake greenstone belts. Schwerdtner, Morgan and Stott (1985) and Stone, Kamineni and Jackson (1992) described several distinct oval structures including those at Dashwa Lake and at Ash Bay of Rainy Lake. Typically, oval structures have a felsic plutonic core and are mantled by belts of greenstone and gneisses that wrap completely around the core pluton. In the case of the Dashwa Lake oval structure, the Eye–Dashwa pluton represents the felsic core enveloped by the Nevison Lake, Steep Rock Lake and Finlayson Lake belts as well as tonalite gneisses. Many northerly trending greenstone belts represent parts of the flanks of oval structures.

Oval structures of the Wabigoon Subprovince have been studied and variously interpreted as due to diapirism (Schwerdtner, Morgan and Stott 1985; Schwerdtner 1988) or cross folding (Schwerdtner 1989). Diapiric models involve the rise of low-density felsic plutons and gneisses within the cores of oval structures accompanied by subsidence of dense, overlying greenstone material into the marginal troughs. In contrast, the cross-fold model is compatible with a dominantly horizontal strain regime such as might have existed during accretion of small Mesoarchean crustal domains or Neoarchean subprovince-scale blocks. Although the origin of oval structures has not been investigated as part of this work, several factors favour the dominant role of vertical tectonism. These factors include the variable age of the mantling greenstone belts, gneisses and core plutons of oval structures that collectively underwent largely ductile deformation during or after the core plutons were emplaced at about 2686 Ma. This ductile deformation is too late and not easily correlated with the largely brittle mode of deformation associated with late Neoarchean accretion. More likely, oval structures developed due to the rise of core plutons attendant with sagging of marginal dense greenstone material during the major late Neoarchean crustal heating event (Kenoran orogeny).

DEFORMATION EVENTS

Systematic analysis of deformation events has been done in study areas adjacent to the central Wabigoon Subprovince area. Percival et al. (2002) identified 5 deformation events in the north-central Wabigoon Subprovince and Sanborn-Barrie et al. (2002) defined 3 deformation events in the western Wabigoon Subprovince. These include early mineral fabrics that are developed at a grain scale in old rocks and are crosscut by later folds and mineral fabrics all of which are cut by young shear zones. Based on U/Pb titanite ages, Percival et al. (2002) inferred that the earliest deformation events developed at 2913 to 2835 Ma in tonalitic rocks, whereas the late shear zones contain 2677 Ma titanite and evidently postdate most magmatic events in the area.

Although the sequence of deformation events has not been studied in the present area, circumstantial evidence suggests that there are likely to have been numerous events. Firstly, geochronology, summarized in Table 2, shows a variety of U/Pb titanite ages ranging from 2722 to 2657 Ma. The growth of titanite or rims on zircon grains are commonly associated with metamorphic or deformation events and the range of titanite ages implies that there could have been several Neoarchean deformation events. Secondly, Stone and Davis (2006) identified 8 magmatic events in the central Wabigoon Subprovince area using U/Pb zircon ages (*see* Table 2) that are thought to represent crystallization ages. These magmatic events are spaced through the interval of 3000 to 2690 Ma and include the development of early tonalitic complexes, various greenstone sequences and late largely posttectonic batholiths. Deformation is not necessarily associated with magmatism although it is probable that deformation is linked to magmatism at least locally. For example, the aureole of an intrusive batholith is likely to be strained during emplacement of the batholith.

STRUCTURAL YOUNGING DIRECTIONS AND FOLDS

Unconformities where supracrustal rocks overlie basement, sedimentary beds that grade from coarse sandy bases to silty tops and volcanic pillows with convex upper surfaces are common features that help to resolve local younging directions and the structure of supracrustal sequences. What is known about the structure of various supracrustal belts on the basis of these features is summarized below.

Jackson (1985b) interpreted the Lumby Lake greenstone belt to be broadly synclinal due to south-younging pillows in the north and mainly north-younging pillows in the south. Since the work of Jackson, geochronology (*see* Table 2) has shown that Lumby Lake belt is composed of at least 3 major assemblages of supracrustal rocks including the Lumby North assemblage that youngs south, the Lumby South assemblage that youngs north and the central Pinecone assemblage that is probably synclinal. Recent mapping (Buse, Lewis and Magnus 2009) has indicated that the structure of the Lumby Lake belt is more complex than that of a simple syncline and includes 2 juxtaposed and oppositely facing assemblages overlain by a third assemblage.

The Graham greenstone belt is too strongly foliated for identification of local younging directions however geologic and geochronologic evidence suggests that it is composed of 2 assemblages. The western assemblage unconformably overlies 2814 Ma basement (*see* Table 2, No. 151) on the west side of the belt and therefore must be younger than 2814 Ma in age, younging east. In contrast, the eastern part of the belt appears to be intruded by 2929 Ma tonalite (*see* Table 2, No. 24) and, accordingly, the mafic metavolcanic rocks in this area must be older than 2929 Ma although their younging direction is unknown. The boundary between these assemblages possibly extends along the central axis of the sinuous Graham greenstone belt. Likewise, the Phyllis Lake greenstone belt is strongly foliated to the extent that local younging directions are difficult to discern; however, an unconformity along the northwest side of the belt implies that strata, at least in the northwest part of the belt, youngs southeasterly.

The Whitton assemblage of the Heaven Lake greenstone belt is composed of black amphibole gneisses in which pillow shapes are not well defined. East of the present area, Hart, MacDonald and Lepine (2001a) showed a few north-younging pillowed metavolcanic sequences in the Whitton assemblage, which is overlain by the synclinal Whistle assemblage. On the basis of these limited data, much of the Whitton assemblage of the Heaven Lake greenstone belt within the study area would appear to represent a north-younging section of volcanic strata.

The Garden Lake greenstone belt of Hart et al. (2000) contains mainly south-younging metavolcanic strata based on pillow shapes observed at scattered localities. Likewise, mainly south-younging pillowed metavolcanic sequences are mapped in the Lac des Iles belt east of the present area (Hart, MacDonald and Lepine 2001b). These would imply that like the Whitton assemblage of the Heaven Lake belt, the Garden Lake and Lac des Iles belts are largely homoclinal sections of volcanic strata.

The Otukamamoan Lake and Raleigh Lake greenstone belts, as well as numerous greenstone slivers in the western central Wabigoon area, represent extensions of the larger Kakagi Lake–Savant Lake greenstone belt, west and north of the present area. Although the greenstone slivers show no recognizable pillows, the Otukamamoan Lake belt has consistent southwest-younging pillows (e.g., Smith 1993) and appears to represent a homoclinal volcanic sequence. The Raleigh Lake belt also has mainly southwest-younging pillows, although this belt is locally curved around intrusive plutons and is possibly folded.

Rocks of the Mine Centre greenstone belt occupy a wedge between the Quetico and Seine River faults. Poulsen (1986, 2000b) described the Mine Centre belt as a zone of wrenching characterized by a series of east-northeast-trending lenticular lithostratigraphic domains with steep boundary faults. Rocks within each domain have a steep mineral fabric and tight folds of east-northeast trend that have developed in response to northwest compression. Poulsen (2000b) identified several stages of deformation including early recumbent folds followed by upright northeast-trending folds and late development of shear zones and kink bands. Eastward at Calm Lake, the Mine Centre belt lies mainly north of the Quetico fault. A few north-younging pillows are observed at scattered localities and Borradaile (1982) interpreted several steeply plunging, sideways-closing isoclinal folds in metavolcanic rocks west of Calm Lake. The metasedimentary rocks north of the Mine Centre belt at Calm Lake show graded beds with variable younging directions at widely spaced localities (Stone, Hallé, Hellebrandt et al. 2007a). These metasedimentary sequences are probably folded, although there are insufficient observations to define fold axes.

The Lac des Mille Lacs greenstone belt is composed of gabbro dikes and massive mafic flows from Atikokan east to Boot Bay of Lac des Mille Lacs. Although the gabbro dikes are folded on the scale of an outcrop, there are no younging directions from which the structure of the belt can be discerned. At Lac des Mille Lacs, the belt is made up of a 2830 Ma largely intermediate metavolcanic assemblage and the younger Puffy Pillow assemblage of mafic metavolcanic rocks. Local younging directions were not observed in the intermediate metavolcanic rocks, but pillows in the mafic metavolcanic sequence young consistently south. Possibly, the Puffy Pillow assemblage represents an allochthonous largely south-facing homoclinal section.

Much of the central Steep Rock Lake greenstone belt is underlain by the 2735 Ma Witch Bay assemblage. Stone, Kamineni and Jackson (1992) noted that the mafic pillow lavas of this assemblage young consistently eastward over a distance of 5 km and possibly represent a tilted lava cone. The Witch Bay assemblage is probably overlain by the Dismal assemblage, which contains the ashrock. Although no younging directions were identified, the Dismal assemblage apparently also faces east and is in fault contact with the west-facing Steep Rock assemblage. Metavolcanic rocks in the western Steep Rock Lake belt are much older than the Witch Bay assemblage and, on the basis of age (*see* Table 2, Nos. 5 and 12) appear to be extensions of the Finlayson East assemblage. Although variable younging directions were observed in metasedimentary rocks at Little Falls Lake, the structure of the western Steep Rock Lake belt is not well known.

The Finlayson Lake greenstone belt is made up of the 3000 Ma Finlayson East assemblage and the 2931 Ma Finlayson West assemblage whose boundary along the axis of the belt is overlain by the 2735 Ma Witch Bay assemblage. The Witch Bay assemblage is interpreted to represent a klippe based on the deformed nature of its contacts and change in metamorphic grade across the contact. Stone, Kamineni and Jackson (1992) identified several reversals in younging directions derived from pillow shapes within what is now known to be the Witch Bay assemblage. The reversals in younging direction probably indicate that the klippe is folded around several northeast-trending axes.

The Quetico Subprovince shows abundant local younging directions based on graded metasedimentary beds particularly near the weakly metamorphosed north boundary of the subprovince. Near the boundary with metavolcanic rocks, the majority of graded beds in Quetico metasediments young north toward the boundary (*see* Map P.2229) although local areas of south-younging beds are noted. Structure of the Quetico metasediments was studied by Borradaile (1982) and Borradaile et al. (1988) who interpreted a series of isoclinal sheath folds with axial traces plunging at variable angles parallel to the boundary. The zones of folded strata have markedly variable structural facing direction and are separated by zones of high strain or “slides” developed due to large transpressive deformation.

Borradaile (1982) noted marked contrast in structural style between metasedimentary rocks of the Quetico Subprovince and metavolcanic rocks of the Wabigoon Subprovince. Quetico sediments have a strong east-west structural trend with isoclinal folds having vertical flanks, but fold hinges that plunge at variable angles. In the Wabigoon Subprovince, the structural trend of metavolcanic rocks changes so as to wrap around felsic intrusions and strata are deformed by steeply plunging, sideways-closing isoclinal folds. In this survey, large areas of apparently unfolded homoclinal volcanic strata are observed.

FAULTS AND FRACTURES

Supracrustal rocks of the central Wabigoon Subprovince area show remarkable variation in age and have probably been faulted in various deformation events through the Mesoarchean and Neoarchean. Many greenstone belts are composed of several assemblages of metavolcanic rocks and the boundaries between these assemblages can, in places, be marked by faults. Such faults are, however, very difficult to recognize because they have been deformed together with country rocks and, in many instances, it is unclear whether assemblages are allochthonous or autochthonous.

The boundary between the Dismal and Steep Rock assemblages in the eastern Steep Rock Lake greenstone belt is unexposed, but a fault (Dismal fault) is interpreted at this locality because the Dismal and Steep Rock assemblages apparently face in opposite directions and, therefore, are probably juxtaposed tectonically. Likewise, the boundary between the Witch Bay and older metavolcanic assemblages in the Steep Rock Lake and Finlayson Lake belts is interpreted to be a fault due to a few exposures at Little Falls Lake and Finlayson Lake where the boundary is a zone of strong subvertically foliated and altered rock. These assemblage boundaries possibly represent deformed thrust faults, although, in many other situations, the structural relation of one assemblage to another is unclear.

Faults and fractures which are of Neoarchean to Proterozoic age and are unaffected by other deformation events are much more easily recognized. Among late faults of regional extent are 2 sets defined on the basis of their strike, which either ranges from east to east-northeast or is consistently north-northeast. The east- to east-northeast-trending faults include the Quetico, Seine River and Shelby faults that are interconnected at the Quetico–Wabigoon subprovinces boundary (*see* Map P.2229). These faults are characterized by zones of schistose rocks up to several hundred metres wide with mylonite developed sporadically where the faults cut felsic plutonic rocks. Locally, zones of cataclastic rocks and pseudotachylite crosscut the mylonite indicating that fault rocks became brittle during deformation. Peterman and Day (1989) obtained an age of 1947±23 Ma using Rb-Sr systematics for pseudotachylite of the Quetico fault indicating that this structure was active into the Proterozoic.

Parts of the Quetico fault have been described by many people through the years (e.g., Hawley 1930a; Kaye 1967; Shklanka 1972; Mackasey, Blackburn and Trowell 1974; Fumerton 1982; Poulsen 2000a) and the fault has been the focus of 2 theses (Bau 1979; Kennedy 1984). Most agree that faulting postdates other deformation in the area and the Quetico fault has a largely dextral transcurrent displacement; several workers have estimated a displacement of up to 150 km.

In the western Central Wabigoon area, the Quetico and Seine River faults are apparently splays of the same structure. Although Fumerton (1982) proposed that the name “Quetico fault” should be applied to the southern splay because this structure more closely follows the subprovince boundary, other workers (e.g., Poulsen 2000a) have continued to apply "Seine River fault" to the southern splay. The present regional mapping shows that the Quetico fault is discontinuous in the central Wabigoon Subprovince area. Although well developed from Rainy Lake through Calm Lake, the fault dies out west of Atikokan by splaying northward into the western Steep Rock Lake greenstone belt. East of Atikokan, the Quetico fault redevelops at the Quetico–Wabigoon subprovinces boundary and several splays extend northeasterly from the main fault at Crooked Pine Lake. At Lac des Mille Lacs, the Quetico fault continues east into the Quetico Subprovince and the Shelby fault, which is effectively a splay of the Quetico fault, curves east-northeast along the Quetico–Wabigoon boundary. At Wakinoo Lake, the Shelby fault extends northeasterly into the Wabigoon Subprovince leaving the Wabigoon–Quetico boundary not faulted in the eastern part of the central Wabigoon Subprovince area. Based on the discontinuous nature of the Quetico and associated faults, it seems unlikely that these structures can have accommodated large displacements.

North-northeast-trending faults include those at Entwine Lake and the Marmion, Pakashkan and Gull Lake faults of which the Marmion is the largest and best known. The Marmion fault extends north-northeast from the Steep Rock Lake greenstone belt through the western Lumby Lake greenstone belt (*see* Map P.2229). Evidently, the Marmion fault does not cut the Steep Rock Lake belt and, therefore, probably predates the Steep Rock assemblage, the age of which is known to be younger than 2780 Ma (*see* Table 2, No. 50). Within the western Marmion batholith, the Marmion fault is a complex braided structure marked by zones of strongly foliated and fractured tonalite cutting early mylonite (Stone 2008b, 2008c). The fault zone has variably shallow to intermediate south-southwesterly plunging slickenlines (Stone, Kamineni and Jackson 1992) and, although the overall sense of slip is unclear, several normal faults with metre-scale displacements were observed in water tunnels at the Ontario Hydro generating station north of Atikokan. In felsic plutonic rocks north-northeast of the Lumby Lake belt, the Marmion fault is marked by mylonite zones. These parts of the fault possibly represent a somewhat deeper level or earlier phase of the deformation compared to segments of the fault exposed southwestward in the Marmion batholith.

Two north-northeast-trending splays of the Entwine fault crosscut the Entwine stock of the sanukitoid suite and, therefore, are younger than 2685 Ma in age (*see* Table 2, No. 120). These faults clearly offset the contact of the Entwine stock in a sinistral sense by a few hundred metres (*see* Map P.2229). It is unclear, however, if this strike separation represent strike-slip because the dip-slip component is unknown. The faults are marked by 10 m wide zones of cataclasis and abundant neomineralization characterized by epidote fracture fillings within broad zones of intensely reddened and altered rock. The Entwine fault has possibly developed late or at a high structural level as indicated by the largely brittle deformation features in the fault.

The Pakashkan fault crosscuts the Graham belt and adjacent plutonic rocks northeasterly. Although poorly exposed, central segments of this fault in the area east of Pakashkan Lake are characterized by a few hundred metres wide zone of mylonite. The contact of the Pakashkan pluton is offset dextrally by about 1 km; however, it is unknown if this observed strike separation is representative of strike-slip.

The Gull Lake fault comprises 2 highly curved splays that crosscut the Heaven Lake greenstone belt and the Tib Lake gabbro intrusion as well as adjacent plutonic rocks. These splays merge and extend north-northeasterly through the northeast corner of the central Wabigoon Subprovince area (*see* Map P.2229). Although poorly exposed, much of the fault is apparently overlain by diabase sills. Both the Tib and Buck Lake intrusions of the Lac des Iles suite are situated along the Gull Lake fault and possibly the mafic magmas, which fed these intrusions, moved upward through the fault or an early stage of the fault. Late stages of deformation associated with the Gull Lake fault crosscut and dextrally offset the contact of the Tib intrusion as well as the Heaven Lake greenstone belt.

Fracture characteristics of the Eye–Dashwa pluton were studied by Atomic Energy of Canada Ltd. (AECL) between 1979 and 1985 (*see* "Previous Mapping and Exploration: Late Regional Work"; *see also* MRD 242: "References to AECL work at Atikokan" document). The AECL research included mapping and analysis of fractures at surface and in boreholes and showed that the Eye–Dashwa pluton was cut by abundant fractures (mean spacing of 0.3 m) dominated by north-northeast- and west-northwest-trending sets of mesoscopic transcurrent faults (Stone 1984). This conjugate set of mesoscopic faults developed in response to northwesterly shortening, which appears to explain some north-northeast-trending major faults such as the Entwine fault discussed here.

Stone and Kamineni (1982) noted that fractures were sealed with crystalline minerals such as epidote and chlorite although many, particularly those within 300 m of surface, had been rejuvenated and were either open or partly resealed with soft materials such as carbonate or clay. Kamineni and Stone (1983) applied various isotopic dating methods to fracture fillings and determined that many fractures developed in the Paleoproterozoic with several episodes of subsequent rejuvenation. Kamineni, Stone and Peterman (1990) went on to propose that much of the western Superior Province had been affected by significant brittle deformation in the Paleoproterozoic.

STRUCTURAL AND MAGMATIC EVOLUTION OF THE CENTRAL WABIGOON AREA

The study area was historically considered to be part of the Wabigoon and Quetico subprovinces representing, in the context of plate tectonic theory, a volcanic arc and accreted metasedimentary prism, respectively. Later work showed that the Wabigoon Subprovince is more complex than a volcanic arc and is divisible into blocks of older crustal material (Marmion terrane) and younger crustal material (Western Wabigoon terrane; *see* Figure 1). To this work (*see* “Subdivisions of Archean Rocks and Tectonic Evolution of the Western Superior Province: A Review”) is added new insight gained from the present study.

Results of the present study show the central Wabigoon Subprovince area to be extremely complex and characterized by many crustal blocks collectively containing magmatic rocks varying in age from 3.0 to 2.7 Ga. For example, the geochronology of Table 2 is summarized in the form of histograms in Figure 17 and shows on the order of 8 magmatic events. Each magmatic event is a discrete peak in the histogram of age determinations and represents an interval of magmatism during which supracrustal and plutonic rocks developed. The clusters of age determinations are the main basis for defining various terranes, domains and assemblages whose spatial distribution is shown in Figure 3. Among these, only the Bar assemblage of the Lumby Lake greenstone belt is represented by a single age determination and remains somewhat enigmatic in terms of whether or not it is a crustal block.

Various terranes, domains and assemblages of Figure 3 are characterized primarily by rocks of similar age. The extent of each terrane, domain or assemblage is inferred by interpolating between sites of similar ages or by extrapolation from sites of age determinations. In the latter case, geologic features including contacts, zones of high strain and variations in metamorphic grade are used to define the boundaries of terranes, domains and assemblages. In many places, these boundaries are defined only on the basis of meagre geologic information and are likely to change as more information becomes available.

What then, in a tectonic sense, do each of the magmatic terranes, domains and assemblages represent? Geochemistry shows that most plutonic rocks and some volcanic rocks have “arc-like” signatures (e.g., Pearce 1996) represented by trace element profiles that are enriched in incompatible elements with distinct Nb, Ta and Ti troughs such as are found in rocks from modern volcanic arcs. Other volcanic rocks have more-or-less flat trace element profiles representative of modern ocean floor environments. These data imply that magmatic rocks representing the various terranes, domains and assemblages of the central Wabigoon Subprovince area could be parts of magmatic arcs and ocean floors that developed episodically through the Meso- to Neoarchean.

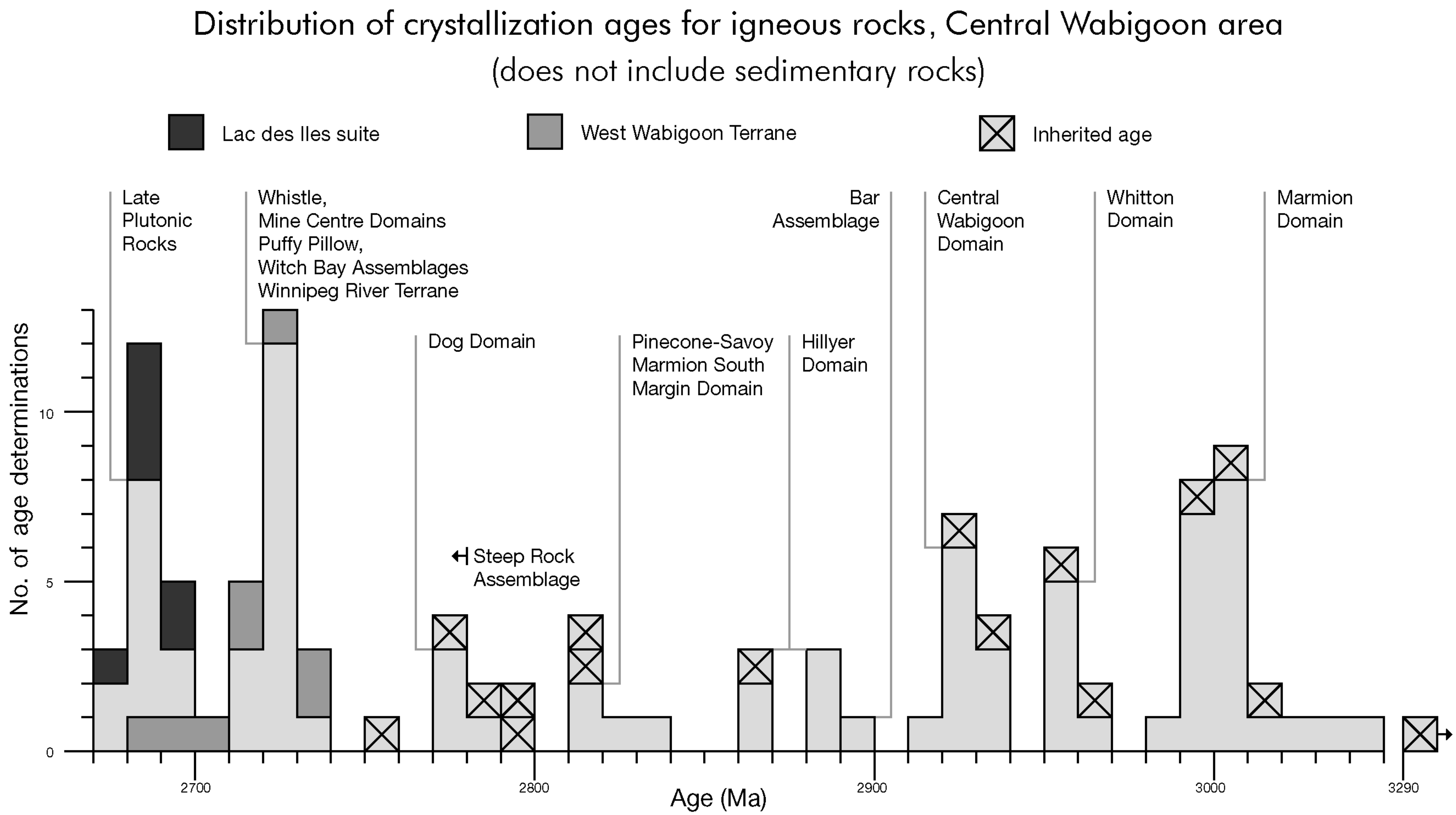

Figure 17. Distribution of crystallization ages for igneous rocks, central Wabigoon Subprovince area.

Stone and Davis (2006) proposed that most domains represent the vestiges of igneous activity that occurred within old magmatic arcs of either oceanic or continental origin. Oldest is the 3.0 Ga Marmion domain, which is composed essentially of the Marmion batholith within slightly older volcanic assemblages of the Lumby Lake, Finlayson Lake and Lac des Mille Lacs greenstone belts. This could represent an early oceanic arc, which was sufficiently large and buoyant to resist subduction. The Marmion domain may have developed into an early microcontinent and acted as a buttress against which younger assemblages and domains were accreted. The progressive outward and northward younging of rocks through the Marmion, Whitton and Central Wabigoon domains is characteristic of large modern continental masses that grew by accretion. Out-of-sequence domains such as the complex Pinecone–Savoy and Marmion South Margin domain may represent rift sequences or overthrust sequences. Unusual linear zones of crosscutting gabbro dikes such as in the Lac des Mille Lacs and Steep Rock Lake greenstone belts are also possible paleo-rift zones.

The Hillyer domain has a highly strained north boundary and high Al-in-hornblende pressures. These features combined with the unusual age of the plutonic rocks within the domain can be explained by a variety of tectonic models in which the Hillyer domain is either a tectonic indentor or intra-thrust sheet of exotic material or alternatively an uplifted and magmatically reactivated part of the Central Wabigoon domain. The presence of older Central Wabigoon material on both the north and south side of the Hillyer domain provides meagre evidence in favour of the Hillyer domain being a sheet of exotic material thrust into the Central Wabigoon domain.

The Dog and Whistle domains are composed of mainly plutonic rocks that have intruded older sequences of the Marmion and Central Wabigoon domains of the present area. Possibly, the Dog and Whistle domains represent magmatic arcs developed on the margin of the older microcontinent. Material similar in age to the Dog domain may have been more extensive than is shown in Figure 3 because it occurs 100 km westward as detrital and inherited grains in the Steep Rock and Dismal assemblages. Several assemblages, including the Puffy Pillow, Witch Bay and Dismal assemblages, comprise mainly pillowed mafic sequences of simatic origin. In contrast to the autochthonous Dog and Whistle domains, these mafic assemblages are probably allochthonous.

The western Wabigoon and Winnipeg River terranes contain isotopically juvenile and recycled material, respectively, that were accreted to the composite Marmion domain at about 2.7 Ga (Tomlinson et al. 2004). The requisite tectonic boundaries between these exotic terranes and the composite Marmion domain have proven difficult to define in the field. For example, Tomlinson et al. (2004) included the Garden Lake greenstone belt with the Winnipeg River terrane because samples of the greenstone belt had negative ε_{Nd} values indicating that the magmas had interacted with older crustal material. Still, no tectonic boundary zone between the Winnipeg River and composite Marmion domain could be mapped in the field, although this boundary is possibly overlain by sheets of late granite.

The Raleigh Lake and Otukamamoan Lake greenstone belts are part of the western Wabigoon terrane in the present area. Numerous greenstone slivers and young plutonic rocks extend east from Otukamamoan Lake as far as the White Otter batholith. This zone of foliated to gneissic rocks (*see* Figure 3, West Wabigoon boundary zone) may represent part of the western Wabigoon terrane or a tectonic mixture of material from the western Wabigoon and Marmion blocks. Alternatively, the boundary zone may represent a deformed magmatic arc that had developed on the west side of the composite Marmion terrane prior to accretion of the western Wabigoon terrane. One sample of tonalite from the boundary zone (*see* Table 2, No. 86) has a young crystallization age and a 3293 Ma inherited zircon grain probably representing material from the Winnipeg River terrane. The significance of this sample and whether or not it indicates that the Winnipeg River terrane extends farther south than shown in Figure 3, is difficult to evaluate with the available information. Possibly detailed Nd isotopic work can help to constrain boundaries between older recycled continental material such as in the Winnipeg River and composite Marmion terranes and juvenile material of the western Wabigoon terrane.

Accretion of Quetico sedimentary sequences and intrusion of voluminous granite batholiths marks the latest major stage of tectono-magmatic evolution of the area. This corresponds with the Central Superior orogeny of Percival et al. (2006) and marks the fairly rapid Neoarchean assembly of the Superior Province from component terranes and superterranes. In the context of the present area, the late Neoarchean tectonism marks an increase by at least an order of magnitude in the scale of crustal elements. For example, the preserved parts of Mesoarchean crustal blocks, such as the Marmion domain, appear to have been of approximately 100 km extent, whereas Neoarchean crustal blocks (e.g., terranes shown in Figure 1) reach more than 1000 km.

Much of the previous discussion has involved the interpretation that Archean crust grew largely by a process in which small crustal elements are created locally, translated horizontally and accreted to form progressively larger crustal elements. This process, which involves modern plate tectonic theory has been widely endorsed for the Neoarchean (e.g., Percival et al. 2006 and references therein) and observations from the present area would appear to indicate that plate tectonism was active as early as 3.0 Ga. In many instances, however, the actual evidence for Mesoarchean plate tectonics is meagre and interpretative. For example, rocks of the Marmion domain have geochemical characteristics of rocks that formed in modern volcanic arc environments, but this does not necessarily prove that volcanic arcs existed at 3.0 Ga because rocks chemically similar to modern arc-rocks could possibly be generated in other environments that existed at the time. Likewise, a layered and progressive outward younging sequence of domains such as from the Marmion through Whitton to Central Wabigoon domains could possibly be achieved through mechanisms other that plate tectonism. Hence, the available evidence supports the interpretation, but does not necessarily prove that plate tectonism was active as early as 3.0 Ga.

The Al-in-hornblende barometry provides an exciting additional perspective on Archean tectonism. The barometry cannot measure horizontal motion, but where combined with geochronology can effectively characterize vertical motion of the crust through time. The overall trend, perhaps best shown in Figure 15b is that the plutonic rocks within various crustal blocks crystallized at 5 to 6 kilobars pressures prior to about 2.72 Ga and about 4 kilobars near 2.7 Ga. These data are interpreted to indicate a period of rapid uplift at about 2.7 Ga and, in consideration with the rocks having been uplifted to surface at present, the data indicate that old crust is progressively uplifted and eroded through time. The period of rapid uplift is not easily explained except to note that it corresponds with the apparent order-of-magnitude growth in the size of crustal blocks. Possibly the development and horizontal interaction of these large crustal blocks at about 2.7 Ga led to correspondingly large uplifts in the obducted material.

Figure 16 shows the spatial distribution of pressure data. The oldest domain (Marmion domain) has possibly been tilted or uplifted in the east as shown by low pressures in the west and high pressures in the east. The areas of high pressure (>7 kilobars) indicate parts of domains that were tilted or bent upward in relation to the domain as a whole. These uplifted areas are small segments of domains and seem to be localized near the boundaries of old crustal domains. Some uplifted areas possibly represent parts of old domains that were overthrust onto other domains in an accretionary plate tectonic environment. After 2.7 Ga, plate interaction occurred at a scale larger than the present area and Figure 16 shows no details of this late-stage and large-scale tectonism except for the widespread, slow but relentless rise and erosion of the Archean crust.

Economic Geology

The central Wabigoon Subprovince area has a long history of prospecting, mineral exploration and mining. Gold, base metals and iron have been historically sought and mined at Mine Centre and Atikokan (*see* "Previous Mapping and Exploration") with the range of commodities and area of search broadened in recent decades. Platinum group metals, rare metals and diamonds are added to the list of sought-after commodities and most major greenstone belts and some plutonic areas have been prospected for these as well as gold and base metals.

Several excellent descriptions have been published for various gold, base metal and iron properties. Descriptions typically include the location, general geology, mineralization and the history of exploration or mining at each property. In some instances, properties are classified according to type of deposit and recommendations for exploration are made. For example, Beard and Garratt (1976) described many gold occurrences at Mine Centre; and Poulsen (2000a, 2000b) provided detailed descriptions of more than 80 gold and base metal occurrences, prospects and past-producing mines. Similarly, at Atikokan, many gold occurrences are described by Wilkinson (1982) and Schnieders and Dutka (1985). Iron deposits and iron mining in the Steep Rock Lake area are reviewed by Shklanka (1972) and McIntosh (1972). Indeed, all reports of previous detailed geologic surveys in the central Wabigoon Subprovince area reviewed local exploration and mineral potential.

In this section, the mineral commodities of economic interest are divided into groups (e.g., "Gold", "Iron", etc.), each of which is briefly reviewed. Rather than repeat the descriptions of all old properties, an effort is made to review important properties within each group and to provide new data in aid of exploration. The new data include details of bedrock geology based on mapping, geochronology and geochemistry, assays of mineral showings and soil surveys over mineralized areas.

GOLD

Gold was the first commodity to be mined in the central Wabigoon Subprovince area and continues to be the focus of active exploration. Mine Centre and Atikokan were the main sites of early gold mining and, although many small mines went into production briefly, the total amount of gold produced was not large (Table 25). Poulsen (2000b), Wilkinson (1982) and Schnieders and Dutka (1985) provided detailed descriptions and assays of the past-producing mines and occurrences in these areas. Current mineral exploration is focussed mainly on the old mining camps with Q-Gold Resources evaluating the Foley Mine in the Mine Centre area (www.qgoldresources.com). VenCan Gold Corporation (now RedPine Exploration Inc.) recently explored the area of the former Fern–Elizabeth gold mine west of the Steep Rock Lake greenstone belt (www.redpineexp.com, *see under* "Moffatt property") and Brett Resources is evaluating the Hammond Reef. A resource of 259.4 million tonnes of ore grading 0.8 g/t Au is defined by drilling at the Hammond Reef property (Brett Resources Inc., news release, November 12, 2009, www.brettresources.com [accessed April 14, 2010]).

In the Mine Centre area, many past-producing gold mines and occurrences are situated within elongate units of biotite tonalite and metavolcanic rocks on opposite flanks of the Bad Vermilion gabbro complex (*see* Stone, Hallé, Hellebrandt et al. 2007b). In most instances, the gold occurs in quartz veins within host rocks, which have been altered to an assemblage of greenschist-facies minerals including quartz, albite, chlorite and carbonate. Metavolcanic rocks and tonalite within the Mine Centre greenstone belt are the same age within error (*see* Table 2, Nos. 70 to 72) and the metavolcanic rocks appear to have been intruded by the undated Bad Vermilion gabbro complex. The available geochronology permits the interpretation that the thermal aureole of the Bad Vermilion gabbro may be responsible for greenschist-facies metamorphism and gold mobilization within the adjacent plutonic and metavolcanic rocks.

Table 25. Mines with significant production within or near the central Wabigoon Subprovince area.

Mine Name	UTM Easting (m)	Location Northing (m)	Commodity	Production (tonnes milled or produced)	Grade	Active Years
Lac des Iles	309200	5449000	Pd, Pt, Au, Ni, Cu	39 million t milled	2.27 g/t Pd, 0.22 g/t Pt, 0.15 g/t Au, 0.061% Cu, 0.074% Ni	1993–
Shebandowan (south of Lac des Mille Lacs)	702900	5386000	Ni, Cu, Pd, Pt, Au	7.9 million t milled	2.06% Ni, 1.0% Cu, 2.7 g/t PGM+Au	1971–1998
Steep Rock	599500	5408000	Fe	99.7 million t milled	59.6% Fe	1944–1979
Caland	602000	5408000	Fe		57% Fe	1960–1979
Canadian Charleston	600500	5403200	Fe (placer)	6.08 million t milled	up to 30% Fe	1958–1965
Atikokan Iron	626400	5404600	Fe	82 300 t shipped	60% Fe	1907–1913
Port Arthur Copper	523200	5400600	Cu	12 t Cu produced	up to 3% Cu	1916–1917
Golden Star	528500	5399300	Au, Ag	0.305 t Au produced	19.2 g/t Au	1898–1900
Foley	526000	5394200	Au, Ag	0.122 t Au produced	5.14 g/t Au	1895–1900
Olive	520500	5401900	Au, Ag	0.10 t Au produced	16.9 g/t Au	1896–1900, 1941
Cone	526500	5395000	Au	900 t milled	34–51 g/t Au	1948–1956
McKenzie–Gray	523800	5392500	Au	409 t milled	5.35 g/t Au	1990, 1995
Sapawe	618300	5405000	Au, Ag	38 009 t milled	5.09 g/t Au	1964–1966
Sunbeam	625700	5413300	Au, Ag	589 t milled	14.7 g/t Au	1904
Sawbill	614700	5422900	Au	2192 t milled, 2679 t milled	8.6 g/t Au, 6.51 g/t Au	1897–1899, 1940–1941
Hammond Reef	613600	5422100	Au	2072 t milled	7.2 g/t Au	1897–1898
Elizabeth	593000	5405600	Au	45 t milled	13.71 g/t Au	1912
Harold	591300	5403700	Au	1026 t milled	20.2 g/t Au	1895–1896

Sources of information: Hawley (1930a), Poulsen (2000a), Schnieders and Dutka (1985), Shklanka (1972), Wilkinson (1982), MacTavish (1999); C. Ravnaas (Resident Geologist Office–Kenora, OGS, personal communications, 2007–2009). Lac des Iles Mine: North American Palladium Ltd., www.napalladium.com. Hammond Reef Mine: Brett Resources Inc., www.brettresources.com.

At Atikokan, gold occurrences are clustered east and west of the Steep Rock belt. To the east, many gold occurrences are situated within a crescentic domain comprising the northwestern, western and southwestern flanks of the tonalitic Marmion batholith and adjacent metavolcanic and/or gabbroic rocks of the Finlayson Lake and Lac des Mille Lacs greenstone belts (*see* Map P.2229). Within this crescentic domain, the tonalite is altered to an assemblage of greenschist-facies minerals and is locally cut by deformed gabbro dikes and shear zones. A particularly large and complexly braided fault system of unknown age and kinematics (Marmion fault; *see* Stone 2008b, 2008c) extends along the northwest margin of the Marmion batholith and hosts several major gold occurrences such as the Hammond Reef and Sawbill. Drilling suggests that the gold mineralization and probably the Marmion fault have an intermediate westerly dip at the Hammond Reef property (Brett Resources Inc., www.brettresources.com).

Available geochronology suggests that mineralized rocks of the Marmion batholith and flanking assemblages of the Lumby Lake, Finlayson Lake and Lac des Mille Lacs greenstone belts are 3.0 Ga in age (*see* Table 2, Marmion Domain). In contrast, the eastern Steep Rock Lake greenstone belt, including the Steep Rock assemblage, is younger than 2780 Ma in age (*see* Table 2, Nos. 50 and 51). The age of the Marmion fault, and potentially related alteration and gold mineralization, is unknown except that the fault appears to be overlain by and, hence, at least partly predates the Steep Rock assemblage (*see* Stone 2008b).

West of Atikokan, gold occurs in intrusive and metavolcanic rocks at the west side of the Steep Rock Lake greenstone belt. Tonalitic intrusive rocks at the Harold Mine and Elizabeth Mine were cut by gabbro dikes, after which they were deformed, locally altered to greenschist-facies minerals and cut by

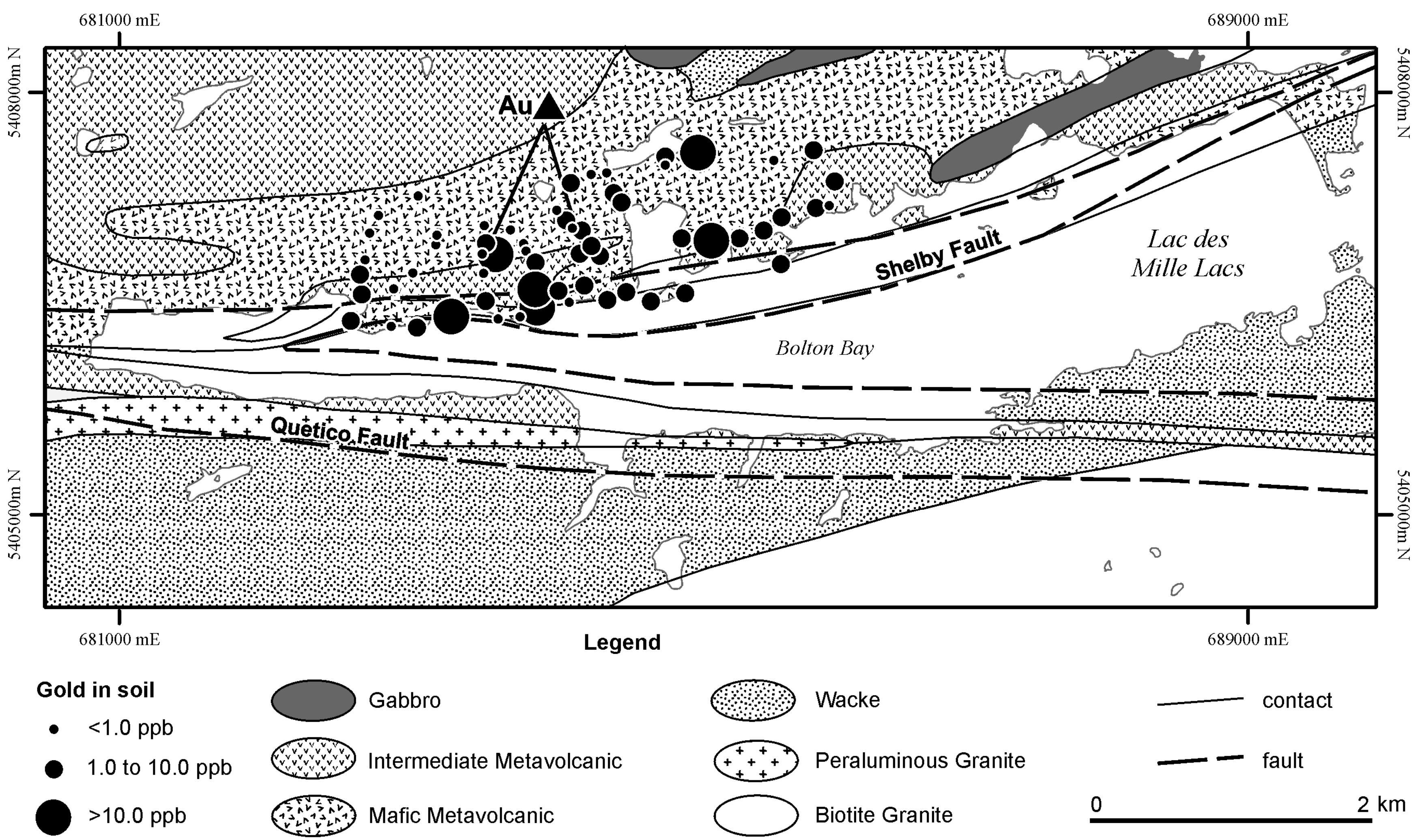

Figure 18. Distribution of gold in soil samples of the Bolton Bay area.

mineralized quartz veins. Gold mineralization at the Harold Mine occurs within tonalitic rocks cut by the Ear Lake fault and, hence, the mineralization is probably related to deformation imposed by the Ear Lake fault. In contrast, deformation and gold mineralization at the Elizabeth Mine is outside of the Ear Lake fault and is possibly related to strain, which was localized at the contact of the greenstone belt or intrusive gabbro masses (*see* Map P.2229). Volcanic rocks of the western Steep Rock Lake greenstone belt are 3.0 Ga in age (*see* Table 2, Nos. 5 and 12). The ages of the tonalitic rocks and gabbro dikes are unknown except that they are younger than the 3.0 Ga volcanic rocks. Thorpe (2008) lists lead isotope model ages ranging from 2511 to 2832 Ma for galena associated with gold mineralization at several gold occurrences including the Hammond Reef, Harold and Elizabeth mines. Although these ages are difficult to interpret, they possibly indicate multiple gold mineralizing events in the area.

Soil surveys were done over 3 areas (Bolton Bay of Lac des Mille Lacs, Crooked Pine Lake (Sapawe) and Harold Lake), each of which contains known gold occurrences. The surveys were done to test the usefulness of soil surveys for detecting gold anomalies and also to provide new targets for gold exploration. Two principal types of soil surveys and related analyses are discussed below. These include direct chemical analysis of soil (usually till) for a range of metals or the separation of discrete grains of heavy minerals from the soil followed, in some cases by analysis of the mineral grains. In the former case, approximately 1 kg soil samples are taken from near bedrock at 0.5 to 1.0 m depth and screened to <63 µm size at the Geoscience Laboratories, Sudbury. Aqua regia digestion is done on 3.0 g of the <63 µm fraction after which the solution is analyzed for a range of metals at ppm level by inductively coupled plasma optical emission spectroscopy (ICP–OES). A 50.0 g aliquot of the <63 µm fraction is prepared using low-level sample methods for analysis of Au, Pt and Pd at ppb level by Ni-sulphide fire assay. The results are provided separately (*see* MRD 242: Tables 27, 28 and 29 for the Harold Lake, Bolton Bay and Sapawe areas, respectively), but are discussed below.

Heavy mineral separations involve 10 kg samples of till, typically collected from just above the bedrock surface at 0.5 to 1.0 m depth. The samples were processed by Overburden Drilling Management Limited, Nepean, Ontario, by passing the material over a shaking table to separate heavy sand-sized mineral grains (e.g., garnet, magnetite, gold) from other largely light (e.g., quartz, feldspar) mineral grains. The heavy mineral grains are further concentrated by magnetic separation and floating the material in heavy liquids after which the resulting concentrate of heavy minerals is hand picked for gold-grains and indicator minerals of metamorphosed massive sulphide deposits and kimberlite. Based on the number and size of gold grains (*see* MRD 242: Tables 27, 28 and 29, "Number of visible gold grains"), the gold content of the sample can be estimated (*see* MRD 242: Tables 27, 28 and 29, "estimated ppb visible gold").

At Bolton Bay of Lac des Mille Lacs, a soil survey was done over an area underlain by 2 known bedrock gold occurrences within mafic and intermediate metavolcanic rocks at the north side of the Shelby Fault (Figure 18). The mineralization occurs within strongly foliated metavolcanic rocks cut by an irregular stockwork of quartz veins (Scott, Magee and Komar 2006). Among the 64 samples of soil processed directly for gold and other metals, 4 samples were also processed on a shaking table and the number of visible gold grains counted with the parts per billion of Au estimated from the size and number of observed gold grains. In the four 10 kg samples processed on the shaking table, the number of visible gold grains varies from 0 to 15 and correlates well with the measured Au (1.1 to 20 ppb) in a 1 kg aliquot of the same samples (*see* MRD 242: Table 28, compare "estimated ppb visible gold" and "Number of visible gold grains").

The amount of gold in the various soil samples is shown by proportional dots in Figure 18. Anomalous gold occurs in soil over or down-ice (south-southwest) from the western gold occurrence, but the eastern occurrence is not clearly identified in the survey. Anomalous gold (>10 ppb) was found at several other localities in the survey area; however, there is no gold mineralization known in bedrock at these localities. Generally, high gold values seem to be found in samples taken over the Shelby fault.

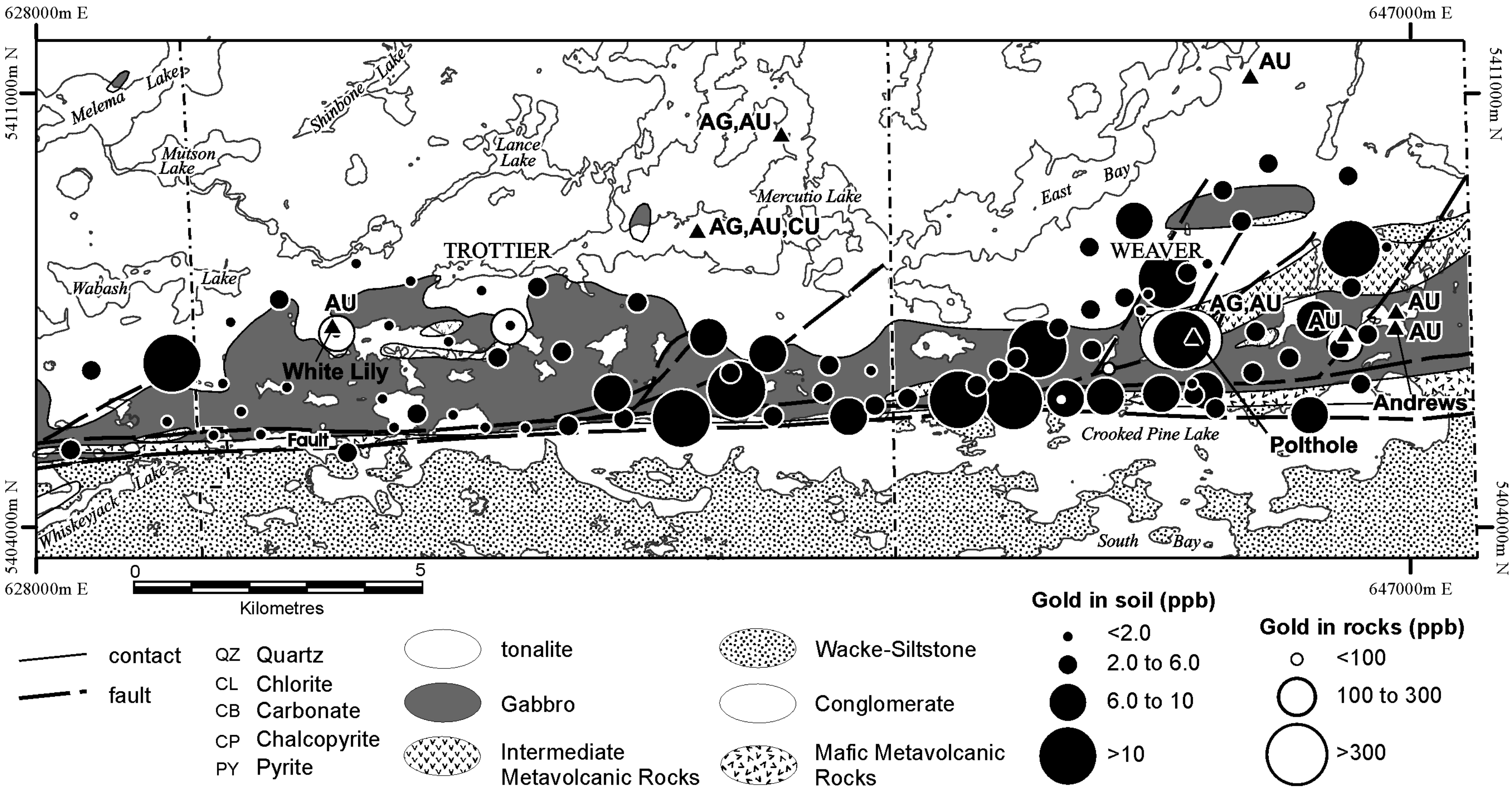

Figure 19. Distribution of gold in soil samples and in bedrock of the Crooked Pine Lake area.

These imply that either the deformed rock contains more gold than less deformed rock or that gold is more readily liberated from the deformed rock than less deformed rock. A more detailed survey (closely spaced samples) might help to define a relation between elevated gold values in soil and discrete sources of gold in the underlying bedrock.

A second soil survey was done in the Crooked Pine Lake area east of Sapawe. In this area, the Quetico fault separates wacke–siltstone sequences to the south from mainly gabbro to the north (Figure 19). Thin units of mafic and intermediate metavolcanic and metasedimentary rocks occur within the gabbro, which intrudes tonalite of the Marmion batholith to the north. Available geochronology suggests that the gabbro, intermediate metavolcanic and tonalite are all about 3.0 Ga (*see* Table 2, Nos. 11, 17 and 1, respectively). The area contains several gold occurrences including the White Lily, Pothole and Andrews. Regional mapping (Stone 2005a) suggests that many gold showings occur along northeasterly trending secondary faults connected to the Quetico fault (*see* Figure 19). The present soil survey was designed to test if gold anomalies occur in soil over or down-ice from the known occurrences or secondary faults.

Among the 86 soil samples of 1 kg size analyzed for ppb gold and other metals, 5 samples of 10 kg size were also processed on a shaking table where the number of gold grains were counted (*see* MRD 242: Table 29). Here, the correlation between the measured gold and number of observed gold grains is poorer than at Bolton Bay(*see* MRD 242: Table 29, compare “estimated ppb visible gold” and “Number of visible gold grains”). The poor correlation may occur due to the “nugget effect” of gold where the presence or absence of one or more gold grains can distort the ppb Au measured from the silt-sized fraction or the number of grains observed in the sand-sized fraction.

The ppb Au in soil and in a few samples of mainly quartz-bearing bedrock is shown by proportional dots in Figure 19. High gold values in bedrock correlate well with high gold values in soil at the Pothole occurrence, but the correlation is poorer at the Andrews occurrence and there is incomplete data for a comparison at the White Lily occurrence. Generally, high gold values seem to occur in samples of soil taken over the eastern part of the Quetico fault and some northeast-trending splay faults, although the data is sparse for an accurate assessment. In this area, a grid of more closely spaced samples is needed to accurately define gold anomalies in soil associated with bedrock gold occurrences and secondary faults. Nonetheless, the available data suggest that segments of the Quetico fault and some northeast-trending secondary faults are the loci of gold occurrences. These faults are mesoscopic-scale features that can be used to guide gold exploration.

A large soil survey was conducted in the Harold Lake area, which spans the western margin of the Steep Rock Lake greenstone belt and includes 2 past-producing gold mines (Harold and Elizabeth), as well as gold occurrences such as the Zephyr and Rebair. In this area, 298 soil samples of 1 kg weight were collected for direct chemical analysis of gold and other elements in the silt-sized fraction. At 42 of these sample sites, large samples of 10 kg weight were also collected and processed on a shaking table where the visible gold grains were counted. The correlation between the ppb Au measured by chemical analysis of the silt-sized fraction and the number of sand-sized gold grains observed in the 10 kg sample is poor (*see* MRD 242: Table 27, compare “estimated ppb visible gold” and “Number of visible gold grains”). Here again, the presence or absence of a small number of gold grains in a sample could distort the results of the chemical analysis and lead to poor correlation with the gold grain count. Hence, isolated anomalous gold values need to be interpreted with caution.

The gold content of soil samples in the Harold Lake area ranges up to a few hundred ppb; this is shown by proportional yellow dots in Figure 20 (back pocket). The number of gold grains observed in the 10 kg samples is also depicted in Figure 20 and is shown by proportional red dots. A cluster of anomalous gold values (yellow dots) occurs in soil samples in the area of the Elizabeth Mine. The distribution of these high values extends north from the Elizabeth veins and implies that the gold occurs mainly at the east margin of an oval embayment of plutonic rocks into the greenstone belt. A diffuse zone of somewhat anomalous gold values is indicated by both the chemical analysis of silt (yellow dots) and number of observed sand-sized grains (red dots) between Modred and Harold lakes extending up to 1 km north of Harold Lake (*see* Figure 20). This zone spans the greenstone–tonalite contact and contains a few showings marked by trenches, but otherwise does not have major known gold occurrences in bedrock. This area is worthy of more exploration for gold in bedrock.

The immediate area of the Harold Mine could not be sampled due to steep local topography and disturbance of the soil by early mining activity. Hence, this area is essentially devoid of soil data. Similarly, only one sample of each type was collected over the Zephyr showing and these samples showed anomalous gold. Two 1 kg samples of soil over the Rebair veins produced only low gold values.

Some 10 kg samples from the southwestern part of Figure 20 seem to show anomalous numbers of gold grains compared with the ppb gold measured in 1 kg samples from the same area. The anomalous gold grain counts come mainly from samples of beach sand taken at the shoreline of the Seine River. Possibly the gold grains have been concentrated in the beach sand or these samples contain a higher proportion of sand-sized particles, either of which could result in higher gold grain counts. Further sampling would be needed to confirm anomalies in this area.

Perhaps the most areally extensive concentration of gold occurs in 1 kg soil samples from the southeast corner of the area more-or-less north of Lower Steep Rock Lake and extending between 2 northerly trending power lines (*see* Figure 20). This area is underlain by mafic metavolcanic rocks cut by three narrow units of quartz-porphyritic tonalite with lenses of conglomerate. These units are strongly carbonate altered and cut by quartz veins to the extent that it is difficult to identify the rock. The area contains no known gold occurrences, although a few trenches are observed. The carbonate-quartz zones, sampled during regional mapping (*see* MRD 242: Table 4, samples 06DS53 to 06DS67 and 07DS05 to 07DS16), contain up to a few hundred ppb Au with the higher gold values generally occurring in samples where sulphide is present.

No 10 kg samples were collected from the area north of Lower Steep Rock Lake for comparison with the results based on chemical analysis of the silt-sized fraction of 1 kg samples. The gold is probably associated with the carbonate alteration zones that occur more-or-less along or north of the Seine River (*see* Figure 20) and also extend northeasterly to the west of Steep Rock Lake (*see* Stone 2008a). These carbonate alteration zones need to be carefully prospected for seams containing even trace amounts of sulphide minerals. Gold normally occurs where sulphide minerals such as pyrite are present.

During the 1995 field season, a regional till sampling survey was done mainly over the area covered by the Sapawe map sheet (*see* location in Figure 2). The 32 samples of approximately 10 kg weight were screened to a 0.25 to 1.0 mm size fraction which was processed on a shaking table and the number of kimberlite indicators and gold grains counted. The goal of the survey was mainly to establish background numbers of kimberlite indicators in soil of the area at an early stage of the project. The results (Table 26) show only one kimberlite indicator mineral and a few gold grains in the till samples. The number of gold grains has probably been biased downward because small gold grains (<0.25 mm) were removed from the sample. Hence, the counts of gold grains from the 1995 regional sampling survey are not directly comparable to the survey results from Bolton Bay, Crooked Pine Lake and Harold Lake areas.

Table 26. Analyses of till samples collected in the Sapawe area in 1995 with the number of gold grains and kimberlite indicator minerals.

Sample Number	Area	UTM Location		Description	Visible Gold		No. of Kimberlite Indicator Grains				
		Easting (m)	Northing (m)		No. grains	ppb Au	Pyrope	Eclogitic Garnet	Chrome Diopside	Ilmenite	Chromite
95DST001	Singleton Lake	615700	5406600	Light brown silty till or gravel beneath 0.3 m silt; no obvious stratification on lee side of outcrop	0	0	0	0	0	0	0
95DST002	N Crooked Pine	636800	5407100	Grey-rusty indurated pebbly till beneath 0.3 m silt on foreside of bedrock knob; probably lake-modified ablation till	1	46	0	0	0	0	0
95DST003	SW Whiskey Jack	624500	5403200	Rusty, stony ablation till draped on outcrop beneath 0.2 m silt, sample from 0.6 m depth	0	0	0	0	0	0	0
95DST004	S Whiskey Jack	628800	5402600	Rusty silty till at 0.5 m depth on bedrock in area of flat outcrop knolls	0	0	0	0	0	0	0
95DST005	Kawene	633000	5402100	Stratified sand and pebbly, silty gravel beneath 0.5 m silt on foreside of outcrop possibly lake-modified till washed down outcrop surface	1	2	0	0	0	0	0
95DST006	S Crooked Pine	636800	5403500	Washed coarse sand on bedrock at 0.6 m depth beneath 0.3 m silty till and 0.3 m silt on flat outcrop	0	0	0	0	0	0	0
95DST007	N Elbow Lake	642800	5402200	Very silty, bouldery red till at 0.5 m depth on high rock ridge on bedrock	0	0	0	0	0	0	0
95DST008	NE Elbow Lake	647800	5402400	Sand beneath 0.3 m silt in crevasse on lee side of outcrop	1	40	0	0	0	0	0
95DST100	Mercutio Lake	637800	5410000	Ablation till (silty, no cobbles) 0.5 m depth on lee side of round outcrop	0	0	0	0	0	0	0
95DST101	E Mercutio Lake	645500	5410700	Pebbly grey till beneath 0.5 m silt on flat top of high bedrock ridge	0	0	0	0	0	0	0
95DST009	Melema Lake	632300	5412100	grey, pebbly, indurated till 0.8 m depth beneath 0.5 m silt in large crevasse on outcrop	0	0	0	0	0	0	0
95DST103	N Kekush Lake	629500	5423700	Lake-modified ablation till, pebbly, some cobbles 10–15 cm in diameter at base of large rounded outcrop, depth of 0.7 m	0	0	0	0	0	0	0
95DST104	NW Caskill Lake	634800	5416700	Pebbly till, no cobble, on top large rounded outcrop, 0.7 m; lake modified or wind blown	0	0	0	0	0	0	0
95DST105	E Bickford Lake	640800	5426200	Silty till, 5–10% angular to subrounded pebbles also present; probably lake modified	0	0	0	0	0	0	0

Table 26. continued.

Sample Number	Area	UTM Location		Description	Visible Gold		No. of Kimberlite Indicator Grains				
		Easting (m)	Northing (m)		No. grains	ppb Au	Pyrope	Eclogitic Garnet	Chrome Diopside	Ilmenite	Chromite
95DST010	Byers Lake	646300	5415000	Beach sand; probably derived from glaciofluvial outwash	0	0	0	0	0	0	0
95DST200	Road 16-4-2	643000	5423200	Ablation till, possibly lake modified, on rounded outcrop, depth of 0.5 m	0	0	0	0	0	0	0
95DST201	Road 16-4-2	646700	5419500	Pebbly, brown-grey till from a large pebbly crevasse on the surface of a rounded outcrop at 0.5 m depth beneath 0.2 m of silt	0	0	0	0	0	0	0
95DST011	S Lumby Lake	624600	5427400	Sandy, pebbly grey indurated till on bedrock beneath silt in crevasse on high outcrop	0	0	0	0	0	0	0
95DST012	N Companion Lake	627100	5419600	Grey, pebbly till at 0.6 m depth on bedrock beneath silt on lee side of flat outcrop	0	0	0	0	0	0	0
95DST013	S Companion Lake	627000	5413800	Grey compact stony till on bedrock 0.8 m below silt (0.4 m) on lee of large outcrop	0	0	0	0	0	0	0
95DST014	S Melema Lake	626800	5407500	Red-grey sandy till on bedrock at 0.5 m depth beneath 0.3 m silt on flat outcrop	0	0	0	0	0	0	0
95DST015	E Belmore Lake	636100	5421300	Red-grey gravel till on bedrock 0.5 m depth in area of boulder ablation till on flat outcrop	0	0	0	0	0	0	0
95DST016	Seine Bay Marmion Lake	621400	5419600	Grey sandy till at 0.7 m depth on boulder on top of flat outcrop	0	0	0	0	0	0	0
95DST106	Trap Bay, Marmion Lake	615400	5421300	Red-grey pebbly till 1 m depth on top of rounded outcrop	0	0	0	0	0	0	0
95DST107	Lower Sawbill Bay, Marmion Lake	610800	5420000	Pebbly, reddish pebbly, sandy till 1 m depth between 2 rounded outcrops	0	0	0	0	0	0	0
95DST018	Central Reserve Bay, Marmion Lake	620800	5410400	Red-brown silty, cobble till 0.8 m depth on bedrock? Lee side of large outcrop	0	0	0	0	0	0	0
95DST019	N Reserve Bay, Marmion Lake	620600	5414600	Grey silty till, 0.9 m on flat outcrop	0	0	0	0	0	0	0
95DST017	S Wilson Lake	637800	5427900	Silty till	0	0	1	0	0	0	0
95DST020	W Sapawe Lake	616950	5402750	Till	0	0	0	0	0	0	0
95DST021	S Lower Seine Basin	612000	5404400	Till	0	0	0	0	0	0	0
95DST108	E Marmion Lake Reservoir	611100	5413900	Till, on top of rounded outcrop, 0.9 m deep, base bedrock	0	0	0	0	0	0	0
95DST109	NE Sawbill Bay, Marmion Lake	617000	5425500	Sandy till, 0.7 m deep, behind pebbly and cobbly beach, base of hole dominated by gray clay, lake modified till	0	0	0	0	0	0	0

IRON

The mining of iron was once an important activity in the central Wabigoon Subprovince area with production coming from the Atikokan Iron Mine, near Sapawe, in 1907 to 1911 and from the Steep Rock and Caland mines near Atikokan in 1944 to 1979 (*see* Table 25; *see* "Previous Mapping and Exploration"). The Atikokan Iron Mine exploited one of several magnetite-bearing deposits associated with high-iron ultramafic rocks of the Dismal assemblage, whereas goethite-hematite deposits, probably representing weathered Archean iron formation were mined from the Steep Rock assemblage near Atikokan. In both cases, substantial iron reserves remained after mining ceased, although it is uncertain whether these can be economically exploited in the future.

The Atikokan Iron Mine represents one of several sulphide-rich magnetite iron deposits within the Dismal assemblage between Atikokan and Sapawe (Atikokan River deposits of MacTavish 1999). These include the Iron Mountain (Atikokan Iron Mine), Shepherd, Pattison–Roberts, Quinn and Garland deposits all of which have been explored and shown to contain reserves of a few million tons of ore grading about 50% Fe with minor Cu and Co (MacTavish 1999). The Atikokan River deposits are herein correlated with pyrite lenses that occur within the Dismal assemblage of the Steep Rock Lake greenstone belt at or near the contact of the Dismal Ashrock and Jolliffe Orezone Formations in the Steep Rock Group (Wright 1959; Jolliffe 1966; Shklanka 1972). Jolliffe (1966) considered the pyritic material to be sedimentary in origin and went on to interpret the zone of pyritic lenses as a member of the Orezone Formation; however, Wilks and Nisbet (1988) showed that the pyritic material is associated with the ultramafic rocks of their Dismal Ashrock Formation. The pyrite lenses are distinguished from the Atikokan River deposits mainly by a larger proportion of sulphide versus oxide-facies iron minerals. In total, however, the Atikokan River deposits and pyrite lenses are probably too small and refractory to be economically exploited as a source or iron. The future development of these deposits may depend upon the potential to economically extract other metals, such as Co and Cu in addition to iron.

The Steep Rock and Caland mines near Atikokan were initially developed during the Second World War by a tremendous engineering effort that involved diverting rivers, draining lakes and pumping lake-bottom sediments from the top of the orebodies (*see* "Previous Mapping and Exploration"). Production ceased in 1979 due to a combination of slope failure at the Hogarth pit and the open pits otherwise approaching their maximum depth of extraction. Various attempts at underground mining proved uneconomical in part due to the low strength and high permeability of the soft and porous Jolliffe Ore Zone Formation.

Shklanka (1972) reported potential ore reserves of 478 000 000 tonnes of iron ore, calculated to a depth of 520 m at the Steep Rock and Caland mines in 1967. Substantial iron ore reserves are supported by deep drilling, such as the drill hole E11-10 (Shklanka 1972) that intersected the iron ore, averaging 49% Fe over 70 m more than 600 m below the Errington pit. Despite large reserves and a high iron-content, the potential for further iron mining in the Steep Rock Lake greenstone belt may be low because near-surface deposits potentially exploited by open-pit mining are largely depleted. A possible exception is the H-orebody under the west arm of Steep Rock Lake, although the grade and tonnage of iron ore in this area is unknown to the author. Likewise, the extraction of iron by underground mining methods has been unsuccessful in the past in part due to strength and permeability problems with the host rock.

The northwest margin of the Bad Vermilion gabbro-anorthosite complex of the Mine Centre greenstone belt is mineralized with discontinuous lenses of magnetite and ilmenite. These have been explored as a source of Fe, Ti and V historically (Shklanka 1968; Rose 1969) and, in recent years, by Numax Resources Inc. (Lichtblau et al. 2008). Other iron occurrences within or near the central Wabigoon Subprovince area include a linear aeromagnetic anomaly of more than 25 km length that occurs in Quetico sediments immediately south of the Lac des Iles belt (Ontario Geological Survey 2002). In the area of this anomaly, thin black layers of magnetite are interbedded with wacke, but have apparently not been evaluated for iron potential. The Bending Lake iron deposit, located at Bending Lake (*see* Figure 2, 10 km north of the Pekagoning Lake area and 10 km west of the Ignace area) represents the largest undeveloped iron deposit near the central Wabigoon Subprovince area. The deposit, represented by magnetite interbedded with wacke and chert is estimated to have proven reserves of 249 million tonnes of ore grading 28.19% Fe occurring mainly within a 330 by 1150 m zone at Bending Lake (Bending Lake Iron Group Limited, www.bendinglakeiron.com [accessed April 14, 2010]).

BASE METALS

Base metals are represented by primarily copper, lead and zinc that are typically found with other metals including nickel, cobalt, molybdenum, gold and silver. Eckstrand, Sinclair and Thorpe (1996) showed that base metals occur in a wide variety of mineral deposit types including those within sedimentary rocks (e.g., Mississippi Valley-type Pb-Zn), metamorphic rocks (e.g., skarn Cu) and felsic intrusive rocks (e.g., porphyry Cu-Mo). Two principal types of base metal occurrences are found within or near the central Wabigoon Subprovince area. These include nickel and copper associated with layered mafic intrusions (*see* “Magmatic Nickel-Copper-Platinum Group Metals”) and copper, lead and zinc within volcanic-associated massive sulphide deposits.

Volcanogenic massive sulphide (VMS) deposits occur as lenses and veins of massive sulphides that formed in volcanic or sedimentary host rocks on the sea floor due to circulation of metal-rich hydrothermal fluids (Galley, Hannington and Jonasson 2007). Within Archean greenstone belts, many VMS deposits are associated with relatively young (2.75 to 2.70 Ga) rhyolitic and andesitic rocks (Fyon et al. 1992). A particular type of rhyolite (FIII rhyolite), characterized by relative flat REE profiles with variable to deep Eu anomalies, low Zr/Y and intermediate to high abundances of high-field strength elements, commonly hosts VMS deposits. Lesher et al. (1986) interpreted that FIII rhyolites are derived from high-level subvolcanic magma chambers that supplied heat to drive the ore-forming hydrothermal system. The movement of metal-rich hydrothermal fluids has commonly altered the host-rock beneath VMS deposits and at greenschist- to amphibolite-facies metamorphism, these alteration zones show distinct coarse-grained assemblages containing phlogopite, cordierite, anthophyllite, muscovite, staurolite, garnet, andalusite and kyanite (Galley, Hannington and Jonasson 2007).

Several base metal occurrences probably representing parts of VMS deposits are found in the central Wabigoon Subprovince area. Poulsen (2000a) described zinc-copper showings including the Gagne Lake, Wind Bay, Lockhart, Farrington and Port Arthur Copper showings in the Mine Centre belt. These are characterized by seams and veinlets of sphalerite, chalcopyrite, galena and pyrite in strongly foliated and interlayered felsic, intermediate and mafic metavolcanic rocks cut by amphibolite sills and dikes. The occurrences are distributed southwesterly more or less along the north side of the Bad Vermilion gabbro complex. The Port Arthur Copper Mine was mined briefly in 1916–1917 and produced 12 050 kg of Cu.

The Mine Centre belt and most other greenstone belts have been explored sporadically for base metals. For example, Wallbridge Ltd. explored gneissic areas of the Hillyer domain (*see* Figure 3) in the late 1990s and defined several sulphide showings characterized by pyrite with minor chalcopyrite. In the course of the present study, numerous rusty zones within greenstone slivers north of Mine Centre and in the Raleigh Lake greenstone belt were sampled. These rusty zones represent pyritic iron formation or disseminated sulphides in metavolcanic rocks and assays gave low values of copper (*see* Stone, Hallé and Chaloux 1998b; *see also* MRD 242: Table 4).

The southern Lumby Lake greenstone belt contains several base metal showings within strongly foliated felsic metavolcanic rocks. These include the Cote zinc showing that represents stringers of up to 10% sphalerite (Jackson 1985b) and the Bernatchez occurrence that consists of disseminated sulphides in strained volcanic breccia. Samples from the Bernatchez showing assayed at greater than 5000 ppm Cu, Pb and Zn and up to 868 ppm Ag (*see* MRD 242: Table 4, samples 95RB01 to 95RB11). Lavigne and Scott (1995) suggested that the Bernatchez showing could represent a remobilized VMS deposit although the high level of strain in the area precludes an accurate assessment.

Perhaps one of the more useful guides to VMS exploration in the central Wabigoon Subprovince area comes from routine analysis of the trace element characteristics of intermediate to felsic metavolcanic rocks. Primitive mantle-normalized trace element profiles are shown in Figure 7 for most major greenstone belts and only the felsic rocks from the Mine Centre greenstone belt show characteristics of FIII rhyolites (*see* Figure 7j). Although the number of samples is too small to accurately define which are the most prospective parts of the Mine Centre belt, the data including the Neoarchean age, presence of FIII rhyolites and numerous base metal showings allude to this belt having good VMS potential.

Averill (2001) identified suites of indicator minerals that are characteristic of certain types of massive sulphide deposits such as VMS deposits. These include the sulphide minerals comprising the deposit and silicate minerals typically found in the alteration zones below VMS deposits (e.g., chalcopyrite, pyrite, goethite, spinel, rutile, kyanite, sillimanite, staurolite, spessartine, fayalite, orthopyroxene and chromite). Samples of surficial material such as till can be screened for these indicator minerals as a means of predicting the potential for sulphide deposits in the bedrock beneath or up-ice from the sample site. During the present study, the soils in several areas near Atikokan were sampled primarily to search for gold anomalies (*see* "Gold" for discussion of the Harold Lake, Crooked Pine Lake (Sapawe) and Bolton Bay areas). A few samples from each area were also processed for indicator minerals of metamorphosed massive sulphide deposits; however, the results were generally negative.

RARE METALS

Rare metals include Li, Rb, Cs, Be, Nb, Ta and Ga and are typically associated with minerals including spodumene, lepidolite, beryl and columbite-tantalite in highly fractionated, commonly dike-like phases of the peraluminous granite suite. Peraluminous granite occurs overwhelmingly in the Quetico Subprovince, which extends through the southern central Wabigoon Subprovince area. The Quetico Subprovince, south of the central Wabigoon Subprovince area, has a few occurrences of rare metals such as in the Nym–Niobe lakes area (south of Sapawe; *see* Map P.2229) where pegmatites are mineralized with small amounts of beryl, cassiterite and columbite-group minerals.

Mineralized peraluminous granite dikes are noted in the Raleigh Lake greenstone belt. For example, Breaks (1993) described a 4 km long zone of mineralized pegmatite dikes cutting mafic to intermediate metavolcanic rocks of the Raleigh belt west of Raleigh Lake (*see* Map P.2229). The mineralized dikes typically have shallow to intermediate dip and are up to several metres in thickness and up to 200 m in length. Most dikes are composed of massive white peraluminous granite pegmatite locally mineralized with spodumene, beryl, manganocolumbite, manganotantalite, bismuthinite and molybdenite (Breaks 1993).

Rare metal mineralization is otherwise uncommon in the central Wabigoon Subprovince area, although peraluminous granite dikes were noted in most greenstone belts during the course of regional mapping. Pegmatitic peraluminous granite dikes commonly occur in marginal areas of greenstone belts adjacent to intrusive batholiths.

DIAMONDS

The potential for diamonds within or near the central Wabigoon Subprovince area was investigated by processing certain till samples for indicator grains of kimberlite—the principal host rock for economic deposits of diamond. Kimberlite typically occurs as small, poorly exposed plugs that are difficult to find by surface mapping and exploration, but can contain a variety of minerals including pyrope garnet, eclogitic garnet, chrome diopside, ilmenite and chromite. These kimberlite indicator minerals (KIMs) are rare in other rocks found at surface and, hence, their abundance in glacially transported till is used as an indication of the potential for kimberlite in the area (Kjarsgaard 2007).

Samples from a sub-area of the central Wabigoon Subprovince (corresponding more-or-less with the area covered by the Sapawe Map P.3350—Revised: *see* Figure 2) as well as a few samples from the Harold Lake and Bolton Bay areas (*see* Figures 20 and 18) were processed for indicator minerals of kimberlite. Indicator minerals of kimberlite include pyrope garnet (GP), eclogitic garnet (GO), chrome diopside (DC), ilmenite (IL), forsterite (FO) and low-chrome diopside (Low-Cr diopside). Various kimberlite indicator minerals are listed in Table 26 for the Sapawe area; complete results are provided separately (see MRD 242: Tables 27 and 28 for the Harold Lake and Bolton Bay areas, respectively). Only one rounded grain of pyrope garnet was found among 32 till samples from the Sapawe area (*see* Table 26, sample 95DST017) and 3 grains of chromite were found in 9 till samples processed for kimberlite indicators from the Harold Lake area. No KIMs were found in the single sample processed from the Bolton Bay area. The results appear to indicate that kimberlite and the associated potential for diamond is low in the central Wabigoon Subprovince area. In comparison, Stone (2001) found KIMs in approximately half the soil samples taken in the northern Superior area, which implies a higher potential for kimberlite there than in the central Wabigoon Subprovince.

Notwithstanding the apparent absence of kimberlite, MetalCORP Limited recently reported recovery of 6 microdiamonds from a 100 kg sample of ultramafic lapilli tuff of the Grassy Portage ultramafic unit of the Mine Centre greenstone belt (MetalCORP Ltd., www.metalcorp.ca, "North Rock property" [accessed April 14, 2010]). The sample site is situated a few kilometres west of the present area, but, as discussed previously (*see* "Mine Centre Greenstone Belt" and "Steep Rock Lake Greenstone Belt"), the Grassy Portage ultramafic unit is texturally and chemically similar to the ashrock in the Steep Rock Lake belt. These ultramafic rocks apparently contain no indicator minerals of kimberlite and their diamond potential is largely unknown based on routine analysis of KIMs in surficial materials.

MAGMATIC NICKEL-COPPER-PLATINUM GROUP METALS

Eckstrand and Hulbert (2007) divided this mineralization into 2 main types dominated by platinum group metals (sulphide-poor ore) and nickel-copper sulphides (sulphide-rich ore). Sulphide-poor ore is thought to develop because it contains only a small amount of S, which is probably derived from the mantle source and into which platinum group metals (PGM) are strongly partitioned along with minor Ni and Cu on cooling. In contrast, sulphide-rich ore has assimilated enough S from country rocks to develop at least a few percent of Ni-Cu sulphides overwhelming the early PGM sulphides. Platinum group metals (PGMs) include Pt and Pd as well as accessory Ru, Rh, Os and Ir.

In the central Wabigoon Subprovince area, known Ni-Cu-PGM occurrences are mainly of the sulphide-poor type occurring in late mafic intrusions of the Lac des Iles, Quetico and sanukitoid suites. Locally, and mainly at localities just outside the central Wabigoon Subprovince area, sulphide-rich mineralization occurs in mafic to ultramafic rocks associated with greenstone belts. North American Palladium Ltd.'s Lac des Iles Mine represents the only recently active mine in the area, which milled more than 39 million tonnes of sulphide-poor ore from the Mine Block intrusion of the Lac des Iles suite between 1993 and 2007. The annual production from this mine in 2006 included 237 388 ounces (6728 kg) Pd, 22 308 ounces (632 kg) Pt, 17 237 ounces (489 kg) Au, 5.156 million pounds (2.3 million kg) Cu and 2.721 million pounds (1.2 million kg) Ni (North American Palladium Ltd., www.napalladium.com, news release, January 17, 2009 [accessed April 14, 2010]).

Various classifications and interpretations have been recently offered for the sulphide-poor and largely PGM-dominated mineralization in the Mine Block intrusion. For example, this type of mineralization is subdivided into the reef-subtype such as occurs in large layered intrusions (e.g., Merensky Reef) and the magmatic breccia or supersolidus intrusion breccia subtype such as is found in smaller stock-like layered intrusions (Eckstrand and Hulbert 2007; Barrie 1996). Other authors write of the contact-subtype of deposit (Peck et al. 2001; Barrie et al. 2002; Hinchey, Hattori and Lavigne 2005), which occurs near the contact of mafic intrusions and is characterized by a low sulphide content, ubiquitous gabbroic pegmatite and the presence of magmatic breccia.

Lavigne and Michaud (2001) showed that PGM mineralization at Lac des Iles is hosted mainly in the matrix of gabbro breccia and in pegmatitic or varitextured gabbro of the Mine Block intrusion. This deposit is classified as magmatic breccia subtype using the division of Eckstrand and Hulbert (2007). Hinchey, Hattori and Lavigne (2005) interpreted that the mineralization was probably introduced to the Mine Block intrusion by late pulses of fertile primitive magma such as is envisaged for mineralized large layered intrusions. In contrast with quiescent magmatism in large layered intrusions, the late magmas were introduced energetically and affected by magmatic fluids leading to complex breccia textures and alteration in the Mine Block intrusion. Hence, the genesis of the mineralization in the Mine Block intrusion appears to have some characteristics in common with those of large layered intrusions.

The Lac des Iles area was extensively explored beginning in the late 1990s with the result that numerous PGM occurrences were revisited or discovered. The area was remapped as part of the present survey and Stone, Fell, Daley et al. (2003) provided assays of many PGM occurrences, most of which contain up to a few g/t Pd with lesser amounts of Pt and Au and up to a few % Cu and Ni. Many PGM occurrences are in small mafic intrusions or at the margins of mafic intrusions and probably represent contact-subtype mineralization. In summarizing exploration, Stone, Lavigne et al. (2003) noted that the extensive mineralized breccias of the Lac des Iles Mine make up a significant part of the Mine Block intrusion, but are only locally developed in other known intrusions of the Lac des Iles suite. Vaillancourt et al. (2003) and Stone, Lavigne et al. (2003) proposed that reef-type mineralization, which is characteristic of many of the world's large PGM deposits in layered intrusions might be found by careful prospecting of layered strata within the Tib and Northern Ultramafic Complex of the Lac des Iles suite. Most other mafic intrusions are too small and lack well-developed layering for this type of mineralization.

The Quetico suite of mafic to ultramafic intrusions is distributed several hundred kilometres west from Lac des Iles more-or-less along the boundary of the Quetico and Wabigoon subprovinces (MacTavish 1999; Pettigrew and Hattori 2006). Although most are small stocks, several including those located 2 km south of Chief Peter Lake are locally mineralized with copper-nickel sulphides and PGMs and are being explored (Canadian Arrow Mines Ltd., www.canadianarrowmines.ca). Three new intrusions of the Quetico type were identified in the Lac des Mille Lacs area during the present survey including a stock located 3 km south of Baril Bay of Lac des Mille Lacs. Boulders probably derived from this ultramafic stock contain a few percent sulphides and have assay values up to 1 g/t combined Pd+Pt+Au (*see* MRD 242: Table 4, samples 06DS68 to 06DS73). These results suggest 1) that the stock south of Baril Bay needs to be evaluated for PGMs, and 2) that additional mafic intrusions worthy of PGM exploration can be found by careful prospecting of the area west of Lac des Iles along the Quetico–Wabigoon subprovinces boundary.

Many PGM occurrences south of the Lac des Iles Mine including the Wakinoo, Stocker, Turtle Hill, Powder Hill, Stinger and Vande (Stone, Lavigne et al. 2003) occur within mafic rocks distributed around the margin of the Shelby Lake batholith of the sanukitoid suite. It is unclear, however, on the basis of regional mapping, if the mafic rocks constitute an early phase of the sanukitoid batholith or remnants of older intrusions of the Lac des Iles suite, which could have been intruded and deformed by the Shelby Lake batholith. Elsewhere, such as in the Entwine stock, PGM mineralization is indeed associated with the more primitive phases of sanukitoid plutons (e.g., Champion Bear Resources Ltd., www.championbear.com, "Eagle Rock project").

The Campbell zone represents a 20 m wide, 1.2 km long mineralized zone within the oval western end of the Entwine stock. Although conformable with the contact, this zone occurs approximately 1 km inside the Entwine stock and is hosted by fairly homogeneous (non layered and non brecciated) diorite and leucogabbro with a mineral assemblage of plagioclase + biotite + hornblende + clinopyroxene + orthopyroxene and accessory magnetite, ilmenite, quartz, potassium feldspar, apatite and titanite. The Campbell zone is locally mineralized with a few percent chalcopyrite, pyrite, pyrrhotite, pentlandite, millerite, PGM tellurides, PGM bismuthides and electrum and assays of up to 1 g/t combined Pd+Pt+Au are common (Stone 2000). This type of PGM mineralization would likely be classified as the sulphide-poor subtype, but is notably richer in copper than nickel. Within the mineralized zone, silicates are altered to actinolite, chlorite, epidote, sericite, albite and calcite.

On the basis of surface mapping, the mineralized part of the Entwine stock does not appear much different from quartz-undersaturated phases of sanukitoid plutons, except for local rusty patches caused by weathered sulphides and greenschist alteration of silicates. Using the Campbell Zone as a model, Stone (2000) suggested that the somewhat complex mineralogy including 2 pyroxenes, 2 Fe-oxides and a suite of greenschist-facies secondary silicates with up to a few percent sulphides can possibly be used as an exploration tool for PGM mineralization in other sanukitoid plutons.

The second major type of magmatic nickel-copper-platinum group metal mineralization includes occurrences dominated by Ni-Cu sulphides with minor PGMs and Au. Two major examples of this type of mineralization occur in mafic to ultramafic rocks adjacent to the central Wabigoon Subprovince area. These include the Shebandowan Mine, which is located 20 km south of Lac des Mille Lacs and milled 8.7 million tonnes of ore at 2.06% Ni, 1.00% Cu and 3.0 g/t Pd+Pt+Au from deformed ultramafic to mafic rocks of the Shebandowan greenstone belt in 1971 to 1998 (Cullen, Brown and Sedore 2007, p.1). The Northrock property is located a few kilometres west of the present area and represents the base of the layered gabbroic Grassy Portage intrusion of the Mine Centre greenstone belt. The North Rock property has an historic reserve of about 1 million tonnes of ore grading 1.2% Cu with minor Ni, PGMs and Au (MetalCORP Ltd., www.metalcorp.ca, "North Rock property" [accessed April 14, 2010]). Poulsen (2000b) attributed the copper-rich nature of this ore to it having segregated from a mafic as opposed to ultramafic source.

The upper parts of large layered mafic intrusions can be mineralized with Fe-Ti oxides (Gross 1996) and here again, 2 examples occur near margins of the central Wabigoon Subprovince area. The upper part of the Grassy Portage intrusion and also the western contact area of the adjacent layered Bad Vermilion intrusion of the Mine Centre greenstone belt are mineralized with magnetite + ilmenite (Poulsen 2000b). These oxides occur as disseminated to massive lenses within dioritic to gabbroic phases of the intrusions. The Seine Bay iron prospects of the Bad Vermilion intrusion have a historic reserve of more than 2 million tonnes of titaniferous magnetite (Poulsen 2000b) and have recently been explored by Numax Resources Ltd. (Lichtblau et al. 2008).

Most major greenstone belts of the central Wabigoon Subprovince area contain gabbro sills and masses although few of these show obvious layering. A possible exception is the eastern Steep Rock Lake greenstone belt where 2 large blocks of layered gabbro were observed at separate localities within otherwise massive gabbro of the Dismal assemblage. Stone (2007) proposed that the blocks may represent fragments of a largely unexposed magma chamber. The massive gabbro either originated from the layered sequence or crosscut the layered sequence thereby incorporating fragments of the layered gabbro.

Nickel-copper mineralization is commonly located in basal parts of mafic intrusions and where feeder-conduits have entered the magma chamber (Eckstrand and Hulbert 2007). Establishing the base and magma conduits are important, but, nonetheless, very difficult steps in exploring any intrusion for nickel-copper mineralization within complexly deformed Archean rocks. Sulphur-rich country rocks such as sediments may also help to make a mafic intrusion prospective for nickel-copper sulphides.

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Metric Conversion Table

Conversion from SI to Imperial			Conversion from Imperial to SI		
SI Unit	*Multiplied by*	*Gives*	*Imperial Unit*	*Multiplied by*	*Gives*
			LENGTH		
1 mm	0.039 37	inches	1 inch	**25.4**	mm
1 cm	0.393 70	inches	1 inch	**2.54**	cm
1 m	3.280 84	feet	1 foot	**0.304 8**	m
1 m	0.049 709	chains	1 chain	20.116 8	m
1 km	0.621 371	miles (statute)	1 mile (statute)	**1.609 344**	km
			AREA		
1 cm^2	0.155 0	square inches	1 square inch	**6.451 6**	cm^2
1 m^2	10.763 9	square feet	1 square foot	**0.092 903 04**	m^2
1 km^2	0.386 10	square miles	1 square mile	2.589 988	km^2
1 ha	2.471 054	acres	1 acre	0.404 685 6	ha
			VOLUME		
1 cm^3	0.061 023	cubic inches	1 cubic inch	**16.387 064**	cm^3
1 m^3	35.314 7	cubic feet	1 cubic foot	0.028 316 85	m^3
1 m^3	1.307 951	cubic yards	1 cubic yard	0.764 554 86	m^3
			CAPACITY		
1 L	1.759 755	pints	1 pint	0.568 261	L
1 L	0.879 877	quarts	1 quart	1.136 522	L
1 L	0.219 969	gallons	1 gallon	**4.546 090**	L
			MASS		
1 g	0.035 273 962	ounces (avdp)	1 ounce (avdp)	28.349 523	g
1 g	0.032 150 747	ounces (troy)	1 ounce (troy)	**31.103 476 8**	g
1 kg	2.204 622 6	pounds (avdp)	1 pound (avdp)	**0.453 592 37**	kg
1 kg	0.001 102 3	tons (short)	1 ton (short)	**907.184 74**	kg
1 t	1.102 311 3	tons (short)	1 ton (short)	**0.907 184 74**	t
1 kg	0.000 984 21	tons (long)	1 ton (long)	**1016.046 908 8**	kg
1 t	0.984 206 5	tons (long)	1 ton (long)	**1.016 046 90**	t
			CONCENTRATION		
1 g/t	0.029 166 6	ounce (troy)/ ton (short)	1 ounce (troy)/ ton (short)	34.285 714 2	g/t
1 g/t	0.583 333 33	pennyweights/ ton (short)	1 pennyweight/ ton (short)	1.714 285 7	g/t

OTHER USEFUL CONVERSION FACTORS

	Multiplied by	
1 ounce (troy) per ton (short)	31.103 477	grams per ton (short)
1 gram per ton (short)	0.032 151	ounces (troy) per ton (short)
1 ounce (troy) per ton (short)	20.0	pennyweights per ton (short)
1 pennyweight per ton (short)	0.05	ounces (troy) per ton (short)

Note: Conversion factors which are in bold type are exact. The conversion factors have been taken from or have been derived from factors given in the Metric Practice Guide for the Canadian Mining and Metallurgical Industries, published by the Mining Association of Canada in co-operation with the Coal Association of Canada.

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Chart A. Accompanies OFR 5422

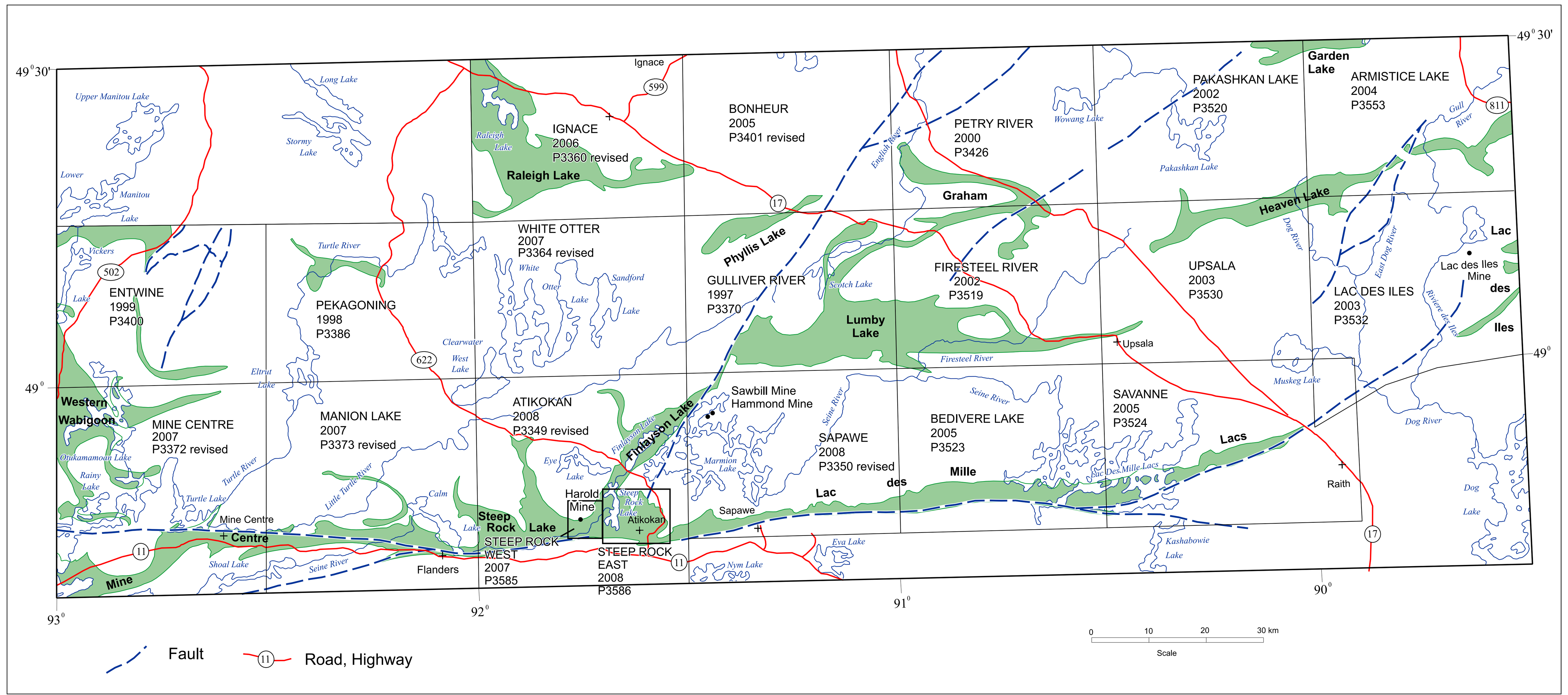


Figure 2. The central Wabigoon Subprovince area showing study area mapped and associated published maps. The map name, year of publication and OGS Preliminary Map number are indicated in each mapped area. The names of area greenstone belts and some past-producing mines are shown. For complete references for the published maps, *see* "References": Stone (2005c, 2005d, 2008a, 2008b, 2008c); Stone, Carter, Hallé et al. (2007); Stone, Carter, Hallé and Pufahl (1997); Stone, Fell and Daley (2004); Stone, Fell, Daley et al. (2003); Stone, Fell and Metsaranta (2003); Stone and Hallé (1999, 2005); Stone, Hallé and Chaloux (1998a); Stone, Hallé, Hellebrandt et al. (2007a, 2007b); Stone, Hallé, Lange et al. (2007); Stone, Hallé, Metsaranta and Petersen (2002); Stone, Hallé and Petersen (2000); Stone and Lennox (2007); Stone and Metsaranta (2002).

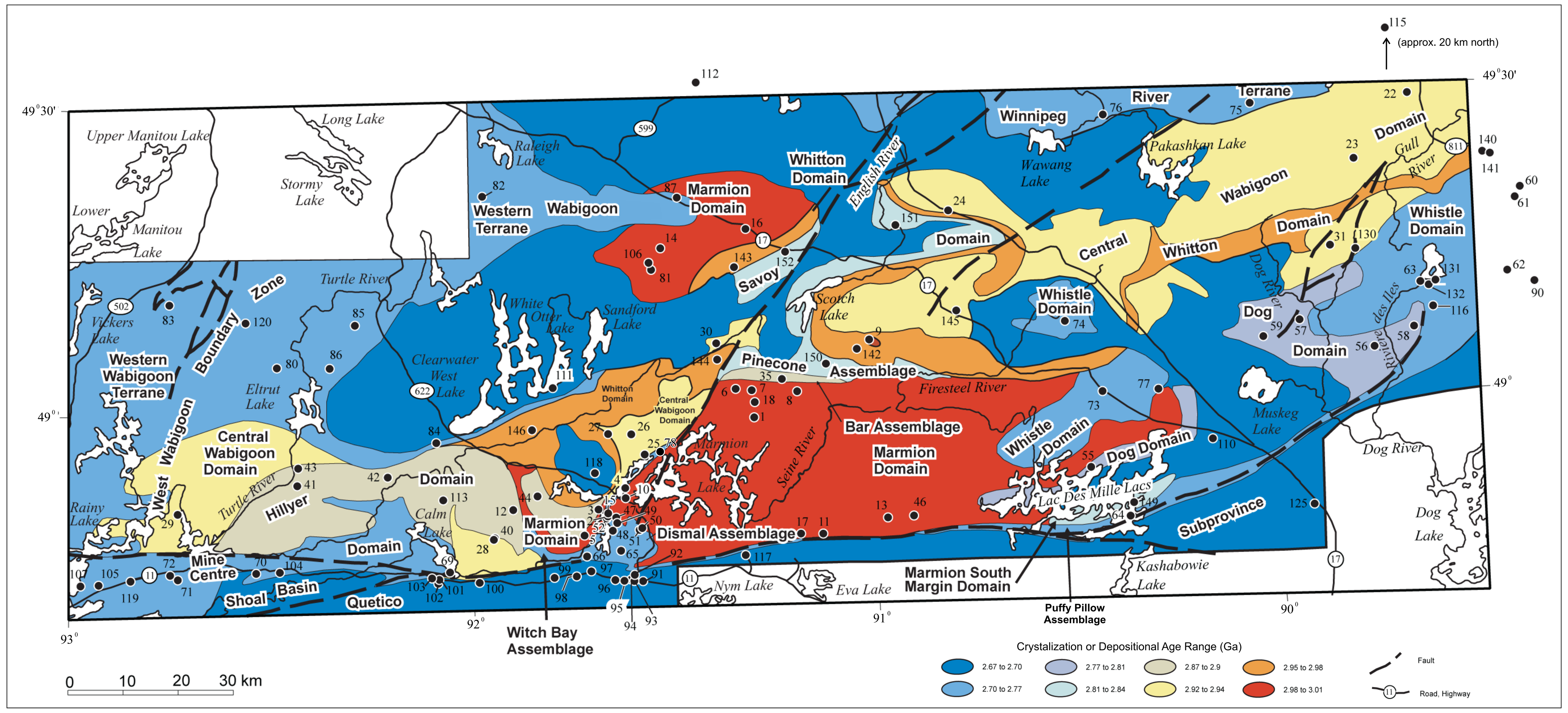


Figure 3. Age domains of the central Wabigoon Subprovince area. Numbers refer to age determination samples listed in Table 2.

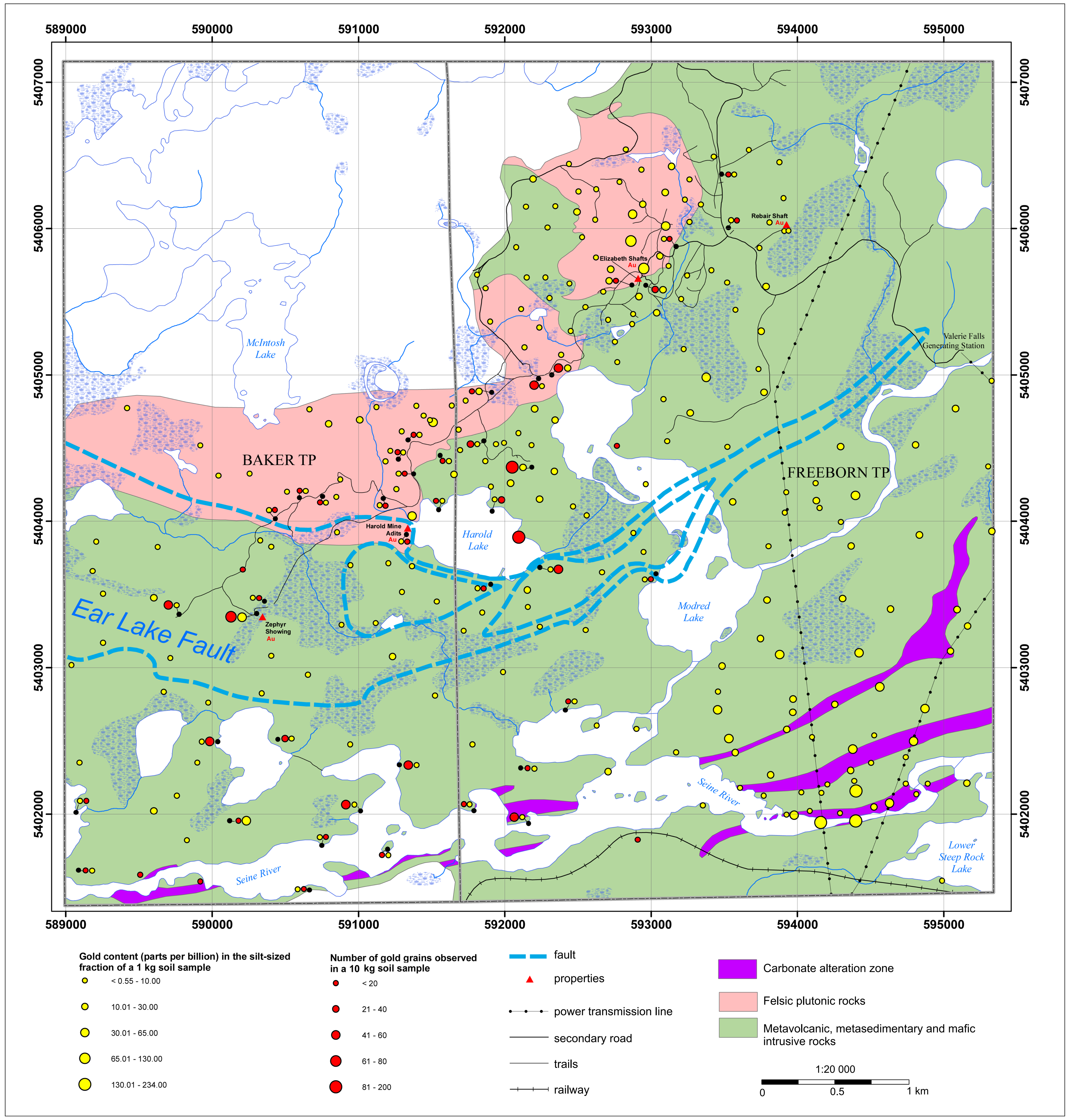


Figure 20. Distribution of gold in soil samples, Harold Lake area, west of Atikokan.

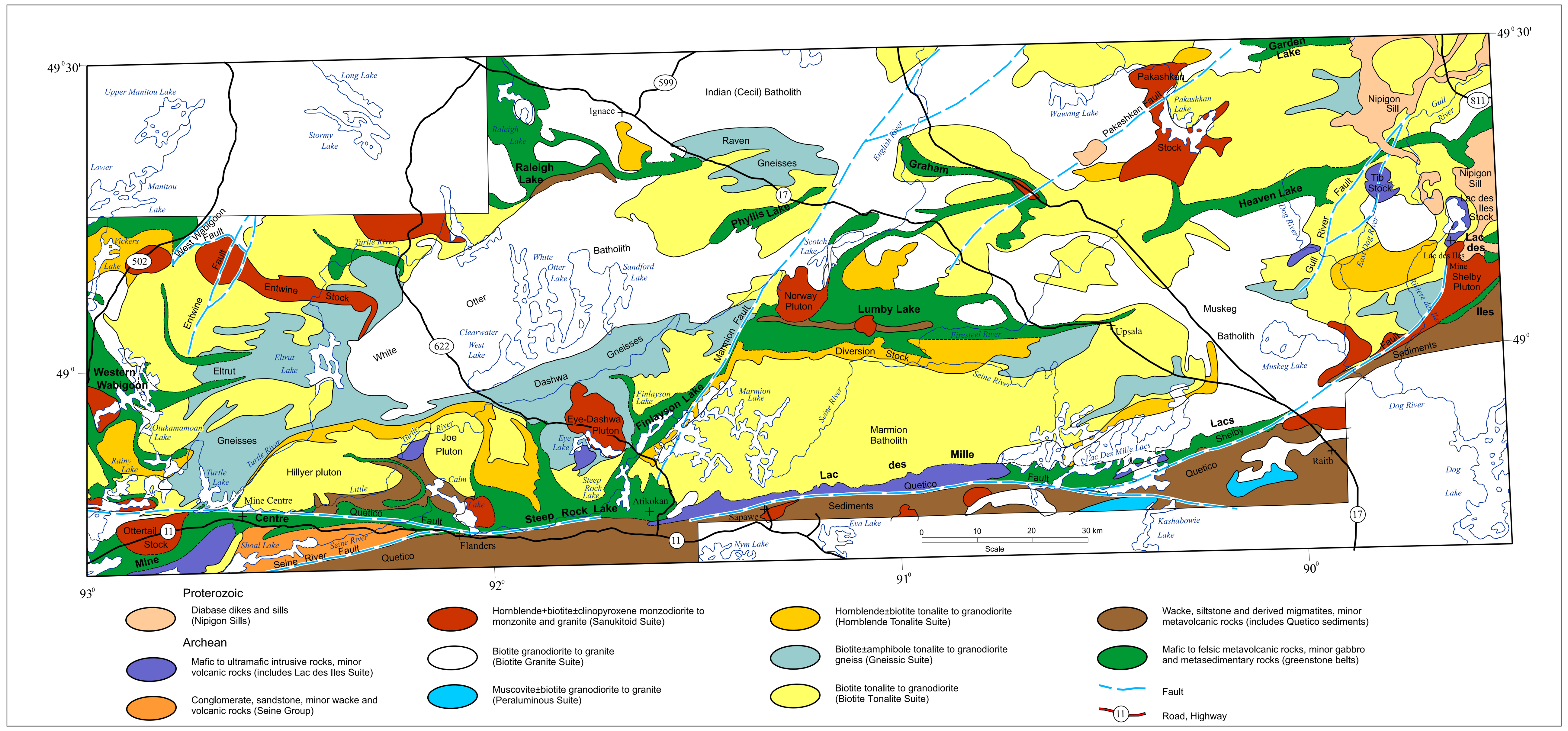


Figure 5. General geology of the central Wabigoon Subprovince area (*see* Stone 2010a (Map P.2229, back pocket) for more detailed geology).

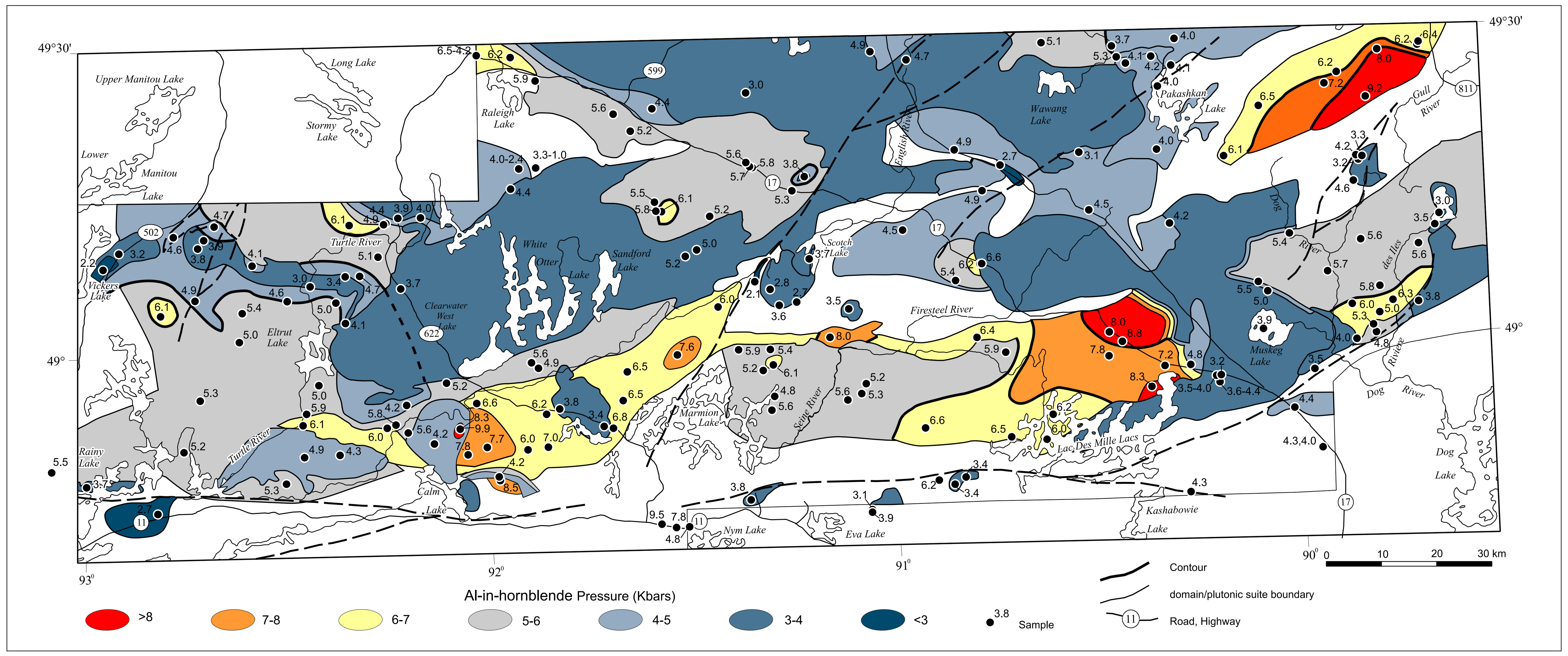


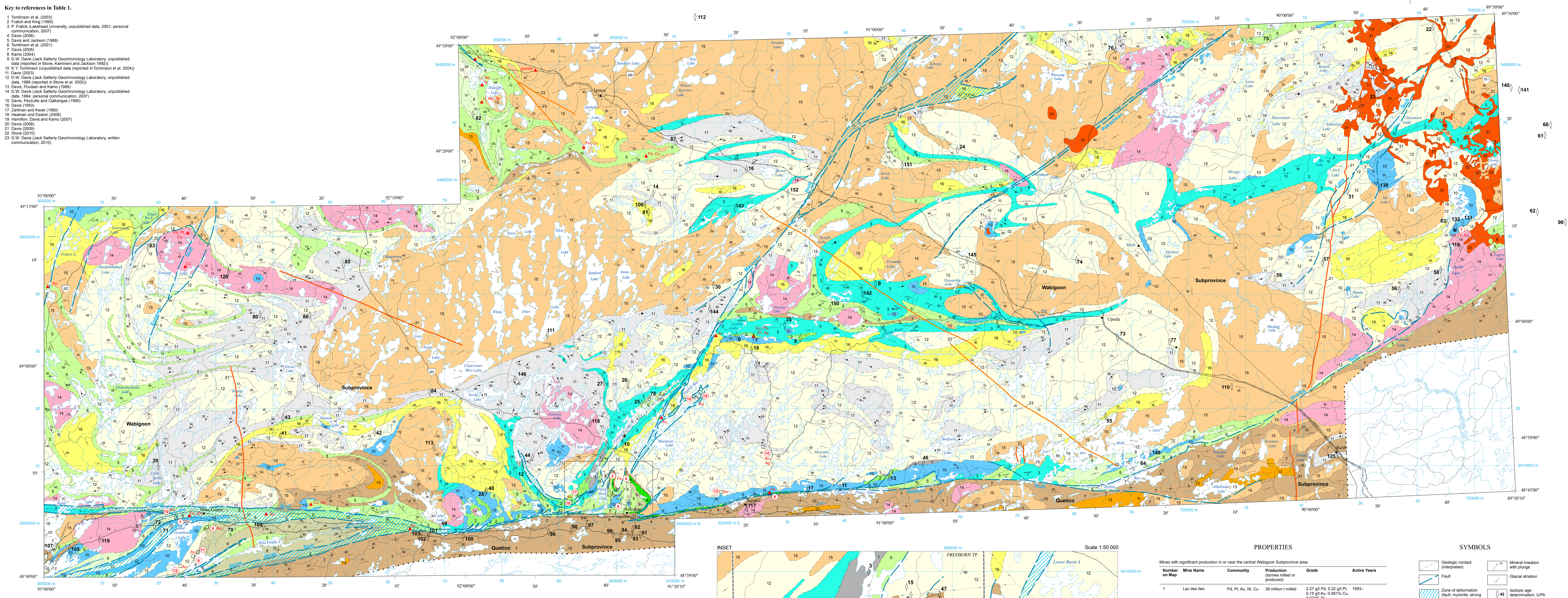
Figure 16. Contoured plot of aluminum-in-hornblende pressure for crustal blocks and late plutonic suites of the central Wabigoon Subprovince area.

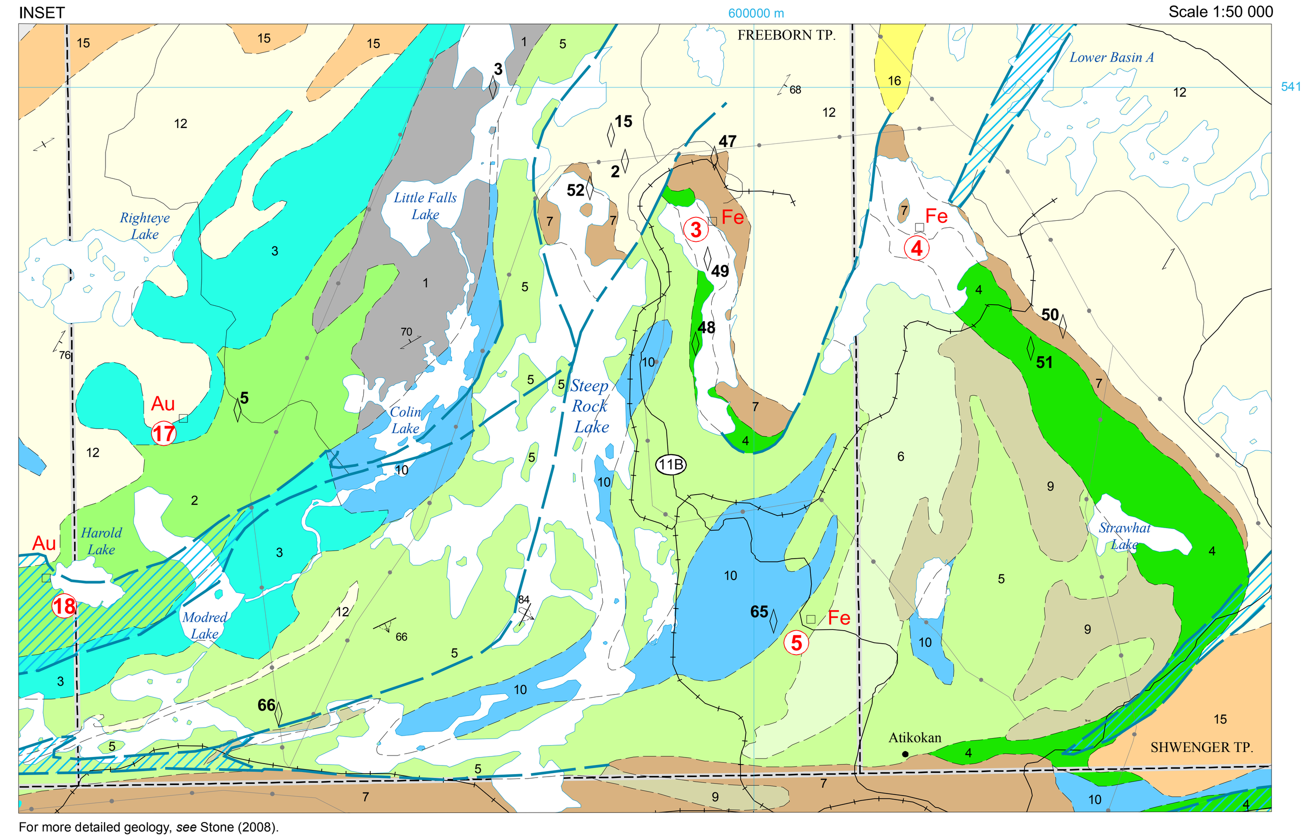
Table 1. U/Pb geochronology data for volcanic, plutonic and sedimentary units of the central Wabigoon Subprovince area. The domains, assemblages, terranes and plutonic suites listed in the table are described in Stone (2010).

Map No.	Rock Type	Assemblage or Plutonic Suite	Age of Crystallization	Detrital Grain Age (Maximum Depositional Age in bold)	Inherited Age	Metamorphic Age	Reference (see key list)
Marmion Domain							
1	tonalite gneiss	Tonalite Gneiss	3009±8				1
2	mafic (hornblende) tonalite	Hornblende Tonalite	3001.6 +3.4/–2.3				1
3[r]	sandstone	Finlayson East		**2997.1±0.9**			3
4	conglomerate, sandstone	Finlayson East		3002, 3001, 3001, 2999, 2999, **2997**			2
5	quartz-feldspar porphyry dacite or subvolcanic intrusion	Finlayson East	2999.4±0.7				4
6	felsic tuff	Lumby South	2999.4±2.9				5
7	rhyolite	Lumby South	3001.1±1.1				1
8	quartz porphyry	Lumby South	2999.1±2.0				5
9	rhyolite	outlier of Marmion Domain	3013.6±1.3				1
10	quartz porphyritic felsic volcanic	Finlayson East	3003.3±0.7				19
11	leucogabbro	mafic tonalite/gabbro	3004.5±1.1				19
12	felsic volcanic	unknown assemblage	2998.5±0.8				20
13	quartz-porphyry felsic volcanic or intrusion	unknown assemblage	3007.6±1.2				20
14	tonalite gneiss	Tonalite Gneiss	2998.1±0.9				20
15	gabbro dike	dike cuts Marmion batholith	3002.1±0.7				20
16[r]	tonalite gneiss	Tonalite Gneiss	2989 +5/–3				6
17	felsic fragmental	unknown assemblage	3006.1±0.7				20
18	hornblende tonalite	Hornblende Tonalite	2998.6±0.7				21
Whitton Domain							
140[t]	felsic dike	Whitton	2953.1±1.6				1
141[t]	felsic tuff	Whitton	2953.7±1.3				1
142	felsic tuff	Lumby North	2963.3±4.5		3016±5		1
143	felsic tuff	Phyllis	2955.8±0.8				1
144	tonalite gneiss	Tonalite Gneiss	2713±4[s]		2950±15[s]	2700±8[s]	19
145	biotite tonalite	Biotite Tonalite	2952±4		2993±11		20
146	tonalite gneiss	Tonalite Gneiss	2957±1				21
Central Wabigoon Domain							
22	biotite tonalite	Biotite Tonalite	2924.1±9			2690 (titanite, lower intercept)	8
23	tonalite gneiss with wacke and gabbro	Tonalite Gneiss	2937.1±1.2*			2722 (monazite)	7
24	biotite tonalite	Biotite Tonalite	2929±4				4
25	felsic tuff	Finlayson West	2931.4±2				1
26	biotite tonalite	Biotite Tonalite	2936				9
27	tonalite gneiss	Tonalite Gneiss	2928				9
28	felsic volcanic fragmental	unknown assemblage	2925.6±1.2				1
29	tonalite gneiss	Tonalite Gneiss	2924±2				6
30	biotite tonalite	Biotite Tonalite	2722±8[s]		2932±5		19
31	biotite tonalite	Biotite Tonalite	2917±12[s]				19
Bar Assemblage							
35	felsic tuff	Bar assemblage	2897.6±2.1		2903 to 2999		1
Hillyer Domain							
40	hornblende tonalite	Hornblende Tonalite	2883±2				10
41	hornblende tonalite	Hornblende Tonalite	2869±4				6
42	biotite tonalite	Biotite Tonalite	2866±8				22 (this study)
43	leuco tonalite gneiss	Tonalite Gneiss	~2880		~3000		21
44	tonalite gneiss	Tonalite Gneiss	2886±7		2875 to 2965		21
Pinecone–Savoy and Marmion South Margin Domain							
149	felsic volcanic fragmental	unknown assemblage	2830.2±1.1				1
150	felsic tuff in komatiite	Pinecone assemblage	2828.3±1.8				1
151	biotite tonalite	Biotite Tonalite	2814.0±1.2				4
152	biotite tonalite	Biotite Tonalite	2817 +4/–2			2694 (titanite)	4
Dog Domain							
55	hornblende tonalite	Hornblende Tonalite	2780.7±1.8				1
56	tonalite gneiss	Tonalite Gneiss	2773.6±1.2				4
57	biotite tonalite to granodiorite	Biotite Tonalite	2777±3				4
58	tonalite gneiss	Tonalite Gneiss	2775±8			2722±3	11
59	tonalite gneiss	Tonalite Gneiss	2729±12[s]		2796±8[s]		19
Steep Rock Assemblage							
46	conglomerate	unknown (possibly Steep Rock assemblage)		3007, 3006, 3004, 3002, 2981, **2927**			7
47	tonalite cobble in conglomerate	Steep Rock assemblage		**2994**			9
48	sandstone	unknown (possibly Seine assemblage)		**3002.6±1.4**			12
49[r]	ultramafic lapilli tuff	Dismal assemblage			2997, 2995, 2780		1
50	sandstone (Wagita Formation)	Steep Rock assemblage		2984±11, **2779±22**			20
51	ultramafic lapilli tuff	Dismal assemblage			2995±11, 2782±15		20
52	wacke (Wagita Formation)	Steep Rock assemblage		2997±4 (84 grains by LA-ICP–MS), **2939±20**			21
Whistle and Mine Centre Domains, Puffy Pillow and Witch Bay Assemblages, Winnipeg River Terrane							
60[t]	felsic tuff	Whistle	2729±2		2813		1
61[t]	quartz+feldspar porphyry intrusion	Whistle	2727.8±1.6		2758 to 2779		1
62[t]	dacite tuff	unknown assemblage	2727.8±1.3				1
63	hornblende tonalite	Hornblende Tonalite	2727.8±1.5				11
64	dacite breccia	Puffy Pillow	2725.0±1.1				4
65	leucogabbro	Witch Bay	2735.2±0.8				4
66	quartz porphyritic tonalite	dike cutting Witch Bay	2729.7±1.9		2865		4
69[r]	felsic volcanic	unknown assemblage	2722.3 +1.3/–1.1				13
70	quartz-phyric volcanic rock	unknown assemblage	2727.0±1.2				13
71	tonalite	Biotite Tonalite	2728.1 +1.8/–1.3				13
72	quartz porphyritic rhyolite	unknown assemblage	2727.8 +4/–2.6				13
73	biotite tonalite	Biotite Tonalite	2716.8±1.0			2707±37 (titanite)	7
74	biotite tonalite	Biotite Tonalite	2712±6[s]				19
75	rhyolite	unknown assemblage	2725.5±1.3				1
76	tonalite gneiss	Tonalite Gneiss	2712±2				10
77	biotite tonalite	Biotite Tonalite	2725±6[s]		2811±24[s]	2721±6[s], 2678±9[s]	19
78	wacke	Witch Bay		61 grains by LA-ICP–MS ranging from **2678** to 3031			23
Western Wabigoon Terrane							
80	tonalite gneiss	Tonalite Gneiss	2732				10
81	hornblende tonalite	Hornblende Tonalite	2721.3±1				6
82	intermediate volcanic breccia	unknown assemblage	2730±7[s]				22 (this study)
83	biotite tonalite	Biotite Tonalite	2716±6[s]				22 (this study)
84	mafic tonalite	Biotite Tonalite	2697.4±0.8				20
85	biotite tonalite	Biotite Tonalite	2709±9			2703±5	20
86	biotite tonalite	Biotite Tonalite	2688±5		2740, 3293±30		20
87	biotite tonalite	Biotite Tonalite	2719		~2923		21
Quetico and Seine Sequences							
90[t]	conglomerate	Seine assemblage		2696			14
91	wacke	Quetico assemblage		3009, 3009, 3004, 3004, 2701, **2698**			15
92	wacke	Quetico assemblage		**2707**			15
93	wacke	Quetico assemblage		**2704**			15
94	wacke	Quetico assemblage		**2739**			15
95	wacke	Quetico assemblage		2901, **2701**			15
96	wacke	Quetico assemblage		**2796**			15
97	wacke	Quetico assemblage		2768, **2706**			15
98	wacke	Quetico assemblage		2719, **2709**			15
99	wacke	Quetico assemblage		**2792**			15
100	wacke	Quetico assemblage		2709, **2704**			15
101	wacke	Quetico assemblage		2913, 2787, **2704**			15
102	wacke	Quetico assemblage		2724, 2705, **2700**			15
103	wacke	Quetico assemblage		**2788**			15
104	tonalite clast in conglomerate	Seine assemblage		**2696.1 +5/–3**			13
105	wacke	Coutchiching/ Quetico assemblage		3059, 3057, 3056, 2940, 2830, 2726, 2721, 2714, 2712, 2704, **2704**			13
106	conglomerate	Seine assemblage		2718, 2710, 2707, **2700**			6
107	felsic sill in Coutchiching sediments	Coutchiching/ Quetico assemblage			2717, 2825		13
Late Plutonic Rocks							
Biotite Granite Suite							
110	potassium feldspar megacrystic granite	Biotite Granite	2686±2		2690		4
111	coarse-grained biotite granite	Biotite Granite	2685±2			2667 (titanite)	16
112[t]	biotite granite	Biotite Granite	2671				10
113	biotite granodiorite	Biotite Granite	2686±6[s]				22 (this study)
Sanukitoid Suite							
115[t]	monzonite	Sanukitoid	2697				10
116	hornblende diorite	Sanukitoid	2690±0.9			2674, 2667 (titanite)	8
117	quartz monzonite	Sanukitoid	2688±4				15
118	biotite-hornblende granite	Sanukitoid	2684±25**			2665 (titanite)	17
119	quartz monzodiorite	Sanukitoid	2686.1 +1.5/–1.4				13
120	monzodiorite	Sanukitoid	2685±8[s]				22 (this study)
Tonalite Gneiss Suite							
125	tonalite to granodiorite gneiss	Tonalite Gneiss	2673.0±1.1			2657 (titanite)	7
Lac des Iles Suite							
130	gabbro pegmatite	Lac des Iles	2685.9±1.6				11
131	gabbro pegmatite	Lac des Iles	2693.3±1.3				18
132	gabbro pegmatite	Lac des Iles	2689.0±1.0				11
133[t]	gabbro	Lac des Iles?	2699.0±0.9				18
134[t]	gabbro	Lac des Iles	2687.6±1.2				18
135[t]	gabbro	Lac des Iles	2686.1±2.2				18
136[t]	gabbro	Lac des Iles	2671±11				18

* the dated zircon grains are rounded and probably are derived from the sedimentary component of the rock
** U/Pb whole rock age
[s] analyzed by sensitive high-resolution ion microprobe (SHRIMP)
[t] located outside the area of this study, but included for comparative purposes
[r] location approximate

Key to references in Table 1.





GEOCHRONOLOGY REFERENCES

PROPERTIES

SYMBOLS

ABBREVIATIONS

LEGEND

Ontario Geological Survey

MAP P.2229

PRECAMBRIAN GEOLOGY

CENTRAL WABIGOON SUBPROVINCE AREA

NORTHWESTERN ONTARIO

SOURCES OF INFORMATION

CREDITS