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1998

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edited by J.A. Ayer, C.L. Baker, J.C. Ireland, R.I. Kelly and P.C. Thurston

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1. Precambrian Geoscience Section

J.A. Fyon
Ontario Geological Survey, Precambrian Geoscience Section

RESPONSIBILITY OF THE PRECAMBRIAN GEOSCIENCE SECTION

In support of the Ontario minerals industry, the Precambrian Geoscience Section (PGS) is responsible for 1) description of the Precambrian bedrock, metallic mineral deposits and their geological histories; 2) interpretation of geologic environments; and 3) inference of unrecognized or new geological environments and their mineral potential. The PGS is also responsible for the geophysical characterization of Precambrian rocks.

PROGRAM DIRECTION — PROVINCIAL STUDIES

Some geological problems require a provincial-scale approach: 1) documentation of craton-scale features to understand controls on the distribution of kimberlite intrusions (Sage, this volume) and petalite- and rare-metal-bearing pegmatites (Breaks, this volume); 2) documentation of metamorphic assemblages to provide insight into the tectonic, thermal and metallogenic histories and hydrothermal alteration assemblages related to mineralization (Easton, this volume); and 3) reprocessing of OGS-sponsored airborne magnetic and electromagnetic data sets (Gupta 1996).

PROGRAM DIRECTION — SUPERIOR PROVINCE

Precambrian Geoscience Section program activities in the Superior Province enhance our understanding of 1) poorly understood areas of perceived high-mineral potential; 2) assemblages that constitute greenstone belts and their metallic mineral deposits (Ayer, this volume); 3) relationships between volcanic, plutonic, thermal and mineralization histories in greenstone belts; and 4) relationships between old and young crust and metallogenic implications (Parker, this volume; Stott, this volume).

PROGRAM DIRECTION — GRENVILLE AND SOUTHERN PROVINCES

Program activities in the Grenville and Southern provinces include studies of 1) petrogenesis and metallogeny of Nipissing intrusions (McRae et al., this volume) and the Sudbury Impact Structure (Spray et al., this volume); and 2) tectonic development, thermal history and metallogenic implications of the Grenville Province (Easton, this volume).

PROGRAM DELIVERY

The PGS geoscience program is delivered through 1) geoscience initiatives (Thurston, this volume); 2) individual, provincial-scale projects; and 3) collaborative projects with private sector, university or other government players. We design PGS projects based on the geoscience problem in an area, PGS technical skills available, amount of bedrock exposure, type and amount of assessment-file information available, and proprietary information available from a client or project collaborator.

Precambrian Geoscience Section acquired and will release unpublished bedrock maps of parts of the Grenville Province from S.B. Lumber's, formerly of the Royal Ontario Museum. The release of these maps fills a void in our understanding of the western Grenville Province and extends the PGS mapping output.

COLLABORATIVE PROJECTS

Precambrian Geoscience Section operates or is a player in 22 collaborative projects involving private sector, university, and other governments (Table 1.1). These collaborative projects 1) complement or extend the Precambrian bedrock geoscience program; 2) broaden the technical geoscience skills available to the PGS; 3) help offset the high cost of delivering new airborne geophysical-survey data. All collaborative project players contribute cash or proprietary data and information not otherwise available to the PGS. All collaborative projects have joint project steering committees to ensure joint planning and co-ordination and to ensure project results meet the needs of the players. The PGS publishes collaborative project data after a period of confidentiality — typically one year.

PROJECT MANAGEMENT AND IMPACT ANALYSIS

The PGS program consists of approximately 50 active projects. Twenty-two are collaborative projects governed by agreements or contracts with external parties. Optimal project delivery to clients requires careful definition and monitoring of project goals, deliverables, milestones, costs, contractual obligations, dependencies between different projects, and project resources. Therefore, project-management practices are an integral part of the PGS program. In addition, we monitor the scientific and economic impacts of PGS projects to ensure our projects
meet their objectives, are results-oriented, and are relevant to mineral industry clients (Churchill, this volume).

INTERJURISDICTIONAL REPRESENTATION

Precambrian Geoscience Section represents the Ontario Geological Survey (OGS) on several inter-jurisdictional committees and associations, including 1) Committee of Provincial Geologists (CPG); 2) National Geological Surveys Committee; 3) Canadian Geoscience Council, for CPG; 4) the federal-provincial Intergovernmental Working Group (IGWG) Task Force on “Alternative Funding Models for Government Geological Surveys” and its derivative task force “How Much Mapping Is Enough”; and 5) Conference Board of Canada Research and Development Impact Network. Client benefits from this representation include 1) shared experiences and approaches dealing with management, operational, technical and political issues; 2) performance benchmark information; and 3) national collaborative activities, such as Team Canada marketing and the national publication meta-data initiative. The national meta-data initiative will post OGS publication meta-data information on an Internet web site maintained by the Geological Survey of Canada.

ONTARIO GEOLOGICAL SURVEY ADVISORY BOARD

In November 1997, the OGS Advisory Board held its inaugural meeting. The Advisory Board advises the Minister of Northern Development and Mines on strategic issues facing the entire OGS, including the geoscience functions of the Resident Geologist Program, Sedimentary Geoscience Section, Precambrian Geoscience Section, Data Services Section and Information Services Section. The Advisory Board struck two committees. A Technical Subcommittee will, in collaboration with the OGS 1) classify Ontario into areas of mapping priority; 2) recommend priorities suitable for development of a five-year plan; 3) participate in an OGS–GSC “geoscience needs analysis” of Ontario; and 4) review annual OGS work plans. A Mines Ontario Subcommittee will review

Table 1.1 Precambrian Geoscience Section Collaborative Projects

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the proposal to deliver geoscience services from an authority arms-length from government.

CLIENT SERVICE

The PGS geoscience program will continue to be technically sound and relevant to Ontario mineral industry because 1) PGS staff geoscientists and support staff are committed to scientific and technical excellence and excellent client service; 2) participation on external committees affords new program opportunities and insights; and 3) clients have the opportunity to comment on the program content and direction through the OGS Advisory Board, its Technical Subcommittee, and at regional client consultation meetings.

REFERENCE

2. Precambrian Initiatives

P.C. Thurston

Ontario Geological Survey, Precambrian Geoscience Section

INTRODUCTION

The initiatives of the Precambrian Geoscience Section represent geographic or functional groupings of projects. These initiatives represent two styles of project execution:

1. integrated team projects in which individual subprojects are part of a tightly integrated whole (e.g., the Hemlo project)

2. project groupings in which several independent projects contribute to a larger scale goal. This represents the NATMAP initiative in which individual OGS and GSC projects contribute to the overall NATMAP goal. Toward the later stages of the NATMAP effort, integrated teams will produce regional-scale compilations of the entire NATMAP area.

The Precambrian Geoscience Section program offered in fiscal 1998–99 is a curtailed version affected by budgetary constraints. Only 2 projects in 1998 had conventional senior assistants contributing to the generation of routine map coverage. However, 4 graduate students are involved in collaborative projects which tie in very directly to the goals of the section initiatives. These collaborative projects include the following:

1. geochronology of rare-metal pegmatite development (Smith, Tindle and Breaks, this volume)

2. a structural study and an alteration study near the Holloway deposit (Luinstra and Benn, this volume; Ropchan, this volume)

3. trace element geochemistry of Abitibi Subprovince metavolcanic rocks (Oliver et al., this volume)

The major initiatives of the Section are 1) Abitibi initiative, 2) NATMAP and LITHOPROBE (concentrated in northwestern Ontario), and 3) mapping in other regions (including regional-scale mineral-deposit studies) and program support. In recent years, we have attempted to split our efforts roughly as follows: Abitibi (30%), NATMAP and LITHOPROBE (30%), mapping in other regions (30%), and program support (10%).

OVERVIEW

By the 1970s the understanding of Ontario’s Precambrian geology had reached the point where it was possible to erect large-scale subdivisions of the Superior Province (subprovinces) and the Grenville Province (Central Metasedimentary Belt, Central Gneiss Belt). By the early 1990s, the OGS improved the understanding of the geology of deformed areas by successfully applying the concept of lithotectonic assemblages to the Precambrian rocks (Williams, Stott and Thurston 1992). As described below, we have now recognized some of the problems with the larger-scale subprovince framework erected in Geology of Ontario. Indeed, it is likely that the newly proposed subdivisions of the Wabigoon Subprovince (Stott, Davis and Parker 1998; Tomlinson et al. 1998) are but the start of subdivisions midway between the assemblage and the subprovince scale (e.g., Ayer et al., this volume). This scale of subdivision likely represents identification of superterranes making up the 1970s-style subprovinces. Many of these superterranes represent fragments of 2.9 to 3 billion-year-old oceanic volcanic rocks that, in southern parts of the Superior Province, are surrounded by 2.7 billion-year-old greenstone belts. Do these fragments represent microcontinental fragments? Are they independent entities or were they all rifted from, for example, the North Caribou Terrane?

These newly proposed subdivisions of the Abitibi Subprovince (Ayer et al., this volume), and the eastern and central Wabigoon Subprovince (Stott, Davis and Parker 1998; Tomlinson et al. 1998) are key to more precisely assessing mineral potential, particularly lode gold mineralization and rare-metal pegmatites. There are hints (Berger, this volume) that we may have to pay more attention to the total structural picture to assess gold potential: “cross faults” not directly related to subprovince-bounding shear zones may also be the locus of lode gold deposits.

ABITIBI INITIATIVE

The Abitibi initiative involves 6 projects headed by a 1:100 000 scale geological compilation of the Abitibi greenstone belt to update and test the Jackson and Fyon (1991) assemblage model. The remaining projects listed below were designed to fill significant gaps in our knowledge throughout the belt.

1. The digital version of the Abitibi greenstone belt map is a “smart” map with data tables electronically linked to the map polygons. Production time, benefits of this style of map and client opinions of this style of map will determine whether future products may emulate the “smart” maps. This project has used computer-processed aeromagnetic, satellite and RADARSAT data (Madon et al., this volume) as interpretive aids. This project, through use of extensive geochronology, has developed the concept of “super-assemblages” which represent amalgamation of the existing plethora of assemblages (Jackson and Fyon 1991). This has changed interpretation of the Abitibi greenstone belt from allochthonous exotic, i.e., disconnected independent assemblages (Jackson, Fyon and Corfu
1994), to an autochthonous model similar to that of Heather, Shore and van Bremeen (1995) for the Swayze belt. This concept makes Abitibi subprovince assemblages more comparable in scale to those seen in northwestern Ontario.

2. Clay-belt-style mapping began with 1:20 000 scale mapping north and east of Timmins compiling exploration drilling supplemented where possible with outcrop data. This year’s project centered on Guibord, Michaud and Garrison townships along the Porcupine—Destor deformation zone (Berger, this volume). As the project progressed eastward, the density of exploration data decreased and therefore the scale has been changed to 1:50 000. Significant observations include the association of gold mineralization with late structures orthogonal to the Porcupine—Destor deformation zone and an association of gold with mafic phases of Timiskaming-related syenites. The former is similar to recent progress in the Yilgarn craton (Vearncombe 1998).

3. Associated with the above project are 2 MSc projects related to mineralization at the Holloway Mine. These are structural investigations by Luinstra and Benn (this volume) and a geochemical study of the alteration system related to the deposit (Ropchan, this volume).

4. Regional-scale mapping is being carried out at 1:50 000 scale by G.W. Johns in the Shining Tree area south-southwest of Timmins. This project has newly defined the presence of 2 relatively young metasedimentary units both unconformably above Keewatin units. These are the Indian Lake Group, equivalent to the Porcupine Group at Timmins, and an equivalent of the Timiskaming Group. If the autochthonous assemblage model (1 above) holds in this region, then explanation must be sought for the stratigraphically high position of the komatiites in this area. Geochemical study of the volcanic rocks in the area is being co-ordinated as a PhD study at Portsmouth University (Oliver et al., this volume).

5. The Sheraton Township project was designed to gather outcrop and diamond-drill data in order to provide a setting, and thus exploration guides, for the mineralization discovered by Golden Knight Resources Ltd. and Cross Lake Minerals Ltd. This project has made significant modifications at a local scale to the position of assemblages and has provided an assemblage-scale setting for the mineralization (Vaillancourt, this volume). As well, the mapping has disclosed the need for revisions to the large-scale mapping in that area.

6. Through a collaborative project with the Geological Survey of Canada Industrial Partnership Program, the OGS, the GSC and 5 industrial collaborators funded 1:50 000 scale mapping by C.T. Barrie to the west and north of Timmins. As the project includes a time-limited confidentiality clause to benefit the industrial sponsors of the work, details are provided in this volume on early phases: the Fox—Stinson area and geochronology. The OGS will publish all maps resulting from this project at the close of the confidentiality period

7. To complement Precambrian Geoscience Section (PGS) programs, the OGS has entered into a number of collaborative agreements (Fyon, this volume). One in particular, an airborne geophysical survey of the Shining Tree area, specifically in parts of Knight, Tyrrell and Leonard townships, will supplement the work in the Abitibi greenstone belt with data that the PGS would have not be privy to otherwise. Products from this survey include very low frequency (VLF), magnetic (total field and measured vertical gradient) and gamma ray (K, U, Th) data. Due to private industry participation, there is a one-year period of confidentiality; after that time the OGS will release the data to the public (late 1999).

**NATMAP**

NATMAP is a Geological Survey of Canada program designed to emphasize regional mapping in collaboration with provincial surveys and academia. It provides top-up funding solely for the GSC, providing them with the means to develop collaborative multi-agency projects which have produced real innovation in the geology of shield regions. The Western Superior NATMAP Project (Percival et al., this volume) has as its objective the creation of an improved large-scale tectonic framework for the Superior Province west of Lake Nipigon. The need for an improved tectonic framework is exemplified by the fact that in *Geology of Ontario* (Blackburn et al. 1991) the Wabigoon Subprovince is portrayed as a single entity. Volcanic rocks and granitoids range in age from 3 Ga to 2.7 Ga, which suggests more appropriate tectonic subdivisions will be possible. Similarly, there is some scantly evidence that the age of volcanism and shear-zone development north of the North Caribou Terrane may young northward (Thurston, Osmani and Stone 1991). Validation of these concepts will improve large-scale understanding and exploration effectiveness. The following paragraphs attempt to put into perspective the efforts of both the Precambrian Geoscience Section initiative and GSC projects:

1. In the area north of the Uchi Subprovince, D. Stone has been conducting a multiyear mapping project at a scale of 1:50 000 in a transect from the north edge of the North Caribou Terrane to the northern limit of exposure of the Superior Province. He has worked closely with the Manitoba Geological Services Branch in the area of Stull Lake (Stone, this volume). Sampling of granitic rocks will permit a petrogenetic synthesis and geochronological reconnaissance with J. Whalen of the GSC. The results thus far have refined knowledge of 1) the limits and character of the North Caribou Terrane, 2) the differing character of supracrustal units north of the North Caribou Terrane including the possible presence of Timiskaming-style rocks at Stull Lake and the Ellard Lake greenstone belt, and 3) the shear-zone control of the gold mineralization at the Sachigo River Mine.
2. In the Uchi Subprovince, the Precambrian Geoscience Section and the GSC are concentrating efforts in the vicinity of the Uchi–Confederation greenstone belt east of Red Lake. The OGS component is twofold: 1) identification and placement into a regional context of the late, Timiskaming-style metasedimentary occurrences, and 2) facies analysis of pyroclastic and epiclastic rocks associated with the 2.7 billion-year-old, Confederation assemblage, South Bay volcanogenic massive sulphide (VMS) deposit. Devaney (this volume) has provided an explanation of the late deformation patterns in the belt by invoking a northward movement of a microcontinental block. The OGS component of the project is running in parallel with structural analysis of the large-scale relationships between 2.9, 2.8 and 2.7 billion-year-old assemblages by the GSC (van Staal 1998). The former view was either arc-arc collision (Williams, Stott and Thurston 1992) or possibly that the 2.8 billion-year-old Woman assemblage represented some sort of Andean margin on the 2.9 billion-year-old Balmer assemblage followed by a collision between the Woman assemblage and the Confederation assemblage at 2.7 Ga (Stott and Corfu 1991). Van Staal (1998) have made much progress highlighted by the observation that the Balmer–Woman assemblage interface was stitched by circa 2.8 billion-year-old dikes.

3. The central Wabigoon region exemplifies the synergies NATMAP has produced between the OGS, the GSC and academic investigators. The GSC is working in the north-central part of the central Wabigoon region (Percival 1998) concentrating on structures in granitoid rocks and the possible relations between the 2.7 billion-year-old Sturgeon Lake greenstone belt and the Obonga Lake greenstone belt. Sanborn-Barrie, Skulski and Whelan (1998) and Skulski, Sanborn-Barrie and Stern (1998) are examining the structural, geochemical and isotopic framework of the Sturgeon–Savant Lake greenstone belt. Tomlinson (Laurentian University) and co-workers, mainly with LITHOPROBE funding, have been using basalt-komatiite geochemistry, isotopes and U-Pb ages to clarify the relationships amongst greenstone belts throughout the central Wabigoon. D. Stone of the OGS has been mapping at a scale of 1:50 000 in the southern part of the central Wabigoon, concentrating upon relationships between central and western Wabigoon and regional controls on mineralization (Stone et al., this volume). Important results are 1) thrust stacking of sheet-like granitoid bodies in the central Wabigoon region at the latitude of Sturgeon and Obonga lakes (Percival et al., 1998), 2) isotopic and geochemical data indicate that the Obonga Lake greenstone belt represents a 2.8 billion-year-old continental margin succeeded to the south by 2.7 billion-year-old arc volcanism (Tomlinson et al., in press), and 3) the central Wabigoon region can be divided into northern and southern Mesoarchean terranes separated byNeoarchean greenstones (Stott, Davis and Parker 1998; Tomlinson et al. 1998), and 4) the architecture of granitoid complexes in the south-central Wabigoon region suggests that young (circa 2.7 Ga?) batholiths overlie circa 3 billion-year-old greenstones (Stone et al., this volume) suggesting that magmatic intraplating was involved in crustal growth in the region.

4. The OGS project in the eastern Wabigoon region is a north-south transect from the ~ 2.7 billion-year-old accretory margin with the Quetico Subprovince on the south, through to the easterly extension of greenstones (over 3 Ga) found in the Armstrong area. Stott, Davis and Parker (1998) have outlined a North Wabigoon block with 3.05 to 2.9 billion-year-old volcanism forming the northern part of the belt overlain by the continental margin 2.74 billion-year-old Marshall assemblage. To the south across the east-trending Humboldt Bay shear zone is the Elmhirst–Rickaby block containing circa 2.7 billion-year-old volcanic rocks. This year’s results have provided additional details on the relationship between the 2.74 billion-year-old Marshall assemblage, various metasedimentary packages and the older assemblages. The elucidation of relationships is critical to the success of a projected reflection seismic transect of this belt during phase 5 of LITHOPROBE.

5. The major metallogenic component of the NATMAP initiative is an OGS project (Parker, this volume) to compare gold and VMS metallogeny of 2.9 billion-year-old greenstones versus 2.7 billion-year-old greenstones, directly related to the NATMAP theme of understanding relationships between older “oceanic” greenstone blocks versus younger “arc” blocks. In this, the first of 3 years, the project emphasized VMS mineralization in 2.9 Ga sequences in and to the north of Red Lake 2.7 billion-year-old sequences mainly in the Uchi Subprovince. Contrasts were documented in terms of geodynamic setting and metal associations. The older VMS-style occurrences are richer in Zn, Pb and Ag, thus they may represent lower-temperature feldspar-destructive reactions in the hydrothermal system versus reactions which supply Cu through destruction of mafic minerals. As such, they are atypical of modern oceanic mineralizing systems. Subsequent phases of the project will examine the evidence for gold mineralization predating 2.74 Ga in the northern Superior Province.

6. This year the PGS began to examine regional controls on rare-metal mineralization building upon our success in discovering the petalite deposit at Separation Rapids (Breaks and Tindle 1997). The project will examine rare-metal occurrences in the region of the Bearhead fault and other occurrences in northwestern Ontario (Smith, Tindle and Breaks, this volume). Results from this year (Breaks et al., this volume) center on the discovery of a zoned, fractionated pegmatite with pollucite and tantalum-bearing oxide minerals along the Bearhead fault within the North Caribou Terrane. The 2697 million-year-old S-type granitoid bodies generating this mineralization are conventionally taken to represent over-thickened
crust (Romer and Smeds 1997) developed during orogenic processes. Implied in these results is that the conventional view that the Berens River–Sachigo Subprovince interface (the Bearhead fault), or in other schemes the North Caribou Terrane, represents a 2.87 billion-year-old orogeny (Williams, Stott and Thurston 1992), is not wholly correct. There was obviously activity at 2.7 Ga in this older terrane. As well, this activity may explain the mineralization at Wawang Lake (Dyer and Breaks 1995), where rare metals in lake sediment geochemical samples are found at the junction of 2.7 and 3 billion-year-old terranes within the central Wabigoon region.

7. In the Sioux Lookout area, the PGS had a project in 1995 and 1996. Preliminary results are described in the article by Devaney (this volume). He provides an admittedly speculative, but comprehensive review of the orogenic architecture for the belt with revisions to stratigraphy and some suggestions for regional-scale controls on gold mineralization.

OTHER INITIATIVES

Approximately 30% of the program is devoted to initiatives other than NATMAP and the Abitibi program. Projects under this initiative include the following: 1) a comprehensive review of kimberlite magmatism throughout Ontario, 2) a study of regional controls on rare-metal mineralization reported on under the NATMAP initiative, 3) continuing mapping and related work in the Grenville province, and 4) a fundamental re-appraisal of the regional-scale controls on the Hemlo deposit.

1. The kimberlite project has 2 phases: 1) the Lake Timiskaming region, and 2) the Attawapiskat region of the James Bay Lowlands. Any additional significant advance kimberlite geology is passed on to clients when opportune. Phase 1 has been released as an Open File Report (Sage 1998) and phase 2 is expected as an Open File Report in 1999. This volume contains a brief summary on regional controls on kimberlites developed over the last few months (Sage, this volume). There are indications in the report on the Sandor occurrence that additional rock types may well be important in assessing the diamond potential of Ontario. The possibility of kimberlite potential in the Geraldton–Longlac area is an innovative suggestion with much evidence in the form of indicator minerals and geological relationships to recommend it.

2. Work in the Grenville Province this year is represented by a brief assessment of the Puzzle Lake area initiated by concerns about mineral potential. The results indicate potential for zinc, molybdenum and wollastonite mineralization. Results of monazite dating in the Geoscience Laboratory are also reported (Easton and Crabtree, this volume).

3. The Hemlo project represents a team approach to re-appraisal of the regional-scale controls of mineralization. The project represents the OGS contribution to a more detailed examination of the immediate mine area by a number of academic researchers under the aegis of the Canadian Mining Industry Research Organization (CAMIRO). The OGS issued an interim progress report as an Open File Report this spring (Jackson, Beakhouse and Davis 1998). That report provides a broad structural model demonstrating an association of the mineralization with late plutonism and shear-zone activity. The Hemlo team (Jackson, Muir and Beakhouse) are in the process of writing a final report on this project. The controversy of syngenetic versus epigenetic origin for barite-rich mineralization in the Hemlo gold deposit is addressed in the article by Muir (this volume).

4. Substantial progress has been made on the metamorphic map of Ontario. We expect this map and publications based on it to offer a substantial re-interpretation of the tectonic framework of all of Ontario (Easton, this volume). The project has already shown, in several poster presentations, the direct benefits to the exploration industry.

SECTION SUPPORT

The Section has several projects underway designed to facilitate the mandate of the Section. These include a number of digital projects for standardizing and improving digital data flow and the geophysical program.

1. The section geophysicist is currently secondment with the Data Services Section completing the Airborne Magnetic and Electromagnetic Survey compilation and levelling. Therefore there is no geophysical component of the program being delivered this year. The intent is to have the geophysical program operate in support of the geological program by using a geophysical perspective to solve particular geological problems such as the position of a particular structure. With a view to strengthening that aspect of the program, 2 staff have received training in the use of aeromagnetic data.

2. Digital mapping standards are required to produce uniform formats for map manuscripts and to facilitate information exchange between software and hardware systems. We released a set of more than 1600 structural symbols in 1995 and are proceeding (Berduco, this volume) with compilation and release of line standards and standards for mineral deposit and alteration symbols. In an effort to standardize procedures in use of Fieldlog, AutoCAD and Arcview, a Guide to Digital Mapping will be made available in 1999. This will facilitate enormously the process of project startup and ensure a uniform look and feel to OGS maps.

3. In the recent past, we have evaluated a number of geophysical visualization software packages for use by staff of the PGS. Individual project staff are in the main using Montaj for geophysical visualization and PCI for visualization of satellite images, etc.

4. We are beginning to use Arcview to organize and visualize data in a number of projects.
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3. The Abitibi Greenstone Belt: a Program Overview

J.A. Ayer

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INTRODUCTION

The Abitibi greenstone belt (AGB) is the world’s largest Archean greenstone belt, covering an area of approximately 100,000 km². The Ontario portion constitutes about 30% of the belt and is a highly productive part of the province with respect to metal production. In 1995, the annual value of production had reached almost $1 billion including about 550,000 ounces of gold, 125,000 t of zinc and 90,000 t of copper. However, high rates of mining in the active deposits, and a general reduction in exploration expenditures because of declining commodity prices, have resulted in a significant reduction in reserves.

The Precambrian Geoscience Section (PGS) of the Ontario Geological Survey is currently focusing 35% of its operational staff on the AGB portion of the province. As outlined below, our program is designed to facilitate exploration for new deposits by providing current geological mapping projects covering the areas of high economic potential at a variety of scales. We are also actively involved in research on the metallogeny and tectonic evolution of the belt and on applied research on specific mineral-deposit-related problems. The applied research projects are being jointly funded by the PGS in conjunction with a variety of mining companies, the Geological Survey of Canada (GSC) and the earth science departments of a number of universities.

NEW DEVELOPMENTS IN THE ABITIBI GREENSTONE BELT

Geological models for the AGB are changing rapidly. The PGS has profoundly modified the understanding of Precambrian greenstone belts by the advancement of the assemblage concept (Thurston and Chivers 1990). Jackson and Fyon (1991) applied the concept to the Abitibi Subprovince in Ontario and identified 71 assemblages. However, as stated by Jackson and Fyon (1991), some assemblages represented a compromise brought about by an imperfect understanding of the extent of units and their character. The assemblage concept is thus intended to be a “straw dog” against which new mapping and tectonic interpretations can be tested. The merit of the concept is that the reclassification permits portrayal of belt-wide lithotectonic units in which contact relationships can be conformable, unconformable, or allochthonous. Jackson, Fyon and Corfu (1994) interpreted the AGB assemblages as Archean analogues to a variety of modern geodynamic settings representing largely allochthonous terranes juxtaposed into their current positions by plate-tectonic processes along active continental margins.

A recent mapping and geochronological sampling program focussed on the Swayze belt in western part of the AGB led to questions about the merit of the allochthonous-assemblage tectonic model (Heather, Shore and van Bremen 1995, 1996). This work revealed that when unfolded, a largely coherent stratigraphic succession existed on the western side of the Kenogami south batholith with similar ages and rock types to those in the central part of the AGB. Evidence against allochthonous assembly of lithotectonic units in the AGB was also indicated by a minor, but widespread, presence of older inherited zircons within the overlying younger volcanic units in both the central (Corfu 1993) and western regions (Heather, Shore and van Bremen 1995, 1996). The above observations lend support to the more traditional AGB stratigraphic model, in which the lithostratigraphic units were deposited in autochthonous successions with their current complex map-pattern distribution developed through the interplay of multiphase folding and faulting (Heather, Shore and van Bremen 1995).

Tectonostratigraphic revision is ongoing as Ayer et al. (this volume), utilizing published and new geochronological data, and the results of regional-scale compilation, view the supracrustal rocks as an autochthonous and largely coherent stratigraphic succession preserved across broad areas of the AGB. Nine distinctive stratigraphic packages have been identified on the basis of geochronological, lithological and temporal criteria, and are correlated using a modified assemblage nomenclature that combines many of the previously identified assemblages into a more regionally coherent framework. Understanding of the distribution of stratigraphic units has economic as well as scientific importance, because the significant mineral deposits are restricted to specific stratigraphic entities or chronological events in the more than 70 million years of AGB supracrustal history (Ayer et al. 1997, this volume).

Geochemical and petrological research has shown that copper-zinc volcanogenic massive sulphide (VMS) deposits are commonly associated with chemically distinctive felsic volcanic rocks (Lesher et al. 1986; Barrie, Ludden and Green 1993; Parker and Ayer 1997) and that these “fertile” rhyolites are restricted to specific stratigraphic intervals in the AGB (Ayer et al. 1997). Barrie, Ludden and Green (1993) and Xie, Kerrich and Fan (1993) indicated that there is also a variety of komatiite types in the AGB based on differing trace-element concentrations, perhaps reflecting different depths of melting in the mantle source regions. Further research is required to relate the various petrographic types of komatiites to their potential to host magmatic nickel deposits, but empirical evidence suggests that only specific komatiite-bearing assemblages
host economic nickel-copper deposits (Ayer et al. 1997). Thus the above types of applied research point to the importance of considering the mineral potential of the various assemblages separately.

**ABITIBI PROJECTS**

The compilation project has produced a hard-copy and digital version of the Timmins-area map and will continue to produce maps covering the AGB in Ontario at a scale of 1:100 000 (Ayer et al., this volume). An additional phase of the project, to be undertaken in conjunction with the Ministère de l’Énergie et des Ressources du Québec, will provide a set of unified compilation maps covering the whole of the AGB at 1:250 000 scale. The compilation maps, in conjunction with new geochronological and geochemical sampling, enable further review of the assemblages, establishment of a more formalized stratigraphy, and the erection of new metallogenic and geodynamic models for the evolution of the AGB. The project also provides the opportunity to test the application of a number of different remote-sensing themes to compilation mapping in the AGB (see Madon et al., this volume). Data collection for the compilation project is being supplemented by more-detailed mapping projects at a variety of scales (discussed below). These projects are designed to focus on areas where we don’t have an adequate understanding of the relationship of mineralization to geology.

The “clay-belt mapping” project conducted by B. Berger utilizes sparse outcrop information in conjunction with diamond-drill data and airborne geophysical data to produce new geological interpretations in the parts of the AGB which are largely covered by thick Quaternary cover. Over the past 7 years, B. Berger has moved progressively eastward from north of Timmins to the current area centred on the Porcupine–Destor deformation zone from Matheson to the Quebec border. The current project is in the second of a three-year mandate to provide 1:50 000 scale coverage of 12 townships in an economically important area where several gold mines have recently been opened and a number of gold and base metal prospects are in various stages of evaluation. The objective of the project is to compile and integrate a wide variety of existing geological maps of different vintages and scales with high-resolution airborne geophysical data and a wealth of diamond-drill data from numerous exploration projects. To date, the work has revealed a number of important lithological, structural and stratigraphic factors which appear to be critical in focussing gold mineralization (Berger and Amelín, this volume).

Under the auspices of B. Berger’s project, a number of jointly funded thesis projects are being conducted on problems related to the controls on gold mineralization in the vicinity of the Holloway and Holt–McDermott mines. One of the MSc theses has the objective of evaluating the timing and structural controls on the gold ore at the Holloway Mine (Luijnstra and Benn, this volume). A second project involves an MSc and a BSc thesis to determine the lithogeochemical signatures of the various types of alteration associated with gold mineralization in the vicinity of the two mines (Ropchan et al., this volume).

In light of the recent base metal discoveries southeast of Timmins, the PGS has initiated a new mapping project in Sheraton Township (Vailancourt, this volume). The mineralization occurs in felsic to intermediate volcanic rocks with an U-Pb zircon age of 2704 ± 2 Ma. Putting the mineralization into a regional context, VMS deposits of this age occur at Kamiskotia, west of Timmins, and in the Val d’Or area of Quebec, all of which are interpreted to be within the newly defined Tisdale assemblage (Ayer et al., this volume). As the township has extensive overburden cover, the project utilizes “clay-belt mapping” techniques by integrating sparse outcrop data with geophysical and diamond-drill data. The objective of the project is to provide a new 1:20 000 map and accompanying report and to better understand the controls on base metal mineralization in order to help focus exploration in this poorly understood part of the AGB.

The Shining Tree area has received a significantly increased level of exploration for gold and base metals following the opening of the former Temagami land cautions area in 1996. Removal of the caution opened the eastern part of the belt, which had been closed to staking for decades. The increased exploration interest and the need to integrate this poorly understood area with the remainder of the AGB has led to an ongoing project (Johns and Amelín, this volume) which will ultimately provide a new 1:50 000 scale map and report covering 12 townships. The project integrates the results of the ongoing synoptic mapping with previous mapping, geochronological and geochemical data and exploration data. Progress to date has led to subdivision of the eastern part of the area into various Keewatin units, unconformably overlain by Timiskaming-assemblage sedimentary and volcanic rocks of shoshonitic and calc-alkaline affinity. Preliminary results suggest gold mineralization in this area is closely associated with deformation zones and high-level shoshonitic intrusions.

The PGS and Portsmouth University have initiated a PhD thesis project to document the geochemical patterns of the various Archean rock packages in the Shining Tree area (Oliver et al., this volume). This data will be used in conjunction with the mapping of G. Johns to help evaluate the evolution and geodynamic framework of this part of the AGB.

The PGS has also entered into a collaborative-funded project with a number of private-sector partners to fly a new airborne geophysical survey in the eastern part of the Shining Tree area. This new data will be used by G. Johns to aid in the mapping and will be released to the public after a one-year period of confidentiality. Products include very low frequency (VLF), magnetic (total field and measured vertical gradient) and gamma ray (K, U, Th) data. This project constitutes the first attempt by the PGS to outline rock units and alteration zones using radiometric data.

The Kidd–Munro Extension project by T. Barrie is a multiyear mapping and lithogeochemical program jointly funded by the PGS, the GSC and a number of mining companies (Barrie, this volume; Barrie and Corfu, this...
The main objectives of the program are to provide an improved understanding of the relationship of VMS and magmatic nickel-copper deposits to komatites and to augment our understanding of the relationship of the "fertile" Kidd–Munro assemblage to other mafic and ultramafic-bearing assemblages in the AGB. The program is in its third year with the current emphasis on the easternmost parts of Kidd–Munro and Stoughton–Roquemaure assemblages in Ontario.

CONCLUSIONS

In summary, many exciting new advances in our understanding of the geology of the AGB are being made. The following reports in this volume demonstrate the merits of a diverse, but balanced, approach to Abitibi geology using the traditional strengths of PGS mapping, the comprehensive understanding of the ore deposits by the mining companies and the research skills of university earth science departments. This balanced approach embraces projects at the wide variety of scales and methodologies that are required for the range of problems that exist. These projects range from belt-wide compilation, to regional synoptic mapping, to township-scale and down to extremely detailed mine-scale mapping. The program is also focussed on the economically important parts of the AGB in Ontario. We feel our integrated approach to applied research through collaboration with corporate clients, the GSC, and the earth science departments of a number of universities is needed to advance our understanding of the geology and its relationship to mineral deposits in this geologically complex, but highly prospective, part of the province. We also feel confident that the maps, reports and theses resulting from this comprehensive program will provide many new ideas and concepts that are, and that will continue to be, directly applicable to exploration for new mineral deposits.

REFERENCES


INTRODUCTION

This multi-year project will produce a series of geological compilation maps covering the Abitibi greenstone belt over the next few years (Ayer and Trowell 1996). Products to date include a 1:100 000 scale map covering the western half of Figure 4.1 in both hardcopy (Ayer and Trowell 1998) and digital (Ayer et al. 1998) versions. Hardcopy and digital maps covering the eastern half of Figure 4.1 will be released in 1999.

As a part of this project, a number of samples from the compilation area were submitted for U-Pb analysis of zircons by F. Corfu and Y. Amelin at the Royal Ontario Museum. Results of the samples received to date are displayed on Figure 4.2. This component of the project is designed to complement geochemistry by other workers (see e.g., Corfu 1993, Bleeker and Parrish 1996, Heather, Shore and van Breeman 1996, Bleeker, Parrish and Sager-Kinsman in press, Barrie and Corfu this volume). The results will help to resolve fundamental questions about the tectono-stratigraphic evolution of the largest Archean greenstone belt on Earth. Understanding of the distribution of stratigraphic units has economic as well as scientific importance, because the economically significant deposits are restricted to specific stratigraphic entities or chronological events in the more than 70 million years of Abitibi greenstone belt supracrustal history (Ayer et al. 1997).

A number of different tectonic models have been proposed for development of the Abitibi greenstone belt. Up until the 1980s interpretation utilizing standard stratigraphic nomenclature was primarily based on superposition as determined by field relationships. The most complete Abitibi greenstone belt-wide subdivision into supergroups, groups and formations was presented in a 1:500 000 scale map (MERQ-OGS 1983). This work was based on the results of mapping at a variety of scales and limited U-Pb zircon ages. Until the mid 1980s, the tectonic models for Abitibi greenstone belt stratigraphy were dominantly based on vertical tectonics (or fixist) models. These models hypothesised development of a belt-wide geosynclinal basin formed in response to riftiing, crustal subsidence and cyclical volcanic and sedimentary infill (e.g., Goodwin 1977, Jensen and Langford 1985).

An increased focus on structural geology, and trace element geochemical data led to the application of plate tectonic (or mobilist) models to the Abitibi greenstone belt in the late 1980s. Jackson and Fyon (1991) subdivided the Abitibi greenstone belt into over 50 different lithotectonic assemblages. This reclassification permitted portrayal of belt-wide lithotectonic units in which contact relationships could be conformable, unconformable, or allochthonous. Jackson, Fyon and Corfu (1994) interpreted these assemblages as Archean analogues to a variety of modern geodynamic settings representing largely allochthonous terranes juxtaposed into their current positions by plate tectonic processes along active continental margins.

A recent mapping and geochronological sampling program focussed on the Swayze belt in western part of the Abitibi greenstone belt led to questions about the merit of the allochthonous assemblage tectonic model (Heather, Shore and van Breeman 1995, 1996). This work revealed that when unfolded, a largely coherent stratigraphic succession existed on the western side of the Kenogamius batholith with similar ages and rock types to those in the central part of the Abitibi greenstone belt. Evidence against allochthonous assembly of lithotectonic units in the Abitibi greenstone belt was also indicated by a minor but widespread presence of older inherited zircons within the overlying younger volcanic units in both the central (Corfu 1993) and western regions (Heather et al. 1996). The above observations lend support to the more traditional Abitibi greenstone belt stratigraphic model, in which the lithostratigraphic units were deposited in autochthonous successions with their current complex map pattern distribution developed through the interplay of multiphase folding and faulting (Heather 1998).

REVISED STRATIGRAPHY

We herein present new geochronological results (see also Barrie and Corfu, Berger and Amelin, Johns and Amelin, this volume), which in conjunction with published data (e.g., Corfu 1993, Bleeker and Parrish 1996, Heather, Shore and van Breeman 1996, Bleeker, Parrish and Sager-Kinsman in press), permits stratigraphic revision. The data supports a predominantly autochthonous stratigraphic succession in the Abitibi greenstone belt, which can be extended into the poorly exposed areas north of the
Figure 4.1. Revised stratigraphic subdivisions for the northern part of the Abitibi greenstone belt in Ontario with geochronology sample locations as discussed in the text.
Figure 4.2. U/Pb isochron plots for samples collected in the northern part of the Abitibi greenstone belt in Ontario (see Figure 4.1 for locations).
Figure 4.2. continued.
Porcupine Destor Fault (Figure 4.1). Based on the lithological and geochemical results and the geochronological data, there appears to have been a period of almost continuous volcanism from about 2745 to 2698 Ma in which 7 distinct volcanic assemblages were deposited, followed by 2 distinctive sedimentary assemblages at about 2695 Ma and 2675 Ma respectively. The term assemblage as used herein supercedes the terminology of Jackson and Fyon (1991), but is still based on the assemblage definition of Thurston (1991) in which assemblages were defined as stratified volcanic and/or sedimentary rock units deposited during a discrete interval of time in a common depositional setting. The rock units typically share common lithofacies and may share additional attributes such as geochemistry, geophysical signatures, and structural style. The term assemblage as used in this report is not intended to have any specific geodynamic implications, but is used as a convenient term to correlate lithostratigraphic units (which are not necessarily contiguous) with common ages, rock types and chemical signatures. A future paper will provide a more formalized stratigraphy upon completion of the compilation project, with a more complete breakdown of the assemblages, into groups, formations and members.

**Pacaud Assemblage**

The Pacaud assemblage has an age range from of about 2745 to 2740 Ma and consists of calc-alkaline and tholeiitic volcanic rocks commonly restricted to the margins of large plutonic complexes such as the Pacaud tuffs (Jens and Langford 1985) in the Kirkland Lake area which have an age of 2747±2 Ma (Corfu 1993) and the 2740 Ma Chester Group on the southwestern margin of the Kenogamissi batholith and south of the Rideout Deformation Zone in the Swayze greenstone belt (Heather 1998). Inherited zircons of this age in younger volcanic sequences have also been identified in the Swayze greenstone belt (Heather 1998) and the southern portion of the Shining Tree greenstone belt (Johns and Amelin, this volume). Volcanic rocks of the Pacaud assemblage have not as yet been detected within the map area (Figure 4.1), but detrital zircons of this age are found in the Porcupine assemblage (see below).

**Deloro Assemblage**

The Deloro assemblage has an age of about 2730 to 2725 Ma and consists of mafic to felsic calc-alkaline volcanic rocks commonly capped by regionally extensive chemical sedimentary units. These units are also typically found adjacent to large plutonic complexes such as the Hunter Mine Group east of the Abitibi batholith (Corfu 1993), and the Marion Group, west of the Kenogamissi batholith (Heather 1998), the Deloro Group east of the Kenogamissi batholith and in antiformal dome structures such as the one centred on Shaw Township (Corfu 1993). That portion of the Deloro assemblage defined as the Deloro Group by Pike (1982) occurs south of the Porcupine Destor Fault in the central part of the Shaw Dome (Figure 4.1). It has been dated at 2725±1 Ma (Corfu 1993) and is now further confirmed by 2725±2 Ma zircons in felsic tuff in Whitney Township (Ayer et al. 1997) and 2724±4 Ma in Langmuir Township (Barrie and Corfu this volume).

Felsic volcanic rocks in Lamarche and Simson townships also have zircon ages of about 2725 Ma (Barrie and Corfu this volume). They occur in a poorly exposed band of calc-alkaline volcanic rocks with iron formation that extends across the northern part of the map area (Figure 4.1).

Calc-alkaline rhyodacite tuff from diamond drill core in southwestern Duff Township (sample # 96-JAA-37) contains small euhedral zircon grains of variable colour. The 207Pb/206Pb ages of three analyzed grains are well within analytical errors of each other, and their weighted mean of 2723±1.2 Ma is thought to yield the crystallization age (Figure 4.2a). This age implies the package of dominantly calc-alkaline volcanic rocks with abundant intercalated iron formations extending westerly from the Buskegau River Fault to the Mattagami River Fault (Ayer and Trowell 1998) belongs to our newly defined Deloro assemblage (see Figure 4.1).

Three analyses were carried out on selections of subhedral zircon tips, euhedral short prismatic crystals and one fraction of unabraded euhedral prisms and tips from a calc-alkaline felsic volcanic sample from drill core in southeastern Thorburn Township (sample # 96JAA-23). The two abraded fractions are concordant and the unabraded fraction constrains a line with 84% probability of fit passing through a lower intercept age of 170 Ma (Figure 4.2b). The upper intercept age of 2719.5±1.5 Ma dates the extrusion of this unit. Based on this age and its occurrence within a thick package of mafic to felsic calc-alkaline volcanic rocks and iron formation extending west of the Mattagami River Fault, the unit is considered to be part of the Deloro assemblage (Figure 4.1). However, further geochronological sampling will be needed to clarify stratigraphic relationships in the poorly exposed supracrustal rocks within the southwestern part of the map area.

**Stoughton–Roquemaure Assemblage**

The Stoughton–Roquemaure assemblage consist of thick magnesium- and iron-rich tholeiitic basalt sequences with localized komatiites and felsic volcanic units. It occurs as an extensive unit in the northeastern part of the map area (see Figure 4.1) and in the Kirkland Lake area where it was identified as the Wabewawa/Catherine groups (Corfu 1993). The Stoughton–Roquemaure Group as defined by Jens and Langford (1985) was lumped together with the Kidd–Munro assemblage by Corfu (1993). However, a confomal contact with underlying calc-alkaline rocks of the Hunter Mine Group in Quebec (Dostal and Mueller 1997) dated at 2730±2 Ma and an age of 2720±2 Ma for the Wabewawa/Catherine Group (Corfu 1993), suggests the assemblage was deposited between the Deloro and Kidd–Munro assemblages (Figure 4.3).

Inherited zircons with an age of 2720±1 Ma have been found in felsic volcanic rocks in the Kidd–Munro
Assemblage in Nova Township (see below) indicating that the Stoughton–Roquemaure assemblage may be also have been present in the western part of the map area.

**Kidd–Munro Assemblage**

The newly defined Kidd–Munro assemblage rocks have an age range from about 2717 to 2712 Ma and are subdividable into 2 distinct suites: 1) a tholeiitic to komatiitic portion which consists of komatiites, magnesium- and iron-rich tholeiites; and 2) a calc-alkaline portion consisting of intermediate to felsic pyroclastic rocks. It is an economically important unit that is host to a number of base metal deposits including the giant Kidd Creek VMS deposit (Barrie and Corfu this volume). In its traditionally defined distribution from Kidd to Munro townships (Jackson and Fyon 1991), it consists of tholeiitic mafic volcanic rocks and komatiites with localized intercalations of high silica FIIIb type rhyolites (Lesher et al. 1986). Comprehensive geochronological investigations have ef-

![Diagram of stratigraphic columns for the north, central, and southern parts of the Abitibi greenstone belt in Ontario.](image)

**Figure 4.3.** Idealized stratigraphic columns for the north, central, and southern parts of the Abitibi greenstone belt in Ontario.
effectively constrained the age of this part of the assemblage to the interval 2717 to 2712 Ma (Corfu 1993, Bleecker and Parrish, 1997, Bleecker, Parrish and Sager-Kinsman in press, Barrie and Corfu, this volume). A sample of quartz-phryic, FIII rhyolite, tuff breccia (sample # 96JAA-88) from diamond drill core in central Reid Township yields zircons with a variety of discordance patterns which imply two lead loss events with lower intercept ages of about 600 and 1400 Ma. Using these different lead loss lines and the average of the 2 least discordant abraded grains (i.e., 0.5% and 0.6%), indicates upper intercepts with a minimum age of 2714 Ma and a maximum of 2719 Ma (Figure 4.2c). These new data indicate the Kidd–Munro Assemblage extends west to the Mattagami River Fault (see Figure 4.1).

Recent regional-scale geochronological studies now indicate the assemblage is more extensive than was originally indicated by Jackson and Fyon (1991). Heather, Shore and van Breeman (1996) expanded the distribution of calc-alkaline volcanic rocks of this age into Keith and Foleyet Townships in the northern Swayze greenstone belt, and the Biscotasing arm portion of the southern Swayze greenstone belt.

Clear colourless and visually uniform zircons were collected from a sample of calc-alkaline felsic tuff (sample #96JAA-94) interbedded with oxide-facies iron formation within a thick sequence of tholeiitic mafic volcanic in Carsscallan Township. Two out of three analyses are concordant, and the third one is 0.7% discordant. All three analyses yielded consistent 207Pb/206Pb ages with the weighted mean of 2713.3±0.9 Ma (Figure 4.2d) representing its crystallization age. Four pale brown elongated zircons were analysed from a calc-alkaline, quartz-phryic felsic volcanic rock (sample #96JAA-41) in northern Nova Township. All four grains are concordant, but their 207Pb/206Pb ages vary: two grains are ca. 2720 Ma, and the other two are 2715 Ma. The difference between the two groups is outside the 2σ error limits. The weighted mean of the two younger grains yield an age of 2715.1±1.9 Ma, which is interpreted as the crystallization age (see Figure 4.2e). The two slightly older grains are interpreted to be inherited xenocrysts with a Stoughton–Roquemaure age of 2720 Ma. The ages of the above two samples indicate that the Kidd-Munro assemblage can now be extended west to the Kapuskasing Structural Zone (see Figure 4.1).

Geochronological results from the eastern part of the study area (Corfu 1993, Berger and Amelin this volume) indicate a predominantly calc-alkaline component of the Kidd–Munro assemblage, previously identified as the Hunter Mine group by Jensen and Langford (1985), extends eastward to the Quebec border.

**Tisdale Assemblage**

The newly defined Tisdale assemblage rocks have an age range from about 2710 to 2702 Ma. Similar to the Kidd–Munro, the Tisdale assemblage is subdivisible into 2 distinct suites: 1) a tholeiitic to komatiitic portion which predominantly consists of komatiites, magnesite- and iron-rich tholeites and FIII rhyolite; and 2) a calc-alkaline portion predominantly consisting of intermediate to felsic pyroclastic rocks. Previously published U-Pb determinations from the komatiitic to tholeiitic portion of the Tisdale assemblage include 2707±3 Ma for gabbro associated with an ultramafic dike intruding the Deloro Group in the Shaw Dome area, and 2707±2 Ma for gabbro and 2705±2 Ma for FIII rhyolite at Kamiskotia respectively (Corfu 1993).

A new age has been determined from long prismatic and colourless zircons that were extracted from a massive, high iron tholeiite mafic flow unit (sample #96JAA-91) in southern Tisdale Township (see Figure 4.1). The sample was collected from an extensive stratigraphic unit known as the “99 Flow” of the Vipond Formation at the base of Pyke’s (1982) Middle Tisdale Group. Two fractions were recovered from clear angular zircon fragments, after abrasion they yield slightly discordant but overlapping analyses (Figure 4.2f). Together with one unabraded fraction they define a line with 98% probability of fit, a lower intercept of 732 Ma and an upper intercept age of 2707±3 Ma which dates crystallization of the unit.

Analyses were carried out on a short prismatic zircon and an anhedral zircon from a calc-alkaline felsic lapilli tuff (sample #96JAA-86) from diamond drill core in central Belford Township. When unabraded, the grains have nearly concordant analyses which when constrained by an unabraded fraction, yield an upper intercept age of 2710±2 Ma (Figure 4.2g). The thin calc-alkaline unit, is intercalated with abundant tholeiitic mafic flows and komatiites which suggests that the tholeiitic to komatiitic portion of the Tisdale assemblage extends westward from Kamiskotia into the northern part of the Montcalm greenstone belt (see Figure 4.1).

Two newly dated FIII rhyolite units intercalated with iron-rich tholeiitic flows also yield Tisdale ages. In southern Mahaffy Township a sample of sphaleritic rhyolite flow (sample #96JAA-87) taken from diamond drill core yielded abraded zircons which define two overlapping analyses near Concordia (Figure 4.2h). The unabraded fraction is discordant and defines a line with a lower intercept of 145 Ma. The upper intercept age of 2703±2 Ma dates the age of crystallization. A rhyolite tuff (sample #96-JAA-93) was collected from drill core in southwestern Jessop Township. Analyses of four abraded euhedral zircons are concordant to very nearly concordant and yield an age of 2702±1 Ma (Figure 4.2i).

South of the Porcupine Destiny Fault, volcanic rocks of the unit previously identified as the Bowman Assemblage by Jackson and Fyon (1991) are now assigned to the newly defined tholeiitic to komatiitic portion of the Tisdale assemblage (Figure 4.1). The Marker Horizon (Corfu 1993) consists of intermediate to felsic calc-alkaline volcanic rocks that overlies the Bowman to the south. Felsic tuff breccia (sample #97JAA-101) from diamond drill core associated with the newly discovered Cross Lake Minerals Ltd. base metal mineralization in Sheraton Township (Vaillancourt this volume) was analysed for zircon age. Three analyzed abraded zircon grains are concordant to slightly discordant. The weighted mean of the samples yields a crystallization age of 2703.7±1.4 Ma.
(Figure 4.2)). This age is correlative with 2706 ± 2 Ma from felsic volcanic rocks in Corfu’s Marker Horizon, along strike to the east in Currie Township, and 2703 ± 2 Ma from felsic volcanic rocks in a unit identified as the Upper Tisdale to the southwest in Douglas Township (Corfu 1993). These features indicate that the calc-alkaline portion of the Tisdale occurs in the upper stratigraphy of the Assemblage south of the Porcupine Detour Fault.

Calc-alkaline felsic volcanic rocks from northeastern Little Township yields ages of 2706 ± 2 Ma (Ayer et al. 1997) and 2709 ± 3 Ma (Barrie and Corfu this volume). This indicates that the calc-alkaline volcanic rocks previously identified as the Duff–Cousen–Rand Assemblage (Jackson and Fyon 1991) belongs to the newly defined Tisdale assemblage east of the Buskegade River Fault, while the calc-alkaline sequences with abundant intercalated iron formations west of the Buskegade River Fault (Ayer and Trowell 1998), are now assigned to the Deloro assemblage (see above) (see Figure 4.1).

Kinojevis Assemblage

The Kinojevis assemblage is restricted to a thick sequence of tholeiitic mafic volcanic rocks in the southeastern part of the map area (see Figure 4.1) previously identified as the Kinojevis Group by Jensen and Langford (1985). Thin felsic flows near the upper portion of the assemblage have been dated at 2701 ± 1 Ma (Corfu 1993, Berger and Amelin, this volume). The limited time interval between the upper calc-alkaline part of the Tisdale south of the Porcupine Detour Fault (see above), and top of the Kinojevis implies the two assemblages are in conformable contact and that the thick succession of Kinojevis mafic volcanic rocks (greater than 10 km) was accumulated in the relatively short time interval of about 2 to 3 million years.

Blake River Assemblage

The newly defined Blake River assemblage consists of calc-alkaline mafic to felsic volcanic rocks in the Ontario portion of the assemblage, but also includes tholeiitic basalt and FII rhyolite in Quebec (Jackson and Fyon 1991). The part of the assemblage classically defined as the Blake River Group (Jensen and Langford 1985) occurs in the southeastern part of the map area (see Figure 4.1). Rhyolite samples from a number of different localities within the Blake River Group in Ontario and Quebec range in age from 2701 to 2698 Ma (Corfu 1993). These ages indicate that the Blake River was also deposited in a relatively short time interval of 2 to 3 million years and implies that it conformably overlies the upper part of the Kinojevis assemblage.

Calc-alkaline units also herein ascribed to the newly identified Blake River assemblage occur as less extensive sequences throughout the Abitibi greenstone belt. Calc-alkaline felsic volcanic rocks of the Krist formation in Tisdale Township have an age of 2698 ± 4 Ma (Corfu 1993). Felsic to intermediate volcanic rocks from the Brett Lake formation in the central part of the Swayze greenstone belt have an age of 2697 Ma ± 1 Ma (Heather 1998).

Calc-alkaline felsic tuff (sample 96JIAA-11) from diamond drill core in southwestern Lennox Township contains a uniform population of small colourless clear euhedral zircon grains. Three analyzed abraded grains of similar appearance yielded variable degree of discordance from 0.1% to 2.6%. The weighted mean yields an age of 2696.6 ± 1.5 Ma (Figure 4.2k). This suggests that poorly exposed portions of the Blake River assemblage also may occur in the northern part of Figure 4.1.

Porcupine Assemblage

The newly defined Porcupine assemblage predominantly consists of wacke, siltstone and mudstone displaying Bouma sequence subdivisions indicating predominantly distal deposition by turbidity currents (Born 1995). Detrital zircons from a number of localities southeast of the Kidd Creek Mine indicate deposition after 2696 Ma (Bleeker and Parrish 1996, Bleeker, Parrish and Sager-Kinsman in press). Polymictic conglomerate and sandstone are more abundant in the portion of the Porcupine assemblage lying to the west of the Mattagami River Fault (Figure 4.1), implying more proximal depositional environment and easier transport. This interpretation is also supported by a unit of conglomerate and sandstone in the northern Swayze greenstone belt with detrital zircons indicating deposition after 2696 Ma (Heather et al. 1996). A maximum age of deposition for Porcupine assemblage sediments is also indicated by the crystallization age of 2698 Ma from conformably underlying felsic volcanic rocks of the Krist Formation in Tisdale Township (Corfu 1993). The Porcupine assemblage unconformably overlies the Kidd–Munro assemblage in Kidd Township (Bleeker and Parrish 1996, Bleeker, Parrish and Sager-Kinsman in press) and the Tisdale assemblage in Jessop Township (see above). These contact relations imply that the Porcupine assemblage basal contact is conformable with the Blake River assemblage in some localities, but unconformably overlying older assemblages in other parts of the Abitibi greenstone belt.

The turbidite-bearing sequence in the northeastern part of Figure 4.1 was previously identified as the Scapa Assemblage (Jackson and Fyon 1991). The unit is laterally extensive continuing eastward for at least another 150 km into Quebec, where it has been called the Chicobi sediments (MERQ-QGS 1984). A wacke sample (sample # 97JAA-60) was collected for geochronology from Purvis Township (see Figure 4.1). Six zircon grains with different appearances were extracted and analyzed from the sample. The youngest zircon yields an age of 2698.8 ± 2.4 Ma (see Figure 4.2l), suggesting it is the maximum age for deposition of the sediment and indicating correlation within the newly defined Porcupine assemblage. Older ages are also evident from the other zircon populations with ages of 2715, 2725, and 2745 suggesting provenance from the Kidd–Munro, Deloro, and Pacond assemblages respectively, and a pre-Abitibi greenstone belt zircon age of 2825. The oldest zircon was most probably derived from gneissic rocks of the Opatica Subprovince (Davis et al.
1994). Taken together, the oldest and youngest detrital zircon ages in this sample, imply the Abitibi and Opatica subprovinces were juxtaposed (accreted together?) by the time of deposition of the Porcupine assemblage. This new data also indicates Mueller and Donaldson's (1992) inclusion of this unit within the oldest sedimentary cycle of the Abitibi greenstone belt (i.e., cycle 1) with ages of greater than 2720 Ma is invalid.

**Timiskaming Assemblage**

Within the map area, the Timiskaming assemblage is restricted to a narrow band of clastic sedimentary rocks deposited unconformably on older assemblages in proximity to the Porcupine Destor Fault (see Figure 4.1). The assemblage consists of polymictic conglomerate and sandstone deposited in subaerial alluvial-fan, fluvial and deltaic environments (Born 1995, Mueller, Donaldson and Doucet 1994). Alkaline volcanic rocks occur as intercalations with the sedimentary units in the vicinity of Kirkland Lake (Jackson and Fyon 1991, Mueller, Donaldson and Doucet 1994). A sandstone sample from northeastern Whitney Township yielded two distinct zircon populations at 2692 and 2679 ± 3 Ma, the former similar in age to many of the large plutonic complexes in the Abitibi greenstone belt (see below) and the latter providing an upper age limit for deposition of the Timiskaming assemblage (Corfu 1993).

**PLUTONISM, SEDIMENTATION AND DEFORMATION**

Widespread geochronological evidence from across the Abitibi greenstone belt (see Corfu 1993, Heather, Shore and van Breeman 1996) indicate that the plutons fall into 3 broad suites: 1) synvolcanic plutons with similar compositions and ages to the volcanic assemblage; 2) syntectonic intrusions that range in age from about 2695 to 2680 Ma. They are massive to foliated and occur within the large marginal bimodal complexes and as rounded plutons within the supracrustal rocks. The earliest intrusions are the most voluminous and range in composition from tonalite to granodiorite. Later intrusions are more potassic and range from syenitic and lamprophyric compositions to granite; 3) post tectonic intrusions range in age from about 2665 to 2640 Ma. They are massive, and consist of granite and aluminous S-type granites.

The similarity in age between the large internal plutons and the external batholiths with the depositional age of the Porcupine assemblage strongly suggests a genetic correlation. Emplacement of these large plutons in the upper crust was temporally related to early regional folding and faulting in the Abitibi greenstone belt (Corfu 1993, Heather, Shore and van Breeman 1996). This plutonism and associated tectonism probably resulted in the uplift, emergence, erosion, and deposition of the Porcupine assemblage sediments. The sediments were deposited in localized, fault-controlled submarine basins around 2695 Ma. About 10 to 20 million years later, a predominantly potassic igneous event was coincident with further uplift and deposition of the Timiskaming assemblage sediments in largely subaerial, fault-controlled basins at about 2685 to 2675 Ma. The close spatial association of the Timiskaming assemblage units with regional-scale faults such as the Porcupine Destor Fault also implies the sedimentation was related to faulting. It is likely that the sediments (± alkaline volcanism) occurred in pull-apart basins where the displacement was predominantly strike-slip, and in grabens where it was dip-slip (Mueller, Donaldson and Doucet 1994, Mueller et al. 1997). It is also possible that the crustal-scale fault systems provided conduits for deep-seated potassic magmatism and the hydrothermal solutions associated with gold mineralization.

**TECTONIC SUMMARY**

A complete tectonic synthesis of the Abitibi greenstone belt is premature at this time, but the belt-wide lithological and geochronological data support a relatively coherent largely autochthonous stratigraphy with distinctive rock suites at specific time intervals (e.g., Goodwin 1977, Pyke 1982, Jensen and Langford 1985, Heather, Shore and van Breeman 1996). This contrasts with recent mobilist models for the Abitibi greenstone belt, in which lithotectonic entities represent variable geodynamic environments allochthonously juxtaposed by collisional plate tectonic processes (Jackson, Fyon and Corfu 1994, Mueller et al. 1997).

However, a simplistic uniform thickness "layer-cake" autochthonous stratigraphy is not completely satisfactory either. The composite stratigraphy discussed above represents a crustal thickness in excess of 20 km. This is unrealistically thick given that even in the lowest parts of the stratigraphy, regional metamorphic grades do not exceed upper greenschist facies (Jolly 1978, Powell, Carmichael and Hodgson 1995). There is also evidence for extensive stratigraphic gaps in areas where there are good bedrock exposures and states of preservation. Examples include the Kidd–Munro assemblage which occurs as a belt-wide unit north of the Porcupine Destor Fault, but is not evident in the central or southeastern parts of the Abitibi greenstone belt (see Figure 4.3). Out of sequence stratigraphy with older units lying above younger units has been detected in a few localities such as the Larder Lake Group overlying the Skead Group in the Kirkland Lake area (see Figure 4.3) (Jackson, Fyon and Corfu 1994) and rocks of the Deloro Group overlying the Tisdale in the southern part of the Shaw Dome (Barrie and Corfu this volume). These stratigraphic complexities are most probably the result of the local folding and faulting, rather than large scale plate tectonic dislocations. There is considerable variation in the thickness of the individual assemblages in different parts of the Abitibi greenstone belt. For example, in the central part, the Kinojevis and the Blake River assemblages are in excess of 10 km and 5 km thick respectively (see Figure 4.1). However, north of the Porcupine Destor Fault, the Blake River assemblage, as represented by the Krist formation, is only several hundred metres thick and is underlain by the lower komatiitic to tholeiitic part of the Tisdale assemblage (see Figure 4.3),
indicating the absence of the Kinojevis assemblage. Stratigraphic evidence of this nature indicates that either the assemblages were thickest proximal to volcanic centres and thin to absent in more distal parts of the basin, or they were removed, or not deposited, as a result of tectonic processes which are as yet poorly understood.

By analogy with modern day plate tectonic geodynamics, the chemical patterns of the Abitibi greenstone belt komatiitic and tholeiitic rock series suggests plume-dominated, ensimatic oceanic rift environments while the calc-alkaline rock series suggest island arc collisional environments (Ludden, Hubert and Gariety 1986, Campbell, Griffiths and Hill 1989, Jackson, Fyon and Corfu 1994). However, autochthonous superposition of repetitions of these distinctive geodynamic environments over the 70 million year history of the Abitibi greenstone belt (i.e., 4 predominantly mafic (± ultramafic) assemblages and 3 predominantly calc-alkaline assemblages) and the extensive intermixture of the different rock series in a number of the assemblages, has not been documented in modern plate tectonic environments. Clearly if Archean plate tectonics were responsible for formation of the Abitibi greenstone belt assemblages, the processes were quite different from those of modern day geodynamics.

**METALLOGENY**

Geochronologic results indicate that there are four distinct volcanogenic massive sulphide-bearing assemblages in the Abitibi greenstone belt (Ayer et al. 1997) the Deloro, Kidd–Munro, Tisdale and Blake River assemblages. Two distinct types of VMS mineralization are recognized within the Deloro assemblage; (1) massive sulphide deposits in mafic to felsic calc-alkaline volcanic rocks spatially associated with FII type felsic volcanic rocks and synvolcanic intrusions of Lesher et al. (1986). A good example of this deposit type is the Normetal deposit of northwestern Quebec, which yielded about 11 Mt of 5% Zn and 2% Cu, and (2) base metal mineralization in sulphide facies iron formations representing proximal exhalative mineralization localized within regional-scale oxide facies iron formations. The Shusby deposit in the Swayze belt is an example of the latter VMS type (Heather 1998).

The second oldest VMS epoch occurs within the Kidd–Munro assemblage. Deposits include the giant Kidd Creek Mine with over 150 Mt of Cu, Zn and Ag ore and a number of smaller deposits to the east in Munro Township (Barrie and Corfu this volume). Results of our new geochronology indicate that the Kidd-Munro assemblage can now be extended west of the Mattagami River fault to the Kapuskasing Structural Zone (see Figure 4.1). This relatively poorly explored area may have good potential for base metal deposits as it appears to be a sinistral offset continuation of the base-metal rich part of the assemblage east of the Mattagami River fault.

The Tisdale assemblage contains a number of base metal deposits including the deposits at Kamiskotia and within the Val d’Or formation in Quebec (Corfu 1993). Thus the Tisdale is considered to be an economically significant stratigraphic unit throughout the Abitibi greenstone belt. Newly identified base metal mineralization at Sheraton Township (Vaillantcourt this volume) indicates that the Tisdale assemblage is host to base metal mineralization south of the Porcupine Destor fault.

The final VMS episode occurs in the Blake River Assemblage. The abundance of base metals associated with highly elevated precious metals in deposits such as the Horne Mine in Noranda, Quebec and the gold deposits to the east, has led Robert and Poulsen (1997) to suggest that some of the gold mineralization in the Blake River Group was related to volcanism. Base metal or gold deposits of this age have not been found in Ontario to date, but the Blake River Group extends west of Noranda and are also found in other parts of the assemblage scattered throughout the Abitibi greenstone belt (see above).

Theories about the origin of gold mineralization in the Abitibi Subprovince have gone through many incarnations from the dominantly structural models up to the late 1070s to a syngentic, exhalative model comparable to that of VMS deposits in the late 1970s and early 1980s, and finally to the present diversity of three different deposit types including (Robert and Poulsen 1997): (1) the exhalative, syn-volcanic type associated with stratabound base metal sulphides (e.g., Horne and Bousquet deposits), (2) intrusion related deposits (e.g., Hollinger–McIntyre and Matachewan deposits) and (3) mesothermal vein type deposits (e.g., Dome and Kerr Addison).

Type 1 deposits appear to be most abundant within the Blake River Group in Quebec, but rocks of this age also occur in Ontario (see above) and may also be prospective for the synvolcanic type of gold deposits. The significantly-sized gold deposits of types 2 and 3 have a strong spatial relationship to crustal scale fault systems such as the Porcupine Destor fault (Hodgson and Hamilton, 1989). Geographic correlation also occurs between these types of gold deposits, the crustal fault systems and the rocks of the Timiskaming Assemblage and are probably genetically related to movements on these breaks.

There may also be a more subtle spatial correlation of the older and more widespread Porcupine assemblage turbidites sedimentary sequences with faulting and gold mineralization. The spatial association of these older sedimentary units with the regional breaks suggests that they may represent distal sedimentation related to early faulting while the Timiskaming Group may represent proximal sedimentation related to late stage reactivations along the crustal-scale fault systems.

**REFERENCES**


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5. Project Unit 97-024. Geological Investigations Along Highway 101; Guibord, Michaud and Garrison Townships

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INTRODUCTION

This multi-year project is designed to improve the geological database along Highway 101, especially in areas covered by overburden in the vicinity of the Porcupine–Destor deformation zone (PDDZ). This report summarizes geochronology from samples collected last year (Berger 1997, Ayer this volume) and reports observations from this year’s field work in Guibord, Michaud and Garrison townships.

GEOCHRONOLOGY

Two samples collected for geochronology in the area mapped last year (Berger 1997) were submitted to the Jack Satterly laboratory, Royal Ontario Museum for analysis. Data are presented on Figure 5.1 and Figure 5.2.

A concordant U-Pb zircon age date of 2716 ± 1.5 Ma was obtained for calc-alkaline intermediate tuff breccia from diamond drill core northeast of the Holloway gold mine (see Figure 5.1). The tuff breccia is part of a calc-alkaline metavolcanic sequence that extends from Quebec to west of Rand Township. This date is correlative with the Kidd–Munro assemblage (Corfu 1993, Jackson and Fyon 1991) and indicates that tholeiitic and calc-alkaline subdivisions are present. A 2716 Ma date also indicates that the calc-alkaline tuff breccia can not be correlated with the Hunter Mine group (ca 2730 Ma) as previously suggested (Jensen and Langford 1985).

A second sample was collected from flow banded rhyolite interlayered with mafic flows correlated with the tholeiitic Kinojevis assemblage in southern Harker Township (Jensen and Langford 1985, Jackson and Fyon 1991)(see Figure 5.1). A concordant U-Pb zircon date of 2701 ± 1.2 Ma was obtained and confirms previously reported dates for Kinojevis assemblage volcanism (Corfu 1993). Mapping indicates that the felsic rocks are extensive units and are laterally transitional to variolitic flows in southern Holloway Township. The flows occur near the contact with the calc-alkaline Blake River assemblage and it is possible that the felsic flows can be used as a marker to signal proximity to this contact.

GENERAL GEOLOGY

This summer’s mapping and compilation in Guibord, Michaud and Garrison townships builds on fieldwork by Prest (1951a), Satterly (1947, 1949) and Troop (1989, 1990) and results are generalized in Figure 5.3. As much of this area is covered by overburden, mapping relied on extensive use of archived and private company diamond drill core. Members of 5 Neoarchean metavolcanic-meta- sedimentary rock assemblages as identified by Jackson and Fyon (1991) are intruded by a distinctive Archean syenite-related suite of ultramafic, mafic, intermediate and felsic plutonic rocks. Paleoproterozoic diabase dikes and rare Jurassic/Cretaceous kimberlite dikes and plugs intruded these rocks. The Archean rocks were tilted to near vertical positions then sheared and folded about westerly to northwesterly trending axes. The Porcupine-Destor deformation zone is the most well known structure in the area, however, several other related deformation zones are inferred to fan westward from Garrison Township. Subsequently, late Archean north-northwest to north-northeast brittle and ductile faulting occurred and reactivation along these structures is inferred at several locations. Gold mineralization is related to the emplacement of the syenitic rocks, to the formation of the major east-west striking deformation zones and to the late Archean brittle to ductile faults.

Kidd–Munro Assemblage

Ultramafic, mafic and intermediate metavolcanic rocks with subordinate related ultramafic and mafic sills of the Kidd–Munro assemblage underlie the northern part of the map area. Massive and pillow mafic flows are predominant, hyaloclastite and pillow breccia are minor. A distinctive variolitic unit that contains pods and bands of coalesced varioles in a high iron tholeiitic groundmass occurs at the base of the assemblage in northern Guibord and Michaud townships and extends at least 5 km to the northwest. Gold mineralization occurs in this unit at the “main zone” on the Pangera property and at the contact with syenite in central Michaud Township (Figure 5.3). It is an important marker and indicates that the Kidd–Munro assemblage is an important host for gold mineralization east of Matheson.

Komatiite flows occur along the north contact of the PDDZ in Garrison Township and as a northwest striking
unit over 1 km thick in Michaud Township. Spinifex and cumulate ultramafic flows are most common, basaltic komatiite, variolitic high magnesium basalts; flow breccia and schist are subordinate. Strongly carbonatized, serici-tized and pyritic ultramafic rocks occur as sheared pods and schistose units at the contact with and as structurally interleaved units within Hoyle assemblage turbidites. The northwest striking unit extends into Munro Township where nickel mineralization is reported (Johnstone 1991). The komatiitic rocks are poorly understood due to poor exposure and limited exploration in this part of the map area.

Andesitic and dacitic lapillistone, tuff and flows are in stratigraphic contact with tholeiitic basalts in northern Garrison Township. The intermediate rocks extend east to

Figure 5.1. Location map for geochronology samples collected in 1997.
Holloway Township where they are dated at 2716 Ma, correlative with the Kidd–Munro assemblage (see above, Berger 1997). The intermediate rocks are calc-alkalic and indicate that the Kidd–Munro assemblage is divisible into tholeiitic and calc-alkalic subdivisions in this part of the Abitibi subprovince.

Peridotite and gabbro sills in northern Garrison Township are similar to but are not contiguous with the Center Hill or Munro–Warden intrusions in McCool and Munro townships. Cumulate peridotite passes into gabbro and quartz gabbro, pyroxenite is rarely reported. Previous exploration concentrated on the extensive asbestos mineralization in these sills. The nickel and platinum group element (PGE) potential of the intrusions appears to be under explored.

Stratigraphy generally is northyounging as indicated by pillow facings, however, south facing pillows were reported by Satterly (1947) and Troop (1990) in Garrison Township.

Hoyle Assemblage

The Hoyle assemblage is composed of clastic and minor chemical metasedimentary rocks and is restricted to the northeastern part of Guibord Township (see Figure 5.3). A, D, E and very rarely C divisions of the Bouma sequence (cf Walker 1992) were observed in outcrop and drill core indicating deposition by turbidity currents. Greater proportions of siltstone and mudstone than previously observed (Berger 1992, 1996) are indicative of distal facies turbidites. Bedded pyritic chert and graphite occur at the southern contact with heterolithic mafic breccia at Guibord Hill. Syenite and syenite-related feldspar porphyry dikes intruded the metasedimentary rocks and ultramafic flows are structurally interleaved with mudstone on the Pangea property.

Graded beds and load casts indicate the turbidites young consistently to the north. Prest (1951b) and Johnstone (1991) concluded that the metasedimentary rocks were older than the Kidd–Munro assemblage based on the stratigraphic facings. However, neither worker recognized a 30 to 50 m wide deformation zone in the Hoyle assemblage where folding and transposition of bedding indicates that the contact with the Kidd–Munro assemblage is structural rather than stratigraphic. Recent geochronology indicates that the Hoyle assemblage in the Timmins area is 2699 Ma, much younger than the Kidd–Munro assemblage (Bleeker and Parrish 1996).

Bowman Assemblage

Mafic and ultramafic metavolcanic rocks south of the Hoyle assemblage and included within the PDDZ are tentatively assigned to the Bowman assemblage. This correlation is based on continuity of the metavolcanic units west of Guibord Township to the type area and on the presence of ultramafic metavolcanic rocks which precludes correlation with the Kinojevis assemblage (Jensen and Langford 1985, Jackson and Fyon 1991). Ultramafic rocks composed of spinifex, flow breccia and cumulate textured flows are interleaved with talc-chlorite-carbonate schist and mafic metavolcanic rocks in southern Guibord and south central Michaud townships (see Figure 5.3). Ultramafic flows and schist also occur north of the Garrison Stock and extend east into Harker Township. These rocks are hydrothermally altered to various degrees and intruded by mafic to felsic alkalic dikes and plutons associated with gold mineralization.

Figure 5.2. a) 97JAA-104 Rhyolite Kinojevis Iris prop. SE Harker Twp; b) 97JAA-102 Intermediate tuff–bx debris flow NW Holloway Twp.

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Figure 5.3. Geology of Guibord, Michaud and Garrison townships.
Pillowed and massive mafic flows are common in central Michaud Township and occur as narrow mappable units throughout central Guibord Township. A distinctive heterolithic breccia that contains phaneritic and aphanitic mafic clasts, wacke, argillite and pyritic mudstone clasts and rare feldspar porphyry and felsic metavolcanic clasts occurs at Guibord Hill. Prest (1951b) and Johnstone and Trowell (1985) concluded that the breccia was intrusive, however, the general lack of reaction rims on the clasts, the increasingly sandy matrix towards the top of the unit and its general conformity with stratigraphy suggest that the breccia may be a basal conglomerate or fault scarp deposit. The breccia is remarkable for its elevated pyrite content as both disseminated (up to 30%) and massive beds near the base are present. However, extensive exploration has encountered only low base and precious metal values.

Variolitic flows of high magnesium basalt composition are generally restricted to the western part of Guibord Township.

Reversals in stratigraphic facings suggest that the assemblage is either anticlinally folded about an east-west axis or is domed around syenitic intrusions in Guibord and Michaud townships.

Kinojevis Assemblage

The Kinojevis assemblage underlies the southern part of the map area and is composed predominantly of massive and pillowed mafic metavolcanic rocks. Several elastic interflow metasedimentary units occur toward the base of the assemblage; the thickest unit is up to 500 m in southeastern Garrison Township (see Figure 5.3). Felsic metavolcanic rocks farther east are not present in the map area (Berger 1997). Variolitic flows are common throughout the assemblage. Pillow facings and graded beds indicate that the assemblage consistently youngs to the south. The PDDZ marks the contact with the Timiskaming assemblage and airborne magnetic data indicate that layer parallel faults and cross faults have disrupted the Kinojevis assemblage.

Timiskaming Assemblage

For the purposes of this report the Timiskaming assemblage includes all metasedimentary rocks that occur within the PDDZ and all syenite related intrusive and extrusive rocks in the map area. The metasedimentary rocks occur as an arcuate band extending from central Garrison southwest through Michaud and west through Guibord townships and are known mostly from diamond drill core (see Figure 5.3). Within this band there are discontinuous remnants of an older metasedimentary rock package characterized by turbidites and laminated magnetite-chert and hematitic iron formation that are not strictly Timiskaming (cf. Jackson and Fyon 1991). The thickest iron formation is preserved in southeast Michaud Township; however, thinly bedded iron formation was observed in Garrison and Guibord townships. The interbedded turbidites are fine-grained, thinly bedded and are interpreted to be distal deposits in a quiet water environment.

The younger metasedimentary rock package composed of polymictic conglomerate, grit, sandstone and mudstone unconformably overlies and has eroded the iron formation. Iron formation and jasper clasts occur within the conglomerate and sandstone beds as do ultramafic and mafic metavolcanic, syenitic, rare albitized and sulphide clasts. Recognition of the unconformity is critical because gold mineralization is localized at or near this contact in Michaud Township. Further, the predominance of sandstone and mudstone over conglomerate combined with small average clast size (3 to 5 cm) and absence of large scale cross bedding suggests deposition in a low energy fluviatile environment, possibly deltaic.

Numerous reversals of stratigraphic facings indicate that both of the metasedimentary packages are folded. Numerous intervals of schist and gouge in the drill core indicate shearing and faulting are also common. For the most part contacts between the metasedimentary and surrounding metavolcanic rocks are sheared or intruded by diabase or syenitic dikes. No clear evidence for the stratigraphic position of the Timiskaming was observed.

Alkaline intrusive rocks ranging in composition from pyroxenite/hornblende to granite intruded the supracrustal assemblages as plutos, dikes, sills and cupolas and are included in the Timiskaming assemblage based on the 2678 ± 2 Ma age for the Garrison Stock (Corfu and Noble 1992). The Garrison Stock is weakly foliated, equigranular hornblende granite that has a distinct low airborne magnetic pattern associated with it. Pyroxenite and biotite lamprophyre occur as xenoliths and dikes in the granite and as dikes in ultramafic metavolcanic rocks along the north margin of the stock. Granitic and monzonitic dikes intruded into variolitic basalts are associated with gold mineralization at the New Buffonta gold deposit near the west contact of the pluto.

The alkalic plutonic complexes in Michaud and northern Guibord townships are more extensive than previously mapped (Prest 1951a, Satterly 1949) and it is possible that they connect at depth (see Figure 5.3). Hornblende, mela-syenite, biotite-amphibole diorite and lamprophyre occur along the northern and eastern margins of the plutos and on the structural footwall to gold mineralization on the Pangea property. Textures vary from pegmatitic, porphyritic, equigranular to aplitic over narrow intervals, which is interpreted as indicating crystallization at high crustal levels probably in subvolcanic magma chambers. These rocks contain greenstone and rare olivine and garnet bearing ultramafic xenoliths that were possibly mantle derived and are surrounded by pronounced mica reaction rims. Felsic syenite, feldspar porphyry and monzonite dikes and pods indicate that the mafic rocks are the oldest members of the plutos. Extensive portions are pervasively hematitized and epidotized, which is interpreted as late stage hydrothermal oxidation of these rocks.

The mafic members of the plutos are intruded by and are transitional into monzonite, syenite and granite. These rocks are most commonly equigranular although porphyritic, pegmatitic and aplitic phases are locally abundant (e.g., Ludgate Lake, Figure 5.3). Disseminated pyrite
(trace to 10%) is widespread and commonly contains geochemically anomalous gold (>30 ppb). Potentially economic gold mineralization occurs where the felsic rocks are sheared, hydrothermally altered and sulphidized. Numerous felsic dikes characterize the southern contacts of these plutons with the supracrustal rocks.

Recent diamond drilling southeast of the main gold zone on the Pangea property encountered mafic and felsic alkaline laminated flows and flow breccia interlayered within monzonite. The flow rocks contain feldspathic pseudomorphs of octagonal crystals inferred by the author to have been primary leucite or analcite. This is the first report of possible foid bearing extrusive rocks in this area and strengthens previous analogies to the Kirkland Lake area (Berger 1997). The flow rocks also contain amygda loidal and fracture controlled fluorite which raises the possibility that the original magma was fluorine rich and related to topaz rhyolites. As such there may be as yet untested potential for tin and tungsten mineralization in this area.

Equigranular and porphyritic granite and syenite comprise a separate pluton in eastern Guibord Township (see Figure 5.3). Biotite lamprophyre appears to be localized near the southern contact with the Bowman assemblage and as narrow dikes within the pluton. Hematized feldspar porphyry dikes similar to that observed along the Pike River in western Guibord Township cut the granite. Extensive pervasive and fracture controlled hematite occurs near the southern contact with the PDDZ.

Numerous syenite, feldspar porphyry, diorite and lamprophyre dikes and small plutons (up to 50 km km by 0.25 km) intruded the Bowman, Hoyle and metasedimentary rocks of the Timiskaming assemblages. They are less common in the Kidd–Munro assemblage and rare in the Kinojevis assemblage. These intrusions most commonly occur at contacts between rock types and have in some places been interpreted as part of the stratigraphy (assessment files, Resident Geologist's Office, Kirkland Lake).

Kimberlite occurs at the New Buffonta pit in Garrison Township, west of the Ludgate Lake gold deposit in Michaud Township, within the syenitic pluton in eastern Guibord Township, as three possible separate occurrences in southern Guibord Township (Buzz #1, #2, #3) and on the Tandem Resources Limited property to the northwest of the Buzz occurrences. Sage (1996) describes all occurrences except the last. The Tandem occurrence is a small kimberlite plug approximately 100 by 75 m in size. Diamond drill core examined by the author contains abundant olivine, garnet, Paleozoic limestone, greenstone and rare eclogite xenoliths. Diamond and ruby crystals are reported in the kimberlite, however, the economic potential of the intrusion is still under investigation.

**STRUCTURAL GEOLOGY**

The structural history of the area is complex and commenced with the tilting of the Neoarchean assemblages into near vertical positions apparently without development of any penetrative foliation. The tilting was prior to or synchronous with intrusion of the alkalic rocks and the start of deposition of the Timiskaming metasedimentary rocks.

Development of the Porcupine–Destor deformation zone (PDDZ) followed crystallization of the alkalic plutons as they are commonly foliated and sheared parallel to the main faults. At the Quebec border the PDDZ is confined to a zone 200 to 400 m wide that gradually widens towards the Holloway and Holt–McDermott mines (cf. Berger 1997). In the map area the PDDZ is interpreted to fan west of the Garrison stock into a number of discrete west, southwest and northwest striking faults each separated by weakly deformed rocks. The main Porcupine–Destor fault is a wide deformation zone broadly coincident with its previously mapped position (Prest 1951a, Satterly 1947, 1949, see Figure 5.3). Northwest striking faults in northern Michaud Township are coincident with the regional Pipistone deformation zone. Westerly striking faults at Ludgate Lake and the Pangea property (see Figure 5.3) appear to be narrow and of limited extent but are nevertheless associated with gold deposits. Explorationists familiar with the map area have long inferred the existence of a west-southwest striking fault not previously portrayed on government maps (“Arrow fault” see Figure 5.3). This structure is characterized by an airborne magnetic lineament that extends from Garrison Township to west of the map area. Northwest striking members of the Kidd–Munro and Hoyle assemblage are terminated and members of the syenite intrusion suite display change in width or strike across this structure. Limited diamond drill testing encountered wide zones of low tenor gold mineralization (0.2 g/ton Au over 124 m) in central Michaud Township; however, much of the structure remains poorly explored.

Folds in the Kidd–Munro, Hoyle and Timiskaming assemblages have axial planar cleavage parallel to the faults of the PDDZ. This suggests that the folding is broadly synchronous with development of the PDDZ and that shearing most likely results from over flattening along the hinges and limbs of the folds (cf. Linnstra and Benn 1997, this volume).

North-northwest to north-northeast striking faults offset the PDDZ with vertical and horizontal displacement of commonly less than 100 m. Many of these faults are characterized by pronounced airborne magnetic lineaments and for the most part are brittle with calcite and limonite healed fractures. However, mylonite zones, crenulation cleavage and penetrative foliation indicates that some of the faults are ductile, which has not been observed previously (Berger 1997, 1996).

**MINERALIZATION**

Gold was extracted from the Buffonta, New Buffonta, Garrcon and Jonpol deposits in Garrison Township. Potentially economic deposits occur on the Pangea property, at Ludgate Lake and on the Moneta Porcupine property in southeast Michaud Township based on reserves and ore grades published by the companies. Significant mineralization where reserves have not been reported
occurs on the Neal and Windjammer properties and there are several areas where low grade gold mineralization (approximately 1 g/t) is reported (see Figure 5.3).

Although each gold occurrence has unique properties there are several common features, which suggest that mineralization is related to 1 or 2 events. Almost all occurrences are spatially associated with alkaline rocks and are within or near major faults or shear zones of the PDDZ. The Ludgate Lake deposit is hosted within sheared syenite whereas the Buffonta, New Buffonta and Pangea deposits are hosted in altered variolitic basalts at the periphery of alkaline plutons. Syenitic or lamprophyric dikes are associated with the Neal, Moneta, Garrcon and Brydges mineralization. Hydrothermal alteration characterized by hematite, albite, carbonate, sericite and silica is common at most occurrences. Disseminated pyrite is ubiquitous. Many of the alkaline rocks and especially the Garrison Stock and small intrusions on the Moneta property (see Figure 5.3) contain geochemically elevated gold even in the absence of pyrite suggestive that the intrusions are primary gold sources.

The following paragenesis was observed locally where alkaline rocks are the primary host to mineralization. Disseminated pyrite that the author inferred to be magmatic is auriferous and commonly accompanied by pervasive albization. The rocks were then hematized and pyrite was corroded as evidenced by hematitic rims and grain size reduction. A second albization event accompanied by quartz and pyrite stringers with or without sericite veining was imposed on the rocks and is commonly accompanied by ore-grade gold mineralization. Where supracrustal rocks are the primary host, hematite and albization accompanied by disseminated and vein pyrite are most evident. It appears crucial that pyrite veining with or without quartz and sericite is necessary to achieve potentially economic gold tenor. These observations support and expand the model proposed by Robert (1997).

The one important exception is on the Moneta property. At this location, gold mineralization occurs where hydrothermally altered Timiskaming conglomerate unconformably overlies hematized magnetite-chert iron formation. Pervasive carbonate, sericite, silica and disseminated pyrite alteration in the conglomerate is cut by vein quartz and pyrite that locally penetrates the iron formation. Gold occurs in the altered conglomerate and later quartz-pyrite veins. The conglomerate probably acted as a permeable conduit for hydrothermal fluids derived from nearby alkaline intrusions and the iron formation acted as a suitable reductant that promoted gold deposition (cf. Colvine et al. 1988).

Gold occurrences at Buffonta, the “jasper zone” near Ludgate Lake and the former Croesus mine in Munro Township indicate that mineralization is associated with late structures orthogonal to the PDDZ. Gold bearing quartz veins are most common in these structures but a north-northeast striking mylonite at Ludgate Lake indicates ductile deformation occurred in contrast to brittle deformation on similarly oriented faults elsewhere in the region.

**RECOMMENDATIONS TO PROSPECTORS**

Gold is associated with alkalic rocks and the PDDZ. The intersection of major structures and plutons is a favourable environment for gold mineralization. Explorationists should be aware that mafic and felsic phases of the alkaline plutons are equally favourable targets. Further, the mafic phases warrant examination for their PGE content especially if disseminated chalcopyrite is present. Recognition of foid bearing alkalic extrusive rocks on the Pangea property strengthens comparison to the Kirkland Lake area and the presence of fluorspar in these rocks raises the possibility of tin and tungsten mineralization analogous to topaz rhyolites.

Recognition of an unconformity between Timiskaming conglomerate and iron formation is crucial because gold mineralization occurs at this contact. The mineralization has a primary stratigraphic control that requires treating the metasedimentary rocks as 2 separate packages that probably underwent different structural histories.

Gold is localized in late north-northeast to north-northwest striking faults orthogonal to the PDDZ. Explorationists should be aware that such faults might be auriferous in the map area.

Diamond and ruby bearing kimberlite occur in the map area. The kimberlite intrusions are generally localized along northwest striking structures parallel to the Timiskaming Structural Zone (cf Sage 1996) and there is close spatial association with the Neoarchean alkaline rocks. Magnetic surveys failed to resolve most kimberlite intrusions from the surrounding rocks so exploration programs will have to rely on techniques other than geophysics.

**REFERENCES**


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INTRODUCTION

Structural investigations of the Holloway mine and vicinity are part of a collaborative project involving Battle Mountain Gold, the Ontario Geological Survey and the University of Ottawa. The principal goals of the project include: 1) identification and description of the major structural features within the deposit, 2) determination of the structural controls on the geometry and distribution of the ore zones, and 3) deciphering the timing relationships of major alteration and mineralization events with respect to the structural evolution at the local and regional scales. The Holloway Mine lies on a splay of the Porcupine–Destor Deformation Zone (PDDZ). Figure 6.1 shows the location of the study area with respect to major geological boundaries. A summary of the principal rock types in the mine sequence can be found in Luinstra and Benn (1997). The stratigraphic group to which the mine sequence belongs is not yet definitively established. Historically, the mine stratigraphy was considered to be part of the Destor–Porcupine Complex (Jensen and Langford, 1985).

Based on rock types and their distributions, Guy (1996) proposed that the mine sequence is part of the Hunter Mine Group. More recent mapping by Berger (this volume) suggests a possible of correlation with either the Stoughton Roquemaure Group or the Kidd–Munro Group.

A first field season of structural mapping and sampling was completed in the summer of 1997. During the ensuing autumn and winter, field observations and

![Diagram](image)

Figure 6.1. Location of the Holloway Mine.
structural data were synthesized and analyzed, and petrographic-microstructural analyses were completed at the University of Ottawa. During the 1998 field season (May 1 to September 1), further detailed structural mapping was carried out underground at the mine, as well as on the surface in the vicinity of the mine. Additionally, a detailed lithological and structural investigation of stored drill core was carried out. Access to the drill core was provided by Battle Mountain Gold.

Ore grade material in the Holloway deposit is generally confined to a highly albitized hyaloclastite unit. This hyaloclastite unit sits structurally above a 5 metre-wide, steeply-dipping shear zone, named the Footwall Contact Shear Zone (Guy 1995). The Footwall Contact Shear Zone affects an ultramafic volcanic unit which is structurally below, and in contact with, the ore zone. Locally, spinifex textures and breccias are preserved within the sheared ultramafic unit. The ore zone is structurally overlain by a sequence of massive and pillowowed mafic volcanic rocks with occasional interflow sedimentary units, generally greywackes interbedded with shales. One particularly interesting unit is a coarse grained heterolithic breccia which, in structural profiles, appears to crosscut the ore body. However, detailed investigations suggest the breccia is a stratigraphic unit and that the ore zones occur structurally both below and above the unit, implying that the ore zone is not confined to one particular stratigraphic horizon.

GEOLOGY AND STRUCTURE OF THE HOLLOWAY MINE

A penetrative S1 foliation occurs throughout the deposit (Figure 6.2), and strikes roughly east-west and dips to the south. Figure 6.2 shows that the penetrative fabric is essentially parallel to bedding throughout the deposit. This fact, in combination with the discontinuous nature of rock types, the presence of intrafolial folds and the high degree of flattening seen in pillowowed flows, suggests the presence of isoclinal folds, possibly accompanied by transposition of the stratigraphy. In the footwall ultramafic rocks, the S1 foliation is defined by the alignment of phyllosilicates, mostly fuschite. In the hangingwall mafic rocks, S1 is defined by flattened pillows, by the alignment of leucocene grains in massive volcanic rocks, by flattened varioles, and occasionally by flattened clasts within the breccia units. In the interflow sedimentary rocks, S1 is defined by the parallel alignment of micaceous minerals. The ore zone itself has a foliation, most evident in thin section, which is marked by the alignment of leucocene and pyrite, as well as by pressure shadows around pyrite grains. Within the ore zone, a cross-cutting intermineral dike was discovered that carries the S1 foliation.

The penetrative regional foliation is truncated at a low angle by the Footwall Contact Shear Zone. Veins which clearly cross cut S1 foliations are also truncated by the Footwall Contact Shear Zone. For this reason, the Footwall Contact Shear Zone is considered to be a D2 shear zone, though the possibility of movement along this fault late during D1 cannot be ruled out. Drag folds of the S1 foliation, asymmetrically sheared quartz veins, and deflection of veins which cross cut S1 into the shear plane suggest a normal sense of movement along the shear zone (south side down).

D3 folds can be found at all scales, ranging from open, megascopic-scale folds with half-wavelengths of 25 to 50 metres, to microscopic folds (crenulations) with half-wavelengths of 1 to 2 millimetres. Figure 6.2 shows the F3 axes are subhorizontal to shallowly east-plunging. A subhorizontal S2 axial planar crenulation cleavage is locally developed in association with F3 folds. The intensity of the S2 cleavage increases in proximity to larger-scale F3 folds. Several subhorizontal, brittle faults occur on the limbs of large F3 folds.

The deposit is also affected by two additional sets of brittle faults. A northeast-southwest trending fault can be followed in structural profiles through the entire deposit. North-south trending joints occasionally show minor offsets of marker horizons, and are consistently east side down.

GEOLGY AND STRUCTURE IN THE VICINITY OF THE MINE

The core from forty-five diamond-drill holes was studied in order to better determine the structural setting of the Holloway deposit. The core was relogged, paying special attention to structural features, the nature of contacts, the orientation of S1, and facing reversals. Structural profiles were constructed along six north-south lines traversing the deposit and the surrounding rocks. Figure 6.3 is a representative profile.

As previously discussed, the deposit itself is situated structurally above the Footwall Contact Shear Zone. Below the shear zone lies an ultramafic unit. This unit is characterized by heavily altered fuschitic and green carbonate-rich komatitites and ultramafic volcanic breccias. In locations where fuschitic alteration is less intense, it is found as a talc-chlorite schist. S1 within this unit is defined by the alignment of fuschite and other minerals, and is particularly well developed within the first five metres of the contact with the overlying units. S2 is well developed proximal to large F3 folds, which also affect the Footwall Contact Shear Zone. The contact with the overlying mafic volcanic rocks is complicated by veinings, yet always represents a sharp change from non-albitized (ultramafic) to albitized (mafic) units.

Description of the rock types within the mafic package which hosts the ore zone can be found above, and in Luinstra and Benn (1997). New observations from this field season and new definition drilling undertaken by Battle Mountain Gold, indicate the occurrence of several lenses of ore grade material away from the Footwall Contact Shear Zone and perhaps hosted by different lithologies. All contacts between ore bodies and hangingwall or footwall rocks are defined by small shear zones with varying kinematics. These contacts are generally parallel to S1 regional penetrative fabrics.
Structurally overlying the mafic package which hosts the main ore body is a sequence of greywackes and shales. The presence of detrital potassium-feldspars in these sediments, as well as the close spatial relationship with jasper bearing, polyolithic conglomerates suggest that they are Timiskaming type sediments (B. Berger, personal communication) Primary structures are indiscernible in this unit due to the intense D₁ and D₃ fabric development. The presence of small scale isoclinal rootless F₁ folds was noted in various holes, however the development of S₃ overprinting fabrics makes the recognition of F₁ folds difficult.

Figure 6.2. Schmidt projections of the principal fabrics in the Holloway Mine. S₀ = bedding; S₁ = main penetrative foliation; S₃ = crenulation cleavage; F₃ = axes of F₃ folds at all scales (including crenulations).
The greywackes and shales are overlain by a relatively thin band of conglomerate. The contact between this conglomerate and the underlying greywackes and shales is complicated by late ankerite veining and sericitization. The conglomerate is composed of a sand to pebble size groundmass rich in feldspar clasts and other mafic minerals. Larger clasts reach up to 10 cm and consist of a variety of rock types, including syenite, various granitic rocks, mafic volcanic rocks, fuschitically-altered ultramafic rocks and red jasper. The composition of this conglomerate suggests that it is of the Timiskiming type (Hyde, 1979; Born, 1995). Several albitized, mineralized clasts, contained within non-mineralized matrices were identified. This implies that at least one major albite alteration event and associated mineralization predated deposition of Timiskiming sediments in this area.

The conglomerates are characteristically a dull reddish colour, due to sericite-carbonate alteration. \( S_1 \) is well developed in these conglomerates, and is defined by the alignment of flattened clasts of all compositions. A pronounced, steeply-plunging \( L_1 \) extension lineation is defined by stretched clasts. \( S_3 \) crenulations are developed only where the sericitization is relatively intense, as is often the case close to lithological contacts.

Structurally above the conglomerates, and possibly repeated by \( F_1 \) folding, is another horizon of interbedded greywackes and shales. A 150 m thick horizon of conglomerates, above this second horizon of greywackes and shales, lies disconformably on top of (but structurally below) the Seagar’s Hill volcanic package (Berger, personal communication), which consists of highly deformed and heavily carbonated mafic pillowed and massive flows that are also found in outcrop south of the mine property. This disconformity predates \( D_1 \).

Another 50 to 75 m thick band of greywackes and shales structurally overlies the Seagar’s Hill volcanic rocks. \( S_3 \) within these rocks is much less intense and primary structures (graded bedding and cross laminations) are preserved. Overlying these greywackes and shales is Figure 6.3. A representative structural profile through the Holloway Mine. A, B and C are \( F_1 \) folds referred to in the text. Drill holes are labelled 96-23, 96-19, and 113.
another thick band of Timiskaming-type conglomerate similar to those described above.

Tops indicators (graded bedding, lode casts and cross stratification) from the uppermost greywacke and shale sequence were used to identify a major facing reversal due to an F1 fold (A in Figure 6.3). The conformable relationship of the Timiskaming type conglomerates with the Seagar's Hill volcanic rocks suggests that facings are downhole at that contact. Due to the general intensity of deformation, reliable tops indicators in the rest of the deposit are rare. However, Hyde (1979) and Born (1995) identified a fining upward sequence within the stratigraphy of Timiskaming type sediments that is also recognizable in drill core and can be used as a possible tops indicator. In addition, the presence of flow top breccias and hyaloclastites may be used as possible tops indicators, although secondary faulting along lithological contacts and flow contacts may also produce tectonic breccias that may be confused with primary breccias. Using these possible tops indicators and lithological distribution patterns, two more F1 folds were interpreted.

The first can be found in the middle of the second greywacke-shale unit (B in Figure 6.3). This fold is largely defined by the profile pattern of the rock types. The second interpreted fold (C in Figure 6.3) runs through the Holloway mafic package. A fourth F1 fold (not shown in Figure 6.3) can be interpreted from bedding cleavage relationships in surface outcrops which lie north of the projected surface expression of the Footwall Contact Shear Zone, on the mine property. The F1 folds are inclined, isoclinal folds with steeply west-plunging axes.

F1 folds and the Footwall Contact Shear Zone have been refolded by F3 folds. These open folds deform stratigraphy throughout the section producing a staircase pattern (see Figure 6.3), and are found on all scales. A late brittle fault, just above the lower contact of the conglomerate, is identified in all drill holes. The fault is identified by the presence of fault gauge, and rubbly core in drill holes, and cross cuts all previous structures. The sense of movement on the fault could not be determined.

ALTERATION

The deposit is pervaded by several stages and types of alteration. The reader is directed to Ropchan et al. (this volume) for a discussion of the chemistry of the different types of alteration. A significant portion of the field season was dedicated to determining the timing of major alteration events with respect to previously documented (Luinstra and Benn 1997) structures in the Holloway mine.

Pre-D1 Alteration

On the 505 level, a well foliated (S1) mafic dike was found that crosscuts all of the alteration and mineralization types discussed in this section. The dike was assayed and had no significant gold values. This dike can be considered to be an intermineral dike (Kirkham 1971). The pre D1 timing of alteration and mineralization events described below is therefore confirmed by the presence of this cross-cutting, well foliated, unmineralized dike.

The first major alteration event is characterized by hematite. The hematization is found only in rocks which have not undergone significant albitionization. This alteration rarely crosses lithological contacts, instead forming pockets restricted to the hyaloclastite unit. The hematite alteration is pre-D1, with S1 foliations defined by hematite within the hematized pockets. Additional evidence for an early hematization event comes from the occurrence of zones of albitionization that cross-cut previously hematized units.

Subsequent albitionization, which cross-cuts earlier hematization, can be split into two major styles. The first (Ab1) is defined by massive replacement of minerals by albite, forming what are essentially albitites. The second albitionization (Ab2), which cross-cuts Ab1 albitites, is characterized by small (<1cm) albite (+ quartz) stringer veins which have been deflected into and rotated parallel to S1. These albitic alterations may represent a single alteration event, however, for the purpose of this study, they are considered to be two distinct events, because of the observed cross-cutting relationships between Ab2 and Ab1.

There are two major pre-D1 pyrite mineralizations. Relative timing of these events with respect to previously discussed alteration styles will not be established with certainty until the petrographic analysis of polished sections have been completed. One pyrite mineralizing event (Py1) is defined by disseminated, large (up to 2cm) pyrite grains, while the other (Py2) is defined by small (<1cm) pyrite stringer veins. Py2 veins can be seen to locally cross cut Py1 pyrites and Ab1 albite veins. Both of these early pyrite mineralizations are confined to the ore-hosting lithology. Py1 pyrites have well developed pressure shadows which, along with aligned leucocenes, define a foliation parallel to S1 within the ore zone, indicating this event likely occurred prior to D1.

Guy (1996) interpreted that Au mineralization was associated with the Py1 and Py2 events based on occurrences and habits of gold grains. Gold grains commonly form on the crystal boundaries and within crystals of Py1 pyrites (Guy 1996). No new evidence has yet been found during this study to confirm Guy’s interpretation.

A sericite-carbonate-fuscite alteration event is considered to be syn- to pre-D1. Sericite associated with this event commonly defines the S1 penetrative fabric within the hangingwall sedimentary and volcanic rocks. The foliation within the intermineral dike is also defined by sericite associated with this event.

Post-D1 Alteration

A second sericite-carbonate-fuscite (Sr2) alteration event cross-cuts S1 fabrics. It is composed of haloes surrounding east-west striking, shallow north dipping quartz + ankerite veins within both hanging wall and footwall units, as well as in the host hyaloclastite unit. Occurring within haloes associated with this alteration is a third pyrite mineralizing event (Py3). Pyrite associated with Py3 mineralization are
large, disseminated grains confined to the alteration halo. Samples collected by Luinstra which are heavily altered by this event were assayed and show elevated values of both arsenic and gold. Locally, ore grade values have been found in hanging wall rocks affected by this alteration. In some cases, the Sr₂ quartz + ankerite veins appear to be en-echelon with respect to D₃ folds and faults. This, combined with the sub horizontal orientation of these veins, suggests that the Sr₂ event is related to D₃. This event may in fact have resulted in remobilisation of fluids during D₃.

FUTURE WORK

Additional samples were collected during the 1998 field season and will be processed for thin sections and polished sections. Microstructural analysis, focusing on the east lens of the Holloway deposit, will be completed over the winter. Polished sections will be studied to identify and describe the differences between the east and west lens ore bodies, and to verify interpretations, based mainly on field observations, of the timing of mineralization with respect to the structural history. Structural profiles through the deposit will be completed and tied-in with surface geology. The profiles should contribute to a better understanding of the third-dimension of the structural geology along this segment of the Porcupine–Detor deformation zone. They will also be useful for correlations of the mine geology with the larger-scale geological picture, and thus be invaluable for future interpretations of the stratigraphic, structural and tectonic setting of the Holloway deposit. All data and interpretations will be compiled in an MSc thesis to be completed the spring of 1999.

REFERENCES


INTRODUCTION

Field work was conducted by J. Ropchan and S. Nattress from late May until the end of August 1998. The objectives of the field season were threefold: 1) continue sampling and data analysis for J. Ropchan’s Master’s thesis on the Holloway Mine deposit, 2) collect samples for S. Nattress’ Bachelor’s thesis on Franco–Nevada and Greater Lenora’s Joint Venture property and 3) mapping of the southern half of Holloway Township for the Ontario Geological Survey (OGS).

The Holloway Mine and Franco–Nevada Joint Venture properties are located near Highway 101 approximately 15 km west of the Quebec border. The Franco–Nevada Joint Venture property overlies the down-dip extension of Barrick Gold Corporation’s Holt–McDermott ore zone. Holloway and Holt–McDermott Mines lie on opposite sides of Highway 101 approximately 1 km apart (Figure 7.1). It is not uncommon in the region to have two gold deposits in such close proximity, however, the contrasting styles of mineralization are noteworthy. Holloway ore is hosted within albite-altered tholeiitic rocks surrounded by haloes of sericite and carbonate alteration. Wherever hematite alteration occurs at the Holloway Mine, ore grades are insignificant but, at the Holt–McDermott Mine, ore is associated with albitic alteration (averaging 9 g/t), and with hematitic and siliceous alteration (averaging 7 g/t), but the highest grades found to date on the Franco–Nevada Joint Venture property are associated with hematitic alteration (Talbot 1998).

HOLLOWAY MINE

The Holloway Mine project is made possible through a partnership between the OGS, Battle Mountain Gold, and
the University of Ottawa. The project serves as a Master’s thesis for J. Ropchan who is studying the alteration and geochemical changes in host rocks of the deposit.

Holloway Mine rocks are characterized as a sequence of metakomatite to iron-rich tholeiitic metabasalts, associated with meta-hyaloclastite and variolitic-rich metabasalts, mafic and felsic dikes and metasedimentary rocks. The mine footwall is mainly comprised of ultramafic flows with some mixed tholeiitic flows and clastic sedimentary rocks. Guy (1996) describes the footwall contact as a structural discontinuity, probably an early fault, which he interpreted as a fluid conduit for the hydrothermal system. Ore zones at the Holloway Mine are hosted by a hanging wall sequence of iron-rich tholeiitic flows which are overlain by clastic sedimentary rocks. The volcanic rocks are probably correlative with Stoughton–Roquemaure Group or the Kidd–Muuro Assemblage, and the metasedimentary rocks with the Timiskaming assemblage (Jackson and Fyon 1991). Rock units at Holloway strike east-west and dip steeply to the south. Guy (1996) describes cleavage development and small and large-scale folding due to post-alteration deformation related to the Kenoran Orogeny (for details see Luinstra and Benn, this volume).

During the 1997 field season a suite of samples was collected from the West Lightning Zone of the Holloway deposit (Ropchan and Fowler 1997). A progression of samples collected leading into the ore zone comprise a spectrum from least to most altered specimens. Similar suites were collected during the 1998 field season from the East Lightning, Middle, Pumphouse and Blacktop mineralization Zones. Samples from a fine-grained dike were collected to aid in determining the paragenetic sequence of alteration and ore minerals. Lastly, samples were collected of a variably altered unit described as a coarse-grained heterolithic breccia. Debate on the true nature of this breccia is focussed on whether it represents a hydrothermal breccia or a volcanoclastic unit? This is one of the questions that will be addressed over the course of the thesis.

Of the samples collected in 1998, 122 were selected for geochemical analysis, and 130 polished thin sections were made. All of the samples were crushed and pulverized during a three week period spent in Ottawa. Wilson iron titrations are underway to determine Fe^{2+} and are expected to be finished by mid-October. The analysis of volatile species (H_{2}O, CO_{2}, and S) is complete. ICP-MS analysis for trace and rare earth elements are expected by November. A week in July was spent in Sudbury to conduct electron microprobe analysis of the following minerals: carbonates, sericite, feldspar, pyrite, chloropyrite and arsenopyrite. Mineral chemical data will be used in conjunction with whole rock data to assess the relative contribution of primary and alteration minerals to the chemical signature of the rocks. Interpretation of the geochemical data is underway and preliminary results are presented and discussed below.

Initial isocron diagrams presented in Figure 7.2 compare samples collected from the East Lightning Zone to those collected from the West Lightning Zone. A “least altered” sample is plotted on the x-axis providing a baseline to compare against altered samples (plotted on the y-axis). An isocron is a line that represents immobile elements; elements that plot above and below the isocron represent elemental gains and losses respectively (Grant 1986).

Figures 7.2a and 7.2b illustrate the gains and losses of major elements for sericitized samples from the West Lightning Zone and East Lightning Zone respectively. Figures 7.2c and 7.2d are comparable plots for albitized samples. Isocron plots for sericite alteration in the East Lightning Zone and West Lightning Zone are consistent in that they show an expected gain of K_{2}O. Similarly, albite altered samples from the East Lightning Zone and West Lightning Zone show a gain in Na_{2}O and a loss in CaO likely due to albization of plagioclase.

Macroscopic field observations show that the East Lightning Zone is less altered than the West Lightning Zone. In turn these observations are reflected on the isocron plots. For instance, Na_{2}O, CaO and MgO are more widely dispersed (see Figure 7.2c versus Figure 7.2d). The East Lightning Zone patterns differ from the West Lightning Zone in two ways. We expected to see a gain in CaO for sericite altered samples for the East Lightning Zone as in the West Lightning Zone due to similar abundances of carbonate minerals, but instead there is a loss. Secondly, there is a major loss of total iron in the albite-altered sample from the East Lightning Zone compared to the West Lightning Zone.

Future work will include plotting selected trace elements on the isocron diagrams in order to identify which elements are mobile and immobile. The isocron technique combined with mineral chemistry will provide information about the chemistry of the mineralizing fluids.

Figure 7.3 shows typical rare earth element (REE) plots for samples from the West Lightning Zone. Several distinct populations of samples can be observed. At the Holloway Mine, least altered and carbonatized samples with flat REE profiles having concentrations approximately equal to 10 times chondrite are common. Albite and variolitic samples have flat patterns with chondrite-normalized REE concentrations around 100 and small negative Eu anomalies. Sericitized samples usually lie in the middle with flat profiles and chondrite-normalized concentrations near 50. Similar flat lying patterns with REE from 10 to 100 times chondrite, and small negative Eu anomalies in their evolved rocks, are typical for basalt to rhyolite of theoleitic suites. The similarity of Holloway’s REE patterns to other tholeiitic suites of the area (e.g., Fowler and Jensen 1988) raises an important question. Are these patterns the reflection of an original suite of tholeiitic rocks or are they due to hydrothermal alteration? These hypotheses will be tested as part of Ropchan’s thesis.

In contrast to the volcanic rocks, sedimentary and coarse-grained heterolithic breccia units of Holloway’s West Lightning Zone have steep slopes (see Figure 7.3) with LREE equal to 200 to 300 times chondrite and HREE equal to 2 to 3 times chondrite. The LREE enriched patterns are characteristic for a calc-alkaline suite of rocks.
It has always been assumed that volcanic rocks in the Holloway Mine originally had similar chemical signatures. Recent observations within the mine lend credence to the idea, based on REE data, that Holloway mine stratigraphy may represent a basalt to rhyolite suite of tholeitic rocks. Some relatively undeformed blocks show what appears to be primary flow banded rock with spherulites aligned along the banding. Cores of this material are to be sampled for petrography. If this proves true then least altered equivalents for the isocon plots will have to be carefully chosen to represent the original chemistry of the rock.

Results from the electron microprobe work completed during the 1998 field season show the carbonate minerals are predominantly ankerite, although some calcite is present in the one “least altered” sample analyzed. The carbonate minerals within an individual sample can contain a range of Fe/Mg ratios. Only three samples contained sericite crystals large enough to analyze with the electron microprobe. The data confirms the micaceous mineral is indeed sericite (muscovite). All of the feldspars analyzed plotted in the albite field. During the electron microprobe work it was noticed that quartz was quite abundant (greater than 50%) in samples from the albite

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**Figure 7.2.** (a) Isocon plot for West Lightning Zone sericite alteration. The isocon is plotted at Y=1. CaO and K₂O show a gain due to the abundance of carbonate and sericite minerals in the sample. (b) Isocon plot for East Lightning Zone sericite alteration. The isocon is plotted at Y=1. K₂O shows a gain due to the abundance of sericite in the sample. (c) Isocon plot for West Lightning Zone albite alteration. The isocon is plotted at Y=1. Na₂O shows a gain while CaO shows a loss due to albition of plagioclase. (d) Isocon plot for East Lightning Zone albite alteration. The isocon is plotted at Y=1. Na₂O shows a gain while CaO shows a loss due to albition of plagioclase. Also observe the unexplained loss of Fe₂O₃.
alteration zone. Chlorite is predominantly iron-rich. Distinct populations of chlorite within each section are found on plots of FeO versus MgO. The significance of this observation is not yet determined. Electron microprobe analysis of pyrite included arsenic in the determination. High arsenic concentrations (approximately 2%) occur within the fine-grained pyrite of stringers, whereas coarse-grained pyrite has lower arsenic concentrations (0.0 to 0.2%). Arsenic concentrations at the Holloway Mine are significantly high especially since there is virtually no arsenic in pyrite at the nearby Holt-McDermott Mine. The importance of arsenic with respect to gold mineralization will be explored in Ropchan’s thesis. Electron microprobe data will be acquired for samples collected in 1998 in order to better define the alteration trends.

FRANCO—NEVADA AND GREATER LENORA JOINT VENTURE PROPERTY

This part of the project involves the collaboration of the OGS, Franco—Nevada and, the Earth Science departments of the Universities of Waterloo and Ottawa. Mattress is completing a Bachelor’s thesis supervised by B. Linen (University of Waterloo), A. Fowler (University of Ottawa) and J. Ayer (OGS).

The Franco—Nevada and Greater Lenora Holloway Joint Venture is located south of Highway 101 approximately 15 km west of the Quebec border. The Joint Venture property is bound to the north by Barrick Gold Corporation’s Holt—McDermott Mine. Talbot (1998), based on geology of the Barrick property, describes mineralization within strongly deformed and brecciated zones formed by regional faulting. Ore in the Holt—McDermott mine is closely associated with the Ghostmount and McKenna Faults. Both faults are splays of the Porcupine—Destor deformation zone (see Figure 7.1).

The study is based on drill core supplied by Franco—Nevada. During the 1998 field season approximately 20 samples were collected from the drill core. From those samples thin sections have been made and samples are being analyzed for major, trace, rare earth elements and volatile species. Initial observations include the recognition of hematite and albite alteration as well as two distinguish forms of pyrite, euhedral and massive (anhdedral). Albite alteration at the Holt McDermott mine hosts very fine-grained disseminated pyrite that is directly associated with gold (D. Talbot personal communication, 1998).

This study is focussed on the alteration and mineralization assemblages of the southern, down-dip extension, of Holt—McDermott’s ore, the South and West Vertical Zones (see Figure 7.1). Several alteration types increasing in intensity towards the mineralized centre enclose these zones. The main ore zones are within deformed bands altered by hematite, albite, chlorite, and quartz. Hematite is of economic importance as it is most closely associated with gold mineralization.

REFERENCES


8. Project Unit 96–003. Reappraisal of the Geology of the Shining Tree Area (East Part), Districts of Sudbury and Timiskaming

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INTRODUCTION

Active mineral exploration in the Shining Tree area and advances in concepts of Archean geology prompted the initiation of a multiyear study to re-evaluate the Archean geology and mineral potential of this mineral-rich portion of the southern Abitibi greenstone belt. The area is bounded by 47°29'30" to 47°45'30" north latitude and 80°55'00" to 81°27'30" west longitude comprising the townships of Miramichi, Asquith, Fawcett, Leonard, Tyrrell, Macmurchy, Churchill, Connaught, Cabot, Kelvin, Natal and Knight. The area was mapped for the Ontario Geological Survey in the early 1970s by M.W. Carter, who produced several detailed geological maps and reports culminating in a 1:50 000 compilation map and report (Carter 1987).

Six weeks were spent in 1996 examining cross sections through the Archean rocks (Johns 1996). Based on these observations, several changes were made to the interpretations portrayed by Carter (1987). The Archean rocks were tentatively subdivided into 2 packages (Johns 1996, Figure 8.1): 1) a lower, Keeewatin-aged, package (package 1) consisting of mafic to intermediate and felsic to intermediate flows and clastic rocks with a subpackage containing komatiitic flows, and 2) an intermediate pyroclastic to epiclastic package (package 2) of unknown age disconformably overlying package 1. Both packages were intruded by gabbros, possibly of early Proterozoic age.

Four weeks were spent in 1997 examining the rocks in Tyrrell Township and in parts of Knight, Natal and Macmurchy townships. The rocks were further subdivided and contacts were interpreted. The designation of packages 1 and 2 was retained until further data was collected. In addition, rocks interpreted to be Timiskaming-type sediments unconformably overlying package 1 metavolcanic rocks were outlined in southern Tyrrell Township (Johns 1997).

The eastern 6 townships of the area (Fawcett, Leonard, Tyrrell, Macmurchy, Natal and Knight) were examined during the 1998 field season. The results of 3 geochronological samples have allowed for a more complete reinterpretation of the area. In addition to the examination of rock types, more than 400 samples were collected to provide a sufficient database to interpret the geochemical evolution of the volcanic succession.

Initial results are reported in this volume (Oliver et al., this volume).

GEOCHRONOLOGY

Four samples collected in the map area were submitted to the Jack Satterly Laboratory, Royal Ontario Museum for geochronological analysis. Concordia diagrams are presented on Figure 8.1. Three samples are from the area outlined within this report; southwest Tyrrell Township, southeast Natal Township and northwest Macmurchy Township (Figure 8.2). A fourth sample is from central Kelvin Township, west of Natal Township.

Northwest Macmurchy Township
(Sample 97-JAA106)

Sample 97-JAA106 is a fine-grained calc-alkaline rhyolite (see Figure 8.2) from a unit which underlies a chemical metasediment of regional extent. Zircon grains in this rhyolite sample vary in appearance and also in 207Pb/206Pb age. Three grains of different appearance, 2 concordant and 1 slightly reversely discordant, have similar 207Pb/206Pb ages of 2725±6 Ma to 2727±4 Ma, the weighted mean of which, 2726±5 ±1 Ma, is considered the best estimate for the timing of crystallization. One grain has a younger 207Pb/206Pb age of 2716±9 Ma, but this grain is 3.8% discordant, and the younger 207Pb/206Pb age may be the consequence of ancient Pb loss, superimposed by recent Pb loss common in zircon. Two additional pale brown, euhedral, long prismatic zircon grains have 207Pb/206Pb ages of around 2741 to 2744 Ma, and are likely to be xenocrysts.

Southwest Tyrrell Township
(Sample 97-JAA105)

Sample 97-JAA105 is a quartz-rich feldspathic arenite from a sedimentary package that unconformably overlies Keewatin-type volcanic rocks (see Figure 8.2). Six detrital zircon grains of different appearance were analyzed from this rock. The 207Pb/206Pb ages fall into 2 distinct groups (3 grains each, see Figure 8.1) with weighted means of 2740±1 ±1 Ma and 2701±6 ±1.4 Ma, respectively. The zircons show no variation of apparent 207Pb/206Pb ages within groups. This pattern is consistent with only 2 main
Figure 8.1. Concordia diagrams with zircon data for rocks in the Shining Tree area.
Figure 8.2. Geology of the Shining Tree area.
sources having contributed to this sedimentary rock. The age of the weighted mean of detrital zircons of the younger group, 2701 ± 1.4 Ma, provides the maximum estimate for timing of sedimentation.

This immature sedimentary package has been termed the Indian Lake group.

**Southeast Natal Township (Sample 96-JAA0097)**

Sample 96-JAA0097 is from a coarse, monolithic, poorly sorted, debris flow comprising feldspar-phryic clasts (see Figure 8.2). Zircon grains in this sample are subhedral and vary in colour from colourless to brown. Three analyzed grains cover the range of colours, and show small degrees of discordance between 0.3 and 1.0%. The weighted mean of $^{207}\text{Pb}/^{206}\text{Pb}$ ages of all 3 grains is 2687.1 ± 1.0 Ma, which corresponds to the crystallization age.

**Central Kelvin Township (Sample 97-JAA107)**

Sample 97-JAA107 is a thin feldspar-rich arenite lying approximately 10 m above an unconformable contact between mafic metavolcanic rocks and the volcano-sedimentary package previously defined as package 2 (Johns 1996, 1997). The age distribution between 6 detrital zircon grains from this rock is similar to that in 97-JAA105. Three dark brown, euhedral, short prismatic zircon grains fall into the younger group with $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2684.4 to 2688.9 Ma; 3 grains with various appearances form the older group with the weighted mean of 2714 ± 1.1 Ma. This distribution implies involvement of 2 or possibly more sources of different ages. The analysis of the youngest $^{207}\text{Pb}/^{206}\text{Pb}$ age is 0.8% discordant, and this may represent either real differences in crystallization ages or ancient Pb loss. Thus the age of 2687 ± 3 Ma covering the range of $^{207}\text{Pb}/^{206}\text{Pb}$ ages in the young group, is considered the reasonable maximum estimate for the age of sedimentation.

Corfu, Jackson and Sutcliffe (1991), in a study of the Timiskaming Group in the southern Abitibi belt, noted an age range of 2677 to 2686 for the group. The 2 age determinations (96-JAA0097 and 97-JAA107) indicate that these rocks also belong within the Timiskaming Group.

**GENERAL GEOLOGY**

The map area is underlain by Neoarchean rocks that can be subdivided into 3 packages: an ultramafic to felsic metavolcanic sequence dated at 2726 Ma in northwest Macmurchy Township unconformably overlying an immature quartz-rich arenite to conglomerate package with a maximum age of deposition of 2701 Ma (herein called the Indian Lake group (ILG)) and an unconformably overlying Timiskaming metavolcanic to metasedimentary package dated at 2687 Ma. The youngest rocks in the area are Proterozoic Huronian Supergroup Gowganda and Lorrain formations intruded by Nipissing Gabbro and will not be described here. See Figure 8.2.

The older Keewatin-age rocks comprise an inter-banded sequence of ultramafic to mafic to felsic flows and volcaniclastic rocks, epiclastic and chemical sedimentary rocks. Basaltic to peridotitic komatitites have been highly altered and display a weak magnetic susceptibility. Mafic volcanic rocks are tholeiitic basalts to calc-alkaline andesites. Intermediate to felsic volcanic rocks are fine-grained calc-alkaline rhyolite to dacite flows and autoclastic and pyroclastic breccias. Rocks have undergone low degrees of metamorphism and have been partially altered to carbonated and sericitic. Chemical sedimentary rocks, which represent stratigraphic markers, comprise gray to black chert with minor Jasper. They are iron poor, containing pyrite with minor magnetite. These rocks are not discernable on aeromagnetic maps.

Immature arenite to conglomerate of the Indian Lake group (ILG) unconformably overlies the Keewatin volcanic rocks. The ILG comprises immature, coarse, quartz-rich, lithic arenite with layers and pods of angular to round clasts of jasper, gray-black chert, intermediate to felsic volcanic and granitoid rocks. Bedding planes are indistinct and there is evidence for scouring and erosion of bedding planes. Felsic to intermediate volcanic rocks are interbedded with these sedimentary rocks.

Timiskaming-type calc-alkaline, volcaniclastic rocks and alkalic mafic flows unconformably overlie the mixed metavolcanic rocks in the northern part of the area. This unconformity has been observed in central Kelvin Township (Johns 1996). The volcaniclastic rocks comprise plagioclase ± amphibole-phenocryst-bearing volcaniclastic to epiclastic rocks. The unit varies from pyroclastic breccia to laminated tuf to arenite.

Red to pink, hornblende-bearing fine-grained to medium-grained sills, dikes and sheets of shoshonitic composition intrude both the Keewatin- and Timiskaming-age supracrustal rocks. In Natal and Knight townships, similar rocks appear to be extrusive. These units are found in Knight, Macmurchy, Tyrrell and Natal townships. Timiskaming volcaniclastic rocks contain clasts of similar rock types.

**KEEWATIN METAVOLCANIC ROCKS**

The ultramafic komatitite flows occur predominantly in the upper part of the volcanic sequence (see Figure 8.2). They are interbedded with lesser mafic and intermediate to felsic metavolcanic rocks. The komatitites are altered to an apple green colour on fresh surface and have a dark chocolate brown weathered surface. Some of these units in the Houston Lake area have been previously interpreted as felsic volcanic rocks (Carter 1977a). Textures normally associated with ultramafic flows (cf. Pyke, Naldrett and Eckstrand 1973) are not common, but are preserved in less altered parts of the flows.

Tholeiitic basalt to calc-alkaline andesite (see Figure 8.2) is found as massive flows, pillow flows with minor
autoclastic breccia, and pillow breccia. Weathered surface colour varies from brown to light buff depending on the degree of alteration. Fresh surface colour varies between grey, grey-green, green and light green. The lighter-coloured mafic flows have been highly saussuritized and may be mistaken for a more felsic rock.

Calc-alkaline dacite to rhyolite units are interbedded with the mafic metavolcanic rocks (see Figure 8.2). They comprise fine-grained flows and/or dome complexes, primary pyroclastic and autoclastic rocks, and resedimented clastic rocks. Weathered surfaces are generally light in colour and the fresh surfaces are light grey to grey. Minor, highly chloritized units are dark green on the fresh surface. The more felsic rocks are generally hard and brittle.

A thin, chemical-sedimentary marker horizon can be traced across the area and comprises thinly laminated, contorted, brecciated, black to gray chert and red chert. These iron-poor sedimentary rocks do not have a magnetic signature, which makes it difficult to trace their true extent. The black and red cherts may have been the source of the chert clasts found in the Indian Lake group described below.

Ultramafic to mafic intrusive rocks are common within the map area. The generally low degree of metamorphism and lack of penetrative deformation make the subdivision between Proterozoic and Archean intrusions difficult. Within Fawcett Township, the peridotitic to gabbroic intrusive rocks are poorly exposed and have been interpreted to be Archean in age as they appear sill-like within the Keewatin metavolcanic rocks (see Figure 8.2). Here they are related to thin peridotite to dunite horizons found along their margins. Carter (1977b) has interpreted the thin sill to be a komatiitic flow horizon. As these rocks appear to occur within a thick, felsic metavolcanic package, they are reinterpreted to be layered peridotite to gabbro sills.

Intermediate to felsic porphyry dikes, sheets and irregular-shaped bodies are found throughout the area, but are concentrated in northern Tyrrell Township (Figure 7.2 in Johns 1997). The porphyries comprise hornblende-feldspar and hornblende porphyry, feldspar, quartz-feldspar and quartz porphyry. Clasts of these porphyry types are commonly found as clasts in the Timiskaming Group. As the intrusive equivalents to these porphyry types have not been noted intruding the Timiskaming rocks, they are inferred to be Keewatin in age.

The area around Porphyry Lake has been flooded with feldspar, hornblende-feldspar and hornblende porphyry. In many outcrops these are the predominant lithology. Crosscutting relations suggest that the feldspar porphyry is the oldest. The hornblende-bearing porphyries may be coeval with the hornblende monzonite to monzodiorite Milly Creek Stock.

INDIAN LAKE GROUP

Indian Lake group is a predominantly sedimentary package of rocks with interbedded intermediate to felsic volcaniclastic rocks that unconformably overlie the Keewatin volcanic rocks. The largest exposed of these rocks is found in southern Tyrrell Township and northern Leonard Township. Smaller isolated outliers are found along the Hydro Creek Fault in western Tyrrell Township, along the south shore of Houston Lake and along the shore of Bigfour Lake (both in northern Macmurchy Township). See Figure 8.2.

These rocks are, for the most part, steeply dipping, but the exposed unconformities in Tyrrell Township are flat lying, with the surface being a fractured regolith filled with coarse sand. The contact can be traced vertically through blocks of regolith decreasing in both size and abundance into the overlying sedimentary rock that is composed of medium- to coarse-grained, immature, quartz-feldspar to feldspar-quartz arenite with lenses of conglomerate.

Clasts comprise red and black chert, vein quartz, felsic volcanic rock, granitoids and quartz-rich granitoids, all of which are subrounded to angular. Conglomerates occur as thin lag beds and thicker lenses and beds that vary from clast to matrix supported. From the immature nature of the deposit and the lack of discrete bedding planes, it is surmised that they were deposited by mass-flow mechanisms that eroded underlying beds.

Calc-alkaline, intermediate to felsic pyroclastic rocks that are interbedded with the sediments are angular, monolithic mass flows ranging in size and sorting from pyroclastic breccia to lapilli tuff. They are massive, with no internal structure and occasionally contain minor black chert fragments. Some of the units have been interpreted to be dome and dome-facies clastic rocks (J. Walker, Shiningtree Resources, personal communication, 1997).

TIMISKAMING GROUP

Alkaline Mafic Metavolcanic Rocks

Alkaline mafic metavolcanic rocks have a limited areal extent in southeast Natal, northeast Macmurchy and north-central Tyrrell townships (see Figure 8.2). The unit is a fresh, hornblende-pyroxene-phyric, fine-grained, crystalline, xenolith-bearing shoshonitic mafic rock. The xenoliths vary greatly in size and shape, some having irregular bomb-like shapes and others subangular block shapes. The unit is homogeneous without sufficient distinguishing features to determine a deposition mechanism.

Calc-alkaline Volcaniclastic to Epiclastic Rocks

Calc-alkaline volcaniclastic to epiclastic rocks occur predominantly in Natal Township (see Figure 8.2), where they unconformably overlie the Keewatin metavolcanic rocks.

This unconformity was observed in central Kelvin Township where the Keewatin mafic flows become bleached and highly silicified as the unconformity is approached and clasts of these bleached mafic rocks are found in the thinly bedded, siliceous siltstone base of the
Timiskaming sedimentary rocks (Johns 1996). Minor clasts of mafic volcanic rocks are found 10 m higher in a polymictic conglomerate.

These rocks are predominantly fragmental, ranging from primary pyroclastic to volcanioclastic epiclastic rocks which contain euhedral to subhedral phenocrysts of amphibole and/or pyroxene, and feldspar. Minor siliceous, chemical metasedimentary units also occur. The epiclastic rocks have been deposited primarily by debris-flow and turbidite mechanisms. Overall, these rocks fine from proximal in the east to distal in the west. More-distal units show evidence of subaqueous redeposition as debris flows and volcanioclastic arenites containing interbedded siliceous siltstones, tuffs and chert.

In southeast Natal Township, fine-grained intermediate crystalline rocks, possibly related to the alkalic to subalkalic intrusive rocks described below, are found and may represent extrusive equivalents to the clastic rocks.

Alkaline to Subalkaline Intrusive Rocks

Fine-grained to very fine-grained, brick red, hornblende syenitic sills and dikes are common, intruding both Keewatin and Timiskaming rocks. The largest of these bodies is an elongate intrusion in northwest Tyrrell Township. It has a coarser-grained monzonite core that is exposed in its southern tip, north of Hare Lake. In the Milly Lake area and in Natal Township, similar-appearing rocks exhibit extrusive-like features.

The Milly Creek Stock, exposed in south central Knight Township, comprises euhedral hornblende monzonite to monzodiorite. It is medium grained, inequigranular and contains xenolith trains of the surrounding country rock. This stock has a similar appearance to some of the hornblende and feldspar-hornblende porphyries intruding the Keewatin rocks and to many of the clasts found within the Timiskaming clastic rocks. The stock may be a high-level intrusion that provided a source for some of the volcanic rocks within the Timiskaming Group.

METAMORPHISM AND ALTERATION

The degree of preservation of both macro- and microscopic features and textures indicates that the rocks in the map area have undergone a low degree of metamorphism and selected areas have not been penetratively deformed. All the rocks have undergone moderate to extensive degrees of carbonatization, sericitization and kaolinitization. The felsic metavolcanic rocks in the region around Bigfour Lake have undergone extensive chloritization and silicification.

STRATIGRAPHY

Three ages of rocks have been documented within the map area. See previous discussion in “Geochronology”. An older, Keewatin, mixed metavolcanic sequence unconformably overlain by a younger, immature metasedimentary package (ILG) is unconformably capped by a mixed volcanic-epiclastic sequence of Timiskaming age. Both the Indian Lake group and the Timiskaming-age rocks can be defined as assemblages (cf. Thurston 1991), while the underlying volcanic rocks await further work.

The Keewatin metavolcanic rocks are a mixed felsic to ultramafic sequence estimated to be 14 km thick. There is not sufficient evidence to account for this thickness by folding, but thrust-fault repetition can not be ruled out. If the chemical sedimentary rocks found within the central part of the volcanic sequence are a single stratigraphic horizon, it may divide the volcanic rocks into upper and lower packages. A detailed geochemical study being carried out by Oliver et al. (this volume) may assist in determining either scenario.

The thickness of the Indian Lake group in Tyrrell and Leonard townships is difficult to determine as the dip of the rocks vary from flat to steeply overturned. Minor opposing tops indicates the presence of some folding. Other ILG outcrop locations are of limited areal extent and may represent erosional remnants of a much more extensive sedimentary package. The source of these quartz-rich sediments may have been volcano-plutonic highlands to the east and southeast currently covered by the Huronian Supergroup, and may represent the remnants of an alluvial fan.

The Timiskaming Group rocks in Natal Township have variable strikes and dips ranging from shallow to overturned (Carter 1983). Isolated outcrops of Keewatin-type volcanic rocks are found within the Timiskaming clastic rocks. These may be inliers within a folded Timiskaming sequence of limited depth extent, which generally fines to the west and north and may represent a prograding, thickening fan with a source in the southeastern Natal Township area.

VOLCANOLOGY

Throughout the area the Keewatin volcanic rocks were deposited in a deep, subaqueous environment, as evidenced by the copious pillows and lack of abundant amygdules in the mafic flows. The lack of abundant interflow sediments in the mafic flows indicates rapid accumulation. Felsic to intermediate volcanic rocks are bounded by pillowd mafic flows and are also presumed to be subaqueous.

The felsic to intermediate volcanic rocks range in depositional mode from flows with autoclastic breccia and flow banding to blocky and ash flows and resedimented debris flows. Size and sorting cover the range from pyroclastic breccia to tuff. With further work the location of vent areas might be surmised.

STRUCTURAL GEOLOGY

In Fawcett and Macmurchy townships the Keewatin metavolcanic rocks strike to the north-northwest in the south changing to westerly strike in northern Macmurchy
Township. Except for minor top reversals, the sequence youngs to the northeast and north. In Leonard, Tyrrell and Knight townships the volcanic rocks strike north-northwest changing to a northerly strike in Knight Township. The sequence youngs to the northeast towards a north-northwest-trending syncline axis in northern Tyrrell Township.

The Timiskaming Group has variable strikes and facing directions. Insufficient data has been collected to identify the cause of the folded aeromagnetic pattern in the area underlain by the Timiskaming Group as portrayed on the recently re-released Ontario Geological Survey aeromagnetic maps (Gupta 1996). It is possible that this aeromagnetic pattern reflects the geology of the underlying Keeawan stratigraphy. In the southeast part of Natal Township, these rocks strike easterly, abutting against the Keeawan volcanic rocks and, for the most part, young to the north.

The entire map area has been disrupted by northwest- and northeast-trending faults, thus making the interpretation of stratigraphic relations difficult. This has also been noted on a property scale in northern Tyrrell Township (J. Walker, Shiningtree Resources, personal communication, 1997). Deformation is restricted to these fault zones and the individual blocks exhibit little evidence of internal deformation.

The Hydro Creek Fault or inferred splay from it in Tyrrell and Natal townships are related to gold mineralization and intense alteration. In some locations there appears to be intense carbonatization, but little ductile deformation is evident. Ductile deformation appears to be more intense on the splay faults. Intense east-northeast-directed shearing and foliation, along with strong carbonatization, affects the rocks in southwest Macmurchy Township north of Gay Lake.

A recently completed lake-sediment and water geochemical survey (Hamilton 1997) included all of Tyrrell, Knight and Leonard townships and 29 additional townships to the north, south and east. A cluster of anomalous gold samples is located in, and to the west of, Spider Lake in northwest Leonard Township. Anomalous elements include gold ± silver, arsenic, copper, lead and zinc. With the exception of the high gold values and lack of nickel, the signature has similarities to that of the Gowganda area. Nipissing Gabbro underlies most of the anomaly, but a significant proportion of the catchment area includes Keeawan volcanic rocks and Indian Lake group. Hamilton (1997) reports, "It is possible that this pattern of anomalous samples represents the signature of more than one type of mineralization".

The felsic to intermediate metavolcanic rocks in Macmurchy Township beneath the chemical metasediment horizon have been strongly chloritized and silicified, and may be prospective targets for base metals (D. Melling, Arcana Consulting Inc., personal communication, 1998). The felsic to intermediate metavolcanic rocks close to chemical metasedimentary rocks throughout the area should be re-examined for their base metal potential.

Disseminated to locally massive sulphide minerals, including sphalerite, chalcopyrite and galena, occur in felsic rocks intruded by an ultramafic to mafic intrusive body hosting 3 000 000 t of ore, grading 1.4% Ni and 0.47% Cu in northwest Fawcett Township (Fort Knox Gold Resources Inc. Press Release, Business Wire, May 29, 1998). If the 2 gabbro-peridotite sills within the felsic metavolcanic rocks to the southeast are related to the aforementioned mineralized body, then they should be examined for nickel and PGE potential. As well, the felsic volcanic rocks should be examined for base metals.

REFERENCES


9. Project Unit 98–15. Preliminary Geochemistry of Metavolcanic Rocks of the Shining Tree Area; Abitibi Subprovince, Ontario

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INTRODUCTION

The Shining Tree area is part of the southern Abitibi subprovince and is situated 120 km north of Sudbury. The area is roughly 40 by 28 km and comprises 12 townships: Cabot, Kelvin, Natal, Knight, Connaught, Churchill, Macmurchy, Tyrell, Miramichi, Asquith, Fawcett and Leonard (Figure 9.1). There is currently considerable exploration interest particularly in gold and to a lesser degree in volcanogenic massive sulphide (VMS) deposits. This paper contributes to the understanding of the potential for economic gold and VMS deposits in the area by presenting preliminary geochemical data characterising the trace element geochemistry of the metavolcanic rocks. The trace element geochemistry of volcanic rocks has been shown to be a reasonable exploration filter, particularly for VMS deposits (Lesher et al. 1986; Kerrich and Wyman 1996). As well, geochemical data can identify alkaline volcanic rocks associated with Late Archaean pull apart basins which have elevated potential for lode gold deposits (Thurston and Chivers 1990 and references therein). This study will help to pinpoint these late sequences and their associated gold potential (Colvine et al. 1988).

The area was first mapped in detail by Carter (1987). The OGS initiated a re-evaluation of the area because of increased exploration interest in the 1990s. In the summer of 1996, G. Johns began a project to compile the geology of the area at a scale of 1:50 000. The summer of 1998 has seen reconnaissance mapping of the 6 eastern-most townships of the area (see Figure 9.1) and in the summer of 1999 the 6 western-most townships will be mapped.

Preliminary analytical data for nearly 100 samples of metavolcanic rocks collected in autumn 1997 by H. S. Oliver and 50 collected by G. W. Johns in 1996 and 1997 form the basis of this report. A further 400 samples were collected by H. S. Oliver during the 1998 field-season. The analyses so far indicate at least three main groups of rocks that are not geochemically related to each other (Figure 9.2). The possible petrogenetic relationships (or lack of them) between these groups are discussed towards the end of the text.

GEOLOGICAL SETTING AND FIELD DESCRIPTIONS

In the Abitibi subprovince, the mafic and associated felsic and ultramafic metavolcanic rocks have a Keewatin age of 2720 to 2700 Ma (Jackson, Fyon and Corfu 1994). The mafic flows are the main metavolcanic component of this greenstone belt. The mafic and ultramafic rocks of the Keewatin sequences in the Abitibi subprovince may be geochemically analogous to basaltic rocks of modern mid-ocean ridge and back-arc settings (Jackson, Fyon and Corfu 1994). The Timiskaming style assemblages comprise a set of volcanic and contemporaneous plutonic alkalic rocks which are most likely to have been associated with late stage arc development and rifting (Jackson, Fyon and Corfu 1994). The sediments of the Timiskaming assemblage have a depositional age of 2685 ± 3 Ma (Corfu, Jackson and Sutcliffe 1991), the alkalic metavolcanic rocks associated with them are approximately 2686 to 2677 Ma (Corfu, Jackson and Sutcliffe 1991).

Johns (1996) divided the metavolcanic rocks of the Shining Tree area into two packages on the basis of their lithostratigraphic relationships. The first constitutes a metavolcanic sequence of mafic to felsic flows, with local komatiitic flows and minor interflool drip water sediments. Geochronological dating (see Johns and Amelin this volume) of rhyolites from Bigfour Lake (Macmurchy township) gives an age of 2726 ± 1 Ma (Keewatin). The second group comprises metasedimentary rocks and associated metavolcanic rocks which have been dated at 2687 ± 3 Ma which indicates that these rocks are Timiskaming in age. The metavolcanic rocks in this package are calc-alkalic and alkalic flows (the later originally had hornblende phenocrysts) and syenite stocks. (Johns 1997, Johns and Amelin this volume). The Timiskaming Assemblage in the Kirkland Lake area to the northeast of the Shining Tree area is predominantly metasedimentary with associated pyroclastic rocks. The metasedimentary rocks at Kirkland Lake have been subdivided into a non-marine association and a resemented association (Hyde 1980). The non-marine association is dominated by pebble-cobble conglomerates and sandstones. Hyde (1980) has interpreted these as representing braided river, floodplain,
lacustrine and aeolian deposits. The resedimented association includes graded and ungraded conglomerates, sandy, silty and laminated turbidites and iron formation (Hyde 1980). The resedimented sequence has been interpreted by Hyde (1980) as representing deposition within a submarine fan. The volcanic rocks are mainly alkalic basalts and trachytes with contemporaneous syenite and augite-syenite intrusions (Cooke and Moorhouse 1969).

A newly defined sedimentary sequence, named the Indian Lake group is recognised in the Shining Tree area (Johns and Amelin this volume). This group is found mainly to the south and east of Indian Lake in Tyrrell Township (see Figure 9.1). It has a U/Pb zircon age based on detrital zircons from metasedimentary rocks of 2701.6 ± 1.4 Ma (Johns and Amelin this volume). Although Keewatin in age, this unit unconformably overlies the Keewatin metavolcanic rocks.

The rocks of the Shining Tree area are thus placed in a subprovincial scale setting. The following lithological descriptions emphasize significant types of alteration likely to perturb geochemical relationships. For a detailed

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**Figure 9.1.** Geological map of the Shining Tree area (Modified from Johns 1997).
description of the lithologies and composite packages, see Johns (1996; 1997) and Johns and Amelin (this volume).

The Keewatin Assemblage

This group is mainly a volcanic sequence that is Keewatin in age (Johns 1996). The sequence youngs to the northeast (see Figure 9.1). It is composed of mafic metavolcanic flows, with interbedded intermediate to felsic flows and pyroclastic. The komatites associated with this package are found in clusters in several places within the mafic metavolcanic rocks.

The mafic and some of the intermediate rocks comprise pillowd and massive flows. In most cases the pillows are well preserved and show typical fine grained margins, with little inter-pillow material preserved. Alteration recognizable in the field includes late pervasive vein carbonate and silification. This has locally imparted a pale grey colour to the metavolcanic rocks. The felsic rocks include aphanitic, aphyric and quartz-phyric flows and associated pyroclastic rocks. The pyroclastic rocks range from tuff to tuff breccia, the most common lithology being lapilli tuff. Alteration includes late sericitization possibly representative of potassium metasomatism and silification. The komatitic flows demonstrate macro- and micro-spinifex textures, polysuturing and brecciation. Thin dunitic cumulate horizons can also be seen. These komatitic flows tend to be black unless silicified or otherwise altered, in which case they can weather to a pale grey colour (Johns 1996; 1997; Johns and Amelin this volume). The komatites are commonly carbonitized. In a few locations close to Shining Tree, they are altered and contain abundant fuchsite. This mineral gives the rocks a strong green colour and this alteration is also associated with abundant silicification.

The Indian Lake Group

The Indian Lake group overwhelmingly comprises a series of metasedimentary rocks including quartzo-feldspathic arenite and wacke and coarse conglomerate and rare mudstone. With an age of 2701.6 ± 1.4 Ma, the Indian Lake group is similar in age to the Porcupine assemblage near Timmins (2689 ± 4 Ma, Corfu et al. 1989) which is mainly composed of wackes. The Ojoolke assemblage (of the Porcupine assemblage) in Matheson, Evelyn, Hoyle and Gowan townships district of Cochrane, is dominated by fine- to medium-grained turbiditic wackes, composed predominantly of feldspar and quartz with minor mafic and felsic volcanic grains. These sediments are thought to represent medial and distal turbidites of a submarine fan formed by large-scale slope failure (Berger 1992; 1994; Pyke 1982) developed during initial deformation of the belt.

The Timiskaming Assemblage

The Timiskaming sediments form a wide strip trending easterly across the northern Connaught and Churchill townships, most of Natal township, and the eastern part of Kelvin township in the north (see Figure 9.1; Johns and Amelin this volume). An exposure of the unconformity has been found between the Timiskaming rocks and the underlying Keewatin sequences (Johns 1996). They were mainly deposited as debris flows and possibly as turbidites (Johns 1996). The volcaniclastic rocks and associated pyroclastic rocks are amphibole and feldspar phenocryst bearing tuff and more common lapilli tuff to tuff breccia. Alkaline volcanic rocks in Natal Township (Carter 1987) (see Figure 9.1) are amphibole-phyric pointing possibly to a shoshonitic bulk composition. Associated with the volcanic rocks are hornblende-phyric syenitic stocks.


**Metamorphism and Structure**

A discussion of the metamorphic and structural features of the area is given by Johns and Amelin (this volume). The present summary is intended to aid in the ensuing interpretation of geochemical relationships.

Adjacent to the felsic intrusions, such as the Miramichi and Togo batholiths, the metamorphic grade reaches amphibolite facies. However, elsewhere the grade of metamorphism is lower greenschist facies. Locally, the levels of metamorphism and deformation are low enough to preserve delicate features such as perlitic fractures in relict glass.

There are two main phases of deformation in this area. The Keevatian rocks predate deformation, with D1 after their eruption and deposition. At a regional scale, Keevatian assemblages are affected by D1 and Timiskaming assemblages are affected by D2. There are two deformation zones in which gold showings have been found one in Macmurchy and the other in Tyrrell Township (see Figure 9.1; Johns 1996, 1997 and Johns and Amelin, this volume) for a more extensive petrographic description of the above packages and geological setting.)

**GEOCHEMISTRY**

**Analytical Techniques**

The samples were analysed by X-ray fluorescence (XRF) spectrometry for major elements and some trace elements at the University of Portsmouth and at the Geoscience Laboratory at the Ministry of Northern Developments and Mines (MNDM). Pressed powder pellets were used to analyse trace elements and glass fusion discs prepared according to the method of Norish and Hutton (1969) for analysis of major elements. Trace elements including the rare earth elements (REE) were also analysed using Inductively Coupled Plasma-Mass Spectroscopy (ICP-MS) at the Geoscience Laboratory at the MNDM. When compared to standard reference materials (SRMs), samples are within acceptable accuracy and precision limits (Tomlinson 1996), with precision of less than 1 wt. % for major elements. Tomlinson (1996) analysed (SRMs) by ICP-MS. Within detection limits, elements found to be within 5% of the recommended values are Y, La, Pr, Nd, Dy, Ho, Er. Elements generally within 10% of the recommended values are Ce, Sm, Eu, Gd, Tb, Tm, Yb, Lu, Ta, W, Th, U. Elements generally outside this accuracy are Rh, Sr, Zr, Hf and Nb. Rb and Sr are considered mobile in greenschist belts and are not used. Zr is within 12% accuracy except for sample DR-N, likewise for Hf which is otherwise accurate to within 5%. Nb returns a value lower than that recommended for all of the SRMs.

**Sample Distribution**

The samples were collected by H. S. Oliver in 1997 and by G.W. Johns in 1996 and 1997. To date trace element analyses have only been performed on the samples collected by Johns, therefore this data is utilised in the following diagrams. In the following year, the samples from the initial sample sets, and those collected in 1998 will be analysed by ICP-MS to complement the XRF major element data. In 1997 the sampling was concentrated along Highway 560. A far more extensive spatial distribution of samples was achieved in 1998. The present coverage extends into all but Connaught, Miramichi and most of Cabot Townships in the far west of the area. Over 350 Keevatian metavolcanic rocks have now been sampled, over 70 metavolcanic rocks from the Timiskaming assemblage, and nearly 50 sedimentary rocks evenly sampled from both packages.

Until quite recently the basaltic rocks comprising greenstone belt sequences were treated as one unit (Condle 1976) leading to the concept of the “average Archaean basalt” which bore little resemblance to modern geodynamic environments. It is now realised that once greenstone belts are subdivided into lithotectonic assemblages the mafic volcanic rocks within individual assemblages can be successfully compared with modern geodynamic principally based on trace element distributions (Jackson, Fyon and Corfu 1994; Tomlinson et al. 1996). As all of the rocks in Archaean greenstone belts are metamorphosed and altered to varying degrees, only immobile elements can be used (cf. Tomlinson 1996). The HFSE (Th, Nb, Ta, Zr, Hf and Ti) are less mobile than the LREE (Jenner 1996). These together with the MREE and HREE, and Al and Y (are also thought to be relatively immobile in Precambrian tholeiites and komatiites) are best used for trace element systematics (Kerrich and Wyman 1996).

Examples of this approach of subdivide a greenstone belt into assemblages and then postulating modern geodynamic equivalents includes the work of Tomlinson et al. (1996) on the Beardmore–Geraldton belt. There, basaltic and andesitic rocks were compared to arc, back arc and oceanic geodynamic settings based on geochemical relations. Tomlinson, Hughes and Thurston (1995) discovered coherent subdivisions of komatiitic and tholeiitic units in the 3 Ga Lusby Lake greenstone belt clearly identifying a plume influence in the genesis of Mesoarchean mafic volcanism.

The samples have been segregated first by field evidence, using contact relationships where possible and obvious lithologic differences, for example, many of the Timiskaming volcanic rocks in the study area are amphibole phyr, but this is not always characteristic of rocks within the Timiskaming Assemblage elsewhere. The samples were then plotted on a Jenson plot (Figure 9.3a) and the other discrimination diagrams presented in Figure 9.3 and any obvious groups noted. The samples were then plotted on REE and extended multielement plots. Using these discrimination diagrams, negative (Zr/Sm)N and (Hf/Sm)N anomalies were noted in the Timiskaming volcanic rocks. These are absent in the Keevatian volcanic rocks. The Keevatian rocks are divided into tholeiitic and calc-alkaline units. The tholeiites are N-MORB and are spatially associated with komatiites. The calc-alkaline rocks mainly consist of felsic-dominated volcanic rocks (see Figure 9.3a). The Timiskaming volcanic rocks are
calc-alkaline and range from basaltic to felsic. The felsic volcanic rocks can be subdivided into a Th enriched group and a standard group.

**Geochemical Characteristics**

**KEEWATIN METAVOLCANIC ROCKS**

**Tholeiitic Basalts**

The tholeiitic basalts have unfractionated primitive mantle normalised (values from G. Jenner, Newpet) REE distributions (\((\text{La/Lu})_N = 1.02 \text{ to } 1.58\); \((\text{La/Sm})_N = 0.94 \text{ to } 1.28\); \((\text{Gd/Lu})_N = 0.93 \text{ to } 1.41\) (see Figure 9.2a) and rock/primary mantle normalised values ranging from 5 to 9. They are slightly depleted in the HFSE Th and Nb (\((\text{Nb/La})_N = 0.58 \text{ to } 0.91\); \((\text{Th/Nb})_N = 0.69 \text{ to } 1.13\)). They are also marginally depleted in Ti (\((\text{Ti/Sm})_N = 0.52 \text{ to } 0.78\) (see Figure 9.2a).

There are both basaltic and peridotite komatiites in the area (Figure 9.3a). Only one komatiitic sample as yet has been analysed for complete trace element and REE data. Thus there is little certainty that this is truly representative of the komatiites of the area. The komatiite is LREE depleted (\((\text{La/Sm})_N = 0.77\) and HREE enriched (\((\text{Gd/Lu})_N = 1.10\), \((\text{La/Lu})_N = 0.73\) (see Figure 9.2a). There is a positive Zr anomaly (\((\text{Zr/Sm})_N = 1.49\)) and a negative Hf anomaly (\((\text{Hf/Sm})_N = 0.67\)) relative to other HFSE.

**Calc-Alkaline Basalt**

There is only one available sample of a Keewatin calc-alkaline basalt. This may be an uncommon lithology within the Keewatin in the Shining Tree area. As such we clearly cannot present an average data array. This sample, like the komatiite, must be treated with caution as they could both be anomalous. Like all of the calc-alkaline volcanic rocks, this is a fractionated basalt, with moderately steep primitive mantle normalised REE distributions (\((\text{La/Lu})_N = 6.21\), \((\text{La/Sm})_N = 2.25\), \((\text{Gd/Lu})_N = 1.85\)) (Table 9.1). This basalt has negative Nb (\((\text{Nb/La})_N = 0.18 \text{ to } 0.48\)) and Ti anomalies (see Figure 9.2b).

![Figure 9.3](image-url)  
A series of basalt discrimination diagrams, note that felsic samples have been included as some of the plots. **a** This (\(\text{FeO}_T\), \(\text{TiO}_2\), \(\text{Al}_2\text{O}_3\), MgO) cation plot (Jenson 1976) shows the overall distribution of the all of the metavolcanic rocks; **b** Calc-alkaline mafic and intermediate volcanic rocks are enriched in Th and fall in the VAB field; **c** Tholeiite basalt plot in a tight MORB group; **d** A clear division is visible between the tholeiites and the calc-alkaline basalt; **e** The tholeiites, calc-alkaline basalt and felsic and intermediate rocks plot in separate groups; **f** again the magma suites plot independently of each other.
Calc-alkaline Intermediate to Felsic Metavolcanic Rocks

The few Keewatin aged intermediate and felsic volcanic rocks in the area are calc-alkaline. They have fractionated primitive mantle normalized REE distributions \((\text{La/Lu})_N = 7.43\) to 24.18, \((\text{La/Sm})_N = 2.95\) to 5.95, \((\text{Gd/Lu})_N = 1.73\) to 3.30\) with the HREE producing a slight trough shape (see Figure 9.2c). These volcanic rocks have HFSE depletion in the form of negative Nb and Ti anomalies \((\text{Ti/Gd})_N = 0.40\) to 0.55).

TIMISKAMING METAVOLCANIC ROCKS

The Timiskaming metavolcanic rocks include calc-alkaline mafic to felsic rocks. There are no tholeiites associated with this assemblage.

Calc-alkaline Basalts

The Timiskaming basalts have strongly fractionated primitive mantle normalized REE distributions \((\text{La/Lu})_N = 7.54\) to 11.43. They are LREE enriched \((\text{La/Sm})_N = 2.15\) to 2.58) with the HREE producing a very slight trough shape on a nearly flat pattern \((\text{Gd/Lu})_N = 2.28\) to 2.43). The mantle normalised multi-element plot indicates depletion in numerous HFSE (see Figure 9.2b), Nb \((\text{Nb/La})_N = 0.16\) to 0.25), Ti \((\text{Ti/Gd})_N = 0.56\) to 0.59) and Sc with respect to the mean gradient of the plot. Zr and Hf are also depleted \((\text{Zr/Sm})_N = 0.79\) to 0.93, \((\text{Hf/Sm})_N = 0.37\) to 0.66) but to a lesser extent. On the other hand, these rocks are enriched in Th \((\text{Th/Nb})_N = 2.89\) to 4.25). These rocks in this plot in the calc-alkaline arc field on the Zr-Th-Nb plot (Wood, Joron and Treuil 1979) (see Figure 9.3b).

Felsic and Intermediate Volcanic and Intrusive Rocks

These rocks have very similar REE and extended multi-element plots to the calc-alkaline basalts (see Figures 9.2b and 9.2c), the most significant difference being the greater abundance of REE and HFSE in the felsic volcanic rocks compared to the basalts. There are two groups of calc-alkaline felsic and intermediate rocks. A Thorium enriched group and a “standard” group (see Figure 9.3b). There is little difference between the two groups (see Table 9.1). The standard and Thorium-enriched felsic rocks do not plot independently of each other except on Figure 9.3b where the Thorium-enriched felsic rocks are drawn towards the Th variable. Both groups have strongly fractionated primitive mantle normalised REE distributions and negative Nb, Ti and Zr, Hf anomalies (see Figure 9.2c). The standard group has a slightly steeper slope \((\text{La/Lu})_N = 14.67\) to 22.96, \((\text{La/Sm})_N = 3.38\) to 4.26, \((\text{Gd/Lu})_N = 4.38\) to 5.43) on REE distributions than the Thorium enriched group \((\text{La/Lu})_N = 32.55\) to 44.75, \((\text{La/Sm})_N = 3.04\) to 3.68, \((\text{Gd/Lu})_N = 3.09\) to 3.52). The extended multi-element plot shows large negative Nb and Ti anomalies, and negative Zr and Hf anomalies (see Table 9.2 for details). The standard group has lower Th \((\text{Th/Nb})_N = 6.16\) to 12.05) than the Th-enriched group \((\text{Th/Nb})_N = 11.19\) to 18.28).

DISCUSSION

The tholeiitic package appears to have geochemical characteristics similar to those of modern normal mid ocean ridge basalts (N-MORB), as implied by the distribution on the basalt discrimination diagrams. The tholeiitic basalts plot in almost every case on the MORB field on basalt discrimination diagrams. However, the tholeiites do not have all of the parameters defining true modern N-MORB (Saunders, Norry and Tarney 1988). The tholeiitic basalts plot as high-iron and high-magnesium tholeiites on the Jenson cation plot (1976) (see Figure 9.3a). In terms of their Ti-Zr-Y ratios (Pearce and Cann 1973) (see Figure 9.3c) and Nb-Zr-Y characteristics (Meschede 1986) (see Figure 9.3d) the tholeiitic basalts plot as a tight cluster in the N-MORB field. In a plot of Th against V (Shervais 1982) (see Figure 9.3e), the tholeiites plot in the MORB, back arc basal fields. On a plot of Zr/TiO2 against Nb/Y (Winchester and Floyd 1977) (see Figure 9.3f), all of the rock types plot in well defined groups.

The characteristics of basalts formed in an arc environment are negative Nb, Ta and Ti anomalies (Pearce 1996). Although Ta has not been plotted in Figures 9.2b and 9.2c negative Nb and Ti anomalies can be seen in all of the calc-alkalic rocks. The anomalies are smaller in the Keewatin volcanic rocks than in the Timiskaming volcanic rocks. The calc-alkalic basalts have strongly fractionated

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**Table 9.1.** A summary of primitive mantle normalised REE and HFSE ratios.

<table>
<thead>
<tr>
<th></th>
<th>((\text{La/Lu})_N)</th>
<th>((\text{La/Sm})_N)</th>
<th>((\text{Gd/Lu})_N)</th>
<th>((\text{Zr/Sm})_N)</th>
<th>((\text{Hf/Sm})_N)</th>
<th>((\text{Ti/Gd})_N)</th>
<th>((\text{Nb/La})_N)</th>
<th>((\text{Th/Nb})_N)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Keewatin</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Komatiite</td>
<td>0.73</td>
<td>0.77</td>
<td>1.10</td>
<td>1.52</td>
<td>0.69</td>
<td>0.84</td>
<td>0.66</td>
<td>0.68</td>
</tr>
<tr>
<td>MORB</td>
<td>1.02-1.58</td>
<td>0.94-1.28</td>
<td>0.93-1.41</td>
<td>0.98-1.30</td>
<td>0.60-1.14</td>
<td>0.61-0.94</td>
<td>0.58-0.91</td>
<td>0.69-1.13</td>
</tr>
<tr>
<td>CAB</td>
<td>6.21</td>
<td>2.25</td>
<td>1.85</td>
<td>1.31</td>
<td>1.20</td>
<td>0.00</td>
<td>0.45</td>
<td>1.80</td>
</tr>
<tr>
<td>Felsic rocks</td>
<td>7.43-24.18</td>
<td>2.95-5.95</td>
<td>1.73-5.30</td>
<td>1.37-1.42</td>
<td>1.11-1.48</td>
<td>0.40-0.55</td>
<td>0.18-0.48</td>
<td>2.54-6.21</td>
</tr>
<tr>
<td><strong>Timiskaming</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CAB</td>
<td>7.54-11.43</td>
<td>2.15-2.58</td>
<td>2.28-2.43</td>
<td>0.79-0.93</td>
<td>0.70-0.86</td>
<td>0.56-0.59</td>
<td>0.16-0.25</td>
<td>2.89-4.25</td>
</tr>
<tr>
<td>Felsic rocks standard</td>
<td>14.67-22.96</td>
<td>3.04-3.68</td>
<td>3.09-3.52</td>
<td>0.61-1.04</td>
<td>0.59-0.86</td>
<td>0.28-0.37</td>
<td>0.09-0.16</td>
<td>6.16-12.05</td>
</tr>
<tr>
<td>Felsic rocks Th-enriched</td>
<td>32.55-44.75</td>
<td>3.38-4.26</td>
<td>4.38-5.43</td>
<td>0.39-0.81</td>
<td>0.37-0.66</td>
<td>0.15-0.24</td>
<td>0.04-0.10</td>
<td>11.19-18.28</td>
</tr>
</tbody>
</table>
REE patterns (see Figure 9.2b). The Keewatin basalt is less fractionated than the Timiskaming basalts. The Timiskaming metavolcanic rocks have negative Zr and Hf anomalies which are absent in the Keewatin metavolcanic rocks (see Figure 9.2b, 9.2c). Eu is also a fractionating element throughout the volcanic rocks of the Shining Tree area. The Timiskaming metavolcanic rocks are more fractionated than the Keewatin metavolcanic rocks and the felsic and intermediate rocks of each assemblage are more fractionated than the respective mafic volcanic rocks.

Thus, the tholeiites may have formed in a marginal basin spreading axis, or greatly attenuated continental crust. The calc-alkaline rocks form a sequence from basalts to rhyolites in the Timiskaming assemblage, but there are also Keewatin aged calc-alkalic rocks. These are clearly influenced by crustal material and have an arc signature. However, the tectonomagmatic implications which can be drawn from these preliminary analyses obviously need to be corroborated by further data.

**SUMMARY**

The preliminary results discussed above demonstrate that the Shining Tree area is a geologically complex one. It does not constitute a simple evolving sequence of basalts to rhyolites. There are two main geochemical units within the Keewatin aged rocks, one tholeiitic with associated komatites and one calc-alkalic with an arc signature. These did not evolve one from the other by fractional crystallisation, but came from two distinct magma sources. The relationship between these two Keewatin units will become clear by the processing of geochemical data as it becomes available and by the continued examination of field data. The Timiskaming assemblage unconformably overlies the Keewatin sequences. There are no tholeiites in the Timiskaming, all of the igneous rocks assessed at this stage are calc-alkalic. The rocks form a sequence from basalts to rhyolites. They have fractionated Nb and Ti, a characteristic of an arc influenced tectonic environment. They have also fractionated Zr and Hf, possibly a diagnostic property to separate Timiskaming and Keewatin rocks. The Timiskaming felsic metavolcanic rocks can be further subdivided into a Th-enriched group. There are many questions unanswered with regard to these rocks and the tectonic setting of this area. The volcanic rocks of both assemblages may be intimately associated or they may represent entirely different packages of a similar age which have become tectonically interleaved as Tomlinson, Hughes and Tomlinson (1995) found in the Wabigoon.

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10. Project Unit 98–013. Geology of Sheraton Township, District of Cochrane

C. Vaillancourt
Ontario Geological Survey, Precambrian Geoscience Section

INTRODUCTION

The discovery of significant copper-zinc-lead-silver-gold mineralization southeast of Timmins has triggered the need to explore the setting of these deposits. A new mapping project at a 1:20,000 scale has been initiated by the Ontario Geological Survey in Sheraton Township to meet this need. The discovery site is located in the Sheraton–Timmins property (100% Cross Lake Minerals Ltd.). Exploration work in the area also covers the Night Hawk Lake Joint Venture property (Golden Knight Resources Inc.–Cross Lake Minerals Ltd.–East West Resource Corp.–Canadian Golden Dragon Resources Ltd.), just north of the Sheraton–Timmins property.

Sheraton Township is located approximately 20 km southwest of Matheson, south of Highway 101 (Figure 10.1). The western side of the township is accessible by the Gibson Lake road and the eastern side by the Currie–Bond road, both extending south from Highway 101. Many branching smaller roads and walking trails provide easy access to most of the township from these two main roads.

Previous mapping covering the area includes the 1:63,360 scale map of the Langmuir–Sheraton area by Berry (1940) and the 1:63,360 scale map of the Watabeag River area by Pyke (1976). Both maps showed very limited outcrop areas, restricted to the eastern part of the township.

The exploration companies currently working on their properties in Sheraton Township have given complete access to their data, which are concentrated in the northwestern part of the township. These include extensive diamond drilling (61 holes by Cross Lake Minerals Ltd. and 24 holes by Golden Knight Resources Inc.), geophysical and litho-geochemical studies.

Geological, geophysical and diamond drilling data from previous exploration projects in the assessment files at the Resident Geologist’s Office, Ministry of Northern Development and Mines, in Timmins have also been used to get better coverage of the entire township.

The purpose of this project is to bring together outcrop mapping and examination of diamond-drill core, along with geochemical and geophysical data, to generate a compilation at a 1:20,000 scale covering Sheraton Township. The map will emphasize the lithological, structural and economical aspects of the geology and relate them to the regional-scale stratigraphic and tectonic framework. A better understanding of the setting of the mineralization and the rock types it is associated with will provide direction for further exploration work in the area.

REGIONAL SETTING

Sheraton Township is predominantly underlain by Archean-aged rocks of the Abitibi greenstone belt, cut by minor Proterozoic diabase dikes. The supracrustal rocks were included within the Watabeag assemblage of Jackson and Fyon (1991), located south of the Porcupine–Destor and north of the Larder–Cadillac major deformation zones. It is surrounded to the north and west by the Bowman and to the east by the Kinojevis assemblages in the classification of Jackson and Fyon (1991). Redefinition of larger scale assemblages by Ayer et al. (this volume) places the township at the contact between the Upper Tisdale and the Kinojevis mega-assemblages (see Figure 10.1). The Jackson and Fyon (1991) assemblages are summarized below.

Watabeag Assemblage

Because it is very poorly exposed, the Watabeag assemblage is not well known but it seems that it is predominantly composed of intermediate to felsic metavolcanics. At the northern contact with the Bowman assemblage, an east-west striking felsic metavolcanic unit has been defined (Leahy 1965; Pyke, Naldrett and Eckstrand 1973). The contacts with the Bowman, as well as with the adjacent Kinojevis, are all interpreted to be faults. The Huronian Supergroup unconformably overlies the western part of the assemblage and its relationships with other assemblages to the south are unknown.

Bowman Assemblage

The Bowman assemblage is composed of peridotitic and basaltic komatiites interlayered with magnesium-rich basalts (Leahy 1965; Pyke 1982). Also, calc-alkalic felsic and intermediate rocks form the top of the assemblage. It is bounded to the north by the Porcupine–Destor Deformation Zone and curls around east and south of the Shaw Dome. The assemblage is thought to face toward the south. The age of the Bowman is constrained to ~2705 Ma (Jackson and Fyon 1991).

Kinojevis Assemblages

The Kinojevis is separated into two assemblages, Kinojevis North and South. The Kinojevis North assemblage is a south-facing succession of pilloww, tholeiitic basalt and minor rhyolite with scarce interflow metasedimentary rocks. The basalts form magnesium-rich and iron-rich distinct units. Some of the flows have variolitic and/or feldspar-phyric textures (Jensen and Langford 1985;
Fowler, Jensen and Peloquin 1987). The assemblage is bounded to the north by the Porcupine–Destor Deformation Zone, where three gold mines and many gold occurrences are present.

The Kinojevis South assemblage is composed of massive and pillowed tholeiitic metabasalts and very local minor interbedded iron formation, tuff and clastic metasedimentary rocks (Thompson and Griffis 1944; Goodwin 1979). It is folded and faulted with structures being east trending and shallowly to moderately plunging. The assemblage is bordered by the Huronian Supergroup to the west and unconformably overlain by the younger Timiskaming assemblage to the south.

Although the exact relationship between the Kinojevis North and South assemblages is not very well known, it seems that the age of ~2701 Ma of a metarhyolite from the North assemblage also represents an approximate age for the South assemblage.

There is evidence indicating that the contact between the Kinojevis and Blake River assemblages is locally of a tectonic nature, although Jackson and Fyon (1991) pointed out an originally stratigraphic contact. Ayer et al. (this volume) suggest conformable overlying of the Kinojevis by the Blake River.

**Blake River Assemblage**

Calc-alkaline metavolcanic rocks ranging from mafic to felsic and small mafic and felsic intrusions form the Ontario part of the Blake River assemblage. However, its Quebec counterpart also includes tholeiitic basalts and FHH rhyolites (Lesher et al. 1986). The marginal units face toward the centre of the assemblage, suggesting a syncline. An age of ~2701 ± 2 Ma (Corfu et al. 1989) has been assigned to a rhyolitic unit in the centre of the assemblage. Only minor base metal mineralization is known to occur in the Ontario side of the Blake River assemblage, but important mineralization is found within a chemically distinct unit of the assemblage in Quebec.

**DESCRIPTION OF ROCK UNITS**

**Felsic to Intermediate Metavolcanic Rocks**

Felsic to intermediate metavolcanic rocks have only been recognized in diamond-drill core in the northwestern part of the township. Two distinct zones are referred to below. The first is the Sheraton Lake zone, directly to the west of Sheraton Lake. The second is the Cross Lake zone, located approximately 1.5 km to the southwest.

On a triangular plot of Al₂O₃ - Fe₂O₃ + TiO₂ - MgO (Jensen 1976), most of the intermediate-felsic pyroclastic rocks define a calc-alkalic trend. Samples plotting in the tholeiitic field can be explained by enrichment in iron due to the presence of abundant pyrite. The pyroclastic rocks are mainly dacitic to rhyodacitic in composition (Bowen 1998). No massive units of any composition have been

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**Figure 10.1.** Revised stratigraphic subdivisions of the area around Sheraton Township, northern Abitibi greenstone belt, Ontario. Modified from Ayer et al. (this volume).
recognized. Only the minor metre-scale rhyolitic intersections could possibly be interpreted as massive, possibly flows, but the very strong, mostly silicic alteration could very well be overprinting the fragmental nature of the protolith. Graphitic argillites and other wacke intervals are locally interbedded with the pyroclastic sequences. The pyroclastic units can be described as mostly lapilli-tuff to fine tuff dominated sequences, and crystal-tuff sequences.

**Lapilli Tuff/Tuff**

In the tuffs and lapilli tuffs, the presence of chloritized pumiceous shards and groundmass quartz, but the absence of quartz phenocrysts, indicates a dacitic composition. The size of the fragments varies greatly from fine ash almost to bombs. The geochemical data of Bowen (1998) give a composition ranging from dacite to rhyodacite for most of the intermediate tuffaceous units. The tuffs contain mostly monolithic fragments. Intervals of heterolithic fragments, along with bedding, infer reworking, at least locally.

A felsic tuff breccia sample taken from drill core in the Cross Lake zone has a U/Pb zircon age of 2703.7 ± 1.4 Ma (Ayer et al., this volume). This age falls into the range of other felsic volcanic rocks along strike to the east in Currie Township (2706 ± 2 Ma; Corfu 1993) and of the Watabag assemblage in Douglas Township (2703 ± 2 Ma; Corfu 1993).

**Crystal Tuff**

The crystal tuff units are distinct from the lapilli tuff as they rarely contain any lapilli-sized lithic fragments. They are mostly composed of plagioclase crystals, fragmented plagioclase crystals and local, small, 1 to 2 mm, black chloritic shards in a very fine ash matrix. A darker greenish tint gives them a more mafic appearance than the lapilli tuff/tuff sequences. Bowen (1998) observed that analyzed samples of this unit are basalts and andesites in composition. This could be the result of a more mafic protolith or alteration. Additional petrography and geochemistry may help to resolve the nature of the compositional differences.

The feldspar-porphyry dikes show many resemblances in appearance and, to a certain extent, in their geochemical composition to the crystal tuffs. This could imply a common magma source.

**Mafic Metavolcanic Rocks**

The mafic metavolcanic rocks of the mapped area can be subdivided in two groups. A pyroclastic calc-alkaline basaltic sequence is found to the south of the felsic to intermediate metavolcanic units and is most probably genetically related to it. The presence of plagioclase and hornblende phenocrysts in the fragmental volcanic rocks south of the Sheraton Lake zone infers its calc-alkaline affinity. It can be described as a tuff breccia, possibly a reworked pyroclastic. As it is stratigraphically on top of the felsic to intermediate rocks, this unit represents the uppermost exposed flow of the sequence.

**Tholeiitic mafic volcanic rocks** occur farther to the south in a different package that could be part of the Kinojevis South assemblage of Jackson and Fyon (1991). A series of mafic massive, pillowed and variolitic lavas and many breccias, probably pillow breccias and hyaloclastites underlie the southeastern part of the township. A strong east-west trending fabric makes accurate identification locally difficult. However, a few well-preserved pillow lavas consistently indicate tops to the north. Variolitic units represent a good marker horizon within the mafic volcanic sequence.

**Metasedimentary Rocks**

Interbedded sedimentary units are found locally in the felsic to intermediate pyroclastic rocks in the Sheraton zone and farther northeast, but are absent in the Cross Lake zone. Facies variations in the fragmental volcanics and the greater abundance of sedimentary units suggest closer proximity to a volcanic centre at the Cross Lake zone.

Many metre-scale horizons of graphitic argillites are present in the Sheraton zone. They commonly contain pyritic nodules and occasionally small specks of red-orange sphalerite. Local normal grading in argillite, siltstone and coarse sandstone consistently indicates younging to the south.

No chemical sediments were observed in diamond-drill holes or in outcrops. Highly silicified sections of the mineralized zones have been interpreted by some as cherts, but the pyroclastic texture of such sections is commonly recognizable.

**Metagabbros**

The metagabbros located in the southeast part of the township are very well preserved but locally are cut by deformation zones that transformed them into chlorite schists. Local primary magmatic layering, centimetre-scale ultramafic patches and pegmatic phases have been observed, but the overall composition is generally homogeneous. These have a very coarse grain size and consist of mainly plagioclase, hornblende and locally magnetite. As the major mafic phase is hornblende, the gabbro could be genetically related to the syenitic pluton to the northeast. Petrographic work along with geochemistry will be required to determine if the gabbro is genetically related to syenite or the mafic volcanic rocks.

**Syenite**

The Bradley Lake Syenite pluton occurs on the eastern boundary of Sheraton Township (Pyke 1976). Two different phases have been observed just outside the township. The main phase is a coarse-grained white unit with a subhedral granular texture. Pink hematization only occurs along fractures. Macroscopically, the only mafic minerals are hornblende and magnetite, together representing around 5% of the rock. The second younger phase is pink with a porphyritic texture. Phenocrysts are potassium-feldspar and plagioclase with well-developed zoning. The phenocrysts vary between less than 1 mm and up to 5 mm.
The mafic minerals are smaller and less abundant but consist of amphiboles and minor magnetite. The latter sometimes occur as inclusions in the zoned feldspar phenocrysts.

**Granodiorites**

Granodiorite was observed in the diamond-drill core southwest of the Cross Lake zone. It is probably similar in composition to the Blackstock pluton and may represent a small apophysis of the pluton. The grain size (2 to 4 mm) and proportion of minerals is very homogeneous in two different holes. Locally the plagioclase is zoned and is larger in size than the other minerals, giving the rock a slightly porphyritic texture. Chlorite is the most abundant mafic mineral, probably replacing biotite and/or amphiboles.

**Porphyry Dikes**

Feldspar and quartz-feldspar porphyritic dikes are found throughout the township in outcrop and in diamond-drill core. The dikes crosscut the mafic volcanic packages of the southeastern part of the township and were also observed in the intermediate to felsic rocks in the northwestern part. Some of these dikes (feldspar porphyritic) could be related to the syenite pluton as they are very similar in mineralogical composition, alteration and deformation style. However, some of the quartz-feldspar porphyry dikes observed in the intermediate to felsic volcanic series are too quartz-rich, foliated and highly altered and are more likely of a syngneous origin.

**Diabase Dikes**

Members of two diabase dike swarms occur in Sheraton Township. They are the Matachewan dikes and the Abitibi dikes of Paleoproterozoic and Mesoproterozoic ages respectively. Only the major dikes are shown on Figure 10.2.

**METAMORPHISM**

Mineralogy and texture in mafic volcanic rocks of the southeast part of the township indicate that metamorphism has achieved upper greenschist facies. Metamorphic recrystallization has locally produced a fine-grained granular texture in the massive volcanic rocks as a result of recrystallization of plagioclase. The presence of garnet in some of the mafic (probably andesitic) volcanic units also indicates elevated metamorphic conditions. Assimilation textures of the volcanic rocks by felsic intrusives have been recognized in the southeast. It may be that the eastern part of the township has experienced contact metamorphism by the Watabeag batholith and associated satellite plutons. Further petrologic and textural work will be required to better estimate the intensity and variation of the metamorphism in the area.

In the northern felsic to intermediate rocks, chlorite, epidote, and, as reported by Bowen (1998), zoisite and clinzoisite indicate degrees of metamorphism between epidote–hornfels and upper greenschist. The local presence of chloritoid in the volcanic rocks is probably indicative of hydrothermal alteration.

**STRUCTURE**

Apart from very few well-preserved sectors, deformation is very strong all across the tholeiitic mafic volcanic rocks. Schistosity is locally developed to the level of forming chlorite schists of unidentifiable protolith. Breccias have everywhere taken a striped look as the fragments are highly stretched. Schistosity is commonly parallel to bedding. It strikes east-west to north-west, with a subvertical dip in the southern part of the township, and north-south, with a moderate dip toward the east in the northern part. As discussed further below, this change in orientation from the southern to the northern parts of the township, along with other lithological evidence, could be a result of large-scale folding.

No obvious mineral or stretching lineation was observed in the mafic metavolcanics. It is possible that the deformation style in the southeast was pure compression with little, if any, shearing. Near the southern township boundary, east-west trending faults at the border of the gabbro show subhorizontal to moderately south-dipping stretching lineation. The faults locally transformed the gabbro in chloritic schists, suggesting that they were active at the time of intrusion.

All pillow tops observed in the mafic volcanic rocks of the southern part of the township indicate younging to the north. From the north-south stratigraphy and tops facing to the south in southern Currie Township, reported by Pyke (1976), a large-scale synclinorium is suspected to form the eastern side of the township (see Figure 10.2).

Deformation in the calc-alkalic felsic to intermediate units is highly variable, from absent to very strong. In a general way, the sericitic alteration seems to have accommodated deformation, thus forming the sericite schists. Rhyolitic fragmental units are less deformed. Projection of the stratigraphic units from the core to the surface indicates the presence of north-northwest trending faults in order to correlate the holes. On a regional scale, these smaller faults are parallel to the Cross Lake Fault. It is possible that the Cross Lake and Sheraton Lake zones were originally continuous, but were sinistrally offset by one of these north-northwest trending faults.

**ALTERATION**

Most of the pyroclastic rocks from the diamond-drill holes are heavily altered. The major alteration consists of silicification, sericitization and rarely chloritization. Silicification affects all pyroclastic units and locally results in a rock almost entirely replaced by quartz. Plagioclase crystal and volcanic shards are locally totally replaced by quartz. In general, silicification is variable in intensity but is almost always present and not restricted to specific compositional units or zones. Sphalerite + galena are locally present in the most silicified zones.
Figure 10.2. Primary interpretation of the geological setting in Sheraton Township, northern Abitibi greenstone belt, Ontario (see Figure 10.1 for locations).
Sericitization is the second most common alteration. It occurs as yellow sericitic bands of variable intensities in the pyroclastic units. The banded nature of highly sericitized sections has been called "tiger striped" because of its appearance. Higher zinc intersections are generally related to intense sericitization.

Chloritic alteration is not well developed anywhere in the intermediate to felsic pyroclastic units. The chlorite component of the tuff matrix locally gives it a greenish colour but it is not necessarily alteration. Petrographic work will bring a better understanding of the nature of the chlorite and other alteration types.

MINERALIZATION

Cross Lake Minerals Ltd. has reported concentrations of zinc in the range of 1% to 6%, with local higher grade pods containing 18% zinc over a true width of 3 m. Reported copper values are 1% to 3% over estimated true widths of up to 12 m (from Internet site http://www.crosslakeminerals.com). These values are from the Cross Lake zone.

The Cross Lake zone shows generally higher concentrations in zinc relative to copper. However, the more distal Sheraton Lake zone generally has higher copper values. Pyrite is the most abundant sulphide in all of the core examined. The sulphides in the volcanic units appear to be largely remobilized as the 1% to 15% pyrite commonly occurs in or around small fractures and quartz or quartz-carbonate veins. The minerals of economic interest are sphalerite, chalcopyrite and galena. The sphalerite is iron-poor, with beige to almost white colour, and is associated with highly sericitized and silicified zones. It is usually concentrated in approximately 1 m to 10 m horizons and mostly disseminated to locally semi-massive. Higher concentrations of chalcopyrite are present in quartz stringers and occasionally form blebs and bands in the lapilli tuffs. Few small highly siliceous areas contain traces of galena.

DISCUSSION AND CONCLUSIONS

A generalized geological interpretation is presented on Figure 10.2. The figure shows contacts between three different assemblages, the mafic tholeitic sequence occurring in the east, a separate tholeitic unit in the southwest and the mafic to felsic calc-alkalic sequence in the north. If the extensive horizon of variolitic pillow basalts in the southeast is used as a marker unit in the tholeiitic mafic rocks (represented as finely dashed lines in Figure 10.1), it suggests synclinal folding. This is consistent with the large-scale synclinorium centred in the Kinonevis and Blake River assemblages east of the mapped area (Jackson and Fyon 1991). Thus the tholeiitic mafic volcanic rocks east of the calc-alkalic units should be part of the Kinonevis assemblage.

As already discussed, the calc-alkalic package is related in age and composition to similar rocks of the Bowman assemblage of Jackson and Fyon (1991). Following Ayer et al. (this volume) mega-assemblages, it would then be part of the Upper Tisdale assemblage. The lack of outcrop in southwestern Sheraton makes it difficult at the moment to be certain of the contact of the intermediate to felsic pyroclastic package to the southwest. However, based on limited diamond drilling in this area and southeastern Thomas Township, the pyroclastic units appear to be in contact with mafic tholeiitic volcanics to the west.

The calc-alkalic pyroclastics are the host of significant base metal mineralization that is still currently under investigation by Cross Lake Minerals Ltd. and Golden Knight Resources Inc. Whether the mineralized assemblage continues to the west or to the south is still a matter of debate due to poor outcropping in the area. It is hoped that clues will be found through more detailed compilation work in the production of the 1:20 000 scale map. The ongoing diamond drilling on the Sheraton-Timmins property by Cross Lake Minerals Ltd. should also bring new data on extensions at depth.

REFERENCES


Bowen, R.P. 1998. A lithogeochemistry and petrographic report on the Sheraton-Bond portion of the Night Hawk project; internal report for Cross Lake Minerals Ltd. and Golden Knight Resources Inc.


11. Project Unit 95–098. Using Radarsat Data for Geoscientific Applications, Abitibi Greenstone Belt, Ontario

Z. Madon, N.F. Trowell, B. Berger and J.A. Ayer

Ontario Geological Survey, Precambrian Geoscience Section

INTRODUCTION

A multiyear project (Ayer and Trowell 1996; Ayer et al. 1997) to produce four 1:100 000 scale geological compilation maps covering the southern Abitibi Greenstone Belt in Ontario (Figure 11.1) is currently under way (Ayer and Trowell, this volume). The first map, covering the Timmins area (northwest sheet, see Figure 11.1), was released in the spring of 1998 (Ayer and Trowell 1998; Ayer et al. 1998).

In 1997 and 1998, the Ontario Geological Survey (OGS) received Radarsat digital radar data covering most of the Timmins–Kirkland Lake portion of the Abitibi Greenstone Belt. These data were received under ADRO (Application Development and Research Opportunity) Project 662 from the Canadian Space Agency.

A key scientific objective of ADRO Project 662 is to integrate Radarsat data with geological, geophysical and other remotely sensed data sets and to enhance the geological interpretation of the Abitibi Greenstone Belt in Ontario. The suitability of Radarsat imagery to geological problems is being evaluated across a variety of bedrock and surficial terrain conditions. Specific objectives include mineral potential assessment, tectonic interpretation, exploration method development and mapping of surficial deposits.

In 1997, a Radarsat study of the Timmins area was carried out (Madon, Trowell and Ayer 1997). This

Figure 11.1. Location map of the Lightning River map area.
summary presents a preliminary structural investigation of the Lightning River area using Radarsat data.

**GENERAL GEOLOGY**

The southern Abitibi Greenstone Belt extends well into northwestern Quebec and is terminated to the west by extensive granitoid complexes along the Ivanhoe Lake Cataclastic Zone. This study confines itself to the Lightning River area (see study area, Figure 11.1).

The Lightning River area was chosen because of its variable terrane conditions and assortment of geological and structural features. It falls within the eastern portion of the Abitibi Greenstone Belt in Ontario and is 44 km by 38 km, covering an area of almost 1700 km$^2$. It is located about 100 km east of Timmins and 50 km northeast of Kirkland Lake, extending to the Quebec–Ontario border (Figures 11.1 and 11.2). The east-west trending Highway 101 bisects the southern half of the area, with logging roads providing secondary access. Lake Abitibi, the largest lake within the Abitibi Greenstone Belt, covers most of the northern third of the study area.

![Diagram of the southern Abitibi Greenstone Belt in Ontario](image)

**Figure 11.2.** The southern Abitibi Greenstone Belt in Ontario.
Figure 11.3 illustrates the most recent geological map of the area (Jensen 1982). Four main Archean volcanic assemblages of generally east-west trending metavolcanic and minor metasedimentary rocks dominate the southern two thirds of the area. These assemblages consist of the older Kidd–Munro and Stoughton–Rocquemaure groups found primarily north of the Porcupine–Dester Fault Zone (PDFZ) and the younger Kinojevis and Blake River groups found to the south of the PDFZ.

Timiskaming-type metasediments and related metavolcanics occupy a narrow east-west band within the PDFZ. All of these assemblages are cut by various ultramafic, mafic and felsic intrusives of Archean age, northeast and north-northwest trending Proterozoic diabase dikes, and minor Jurassic to Cretaceous kimberlite dikes. The Lake Abitibi Batholith underlies the northern third of the area and is composed of trondhjemites, quartz diorites and diorites.

Over 90% of the area is covered by a variety of glacial material, including glaciolacustrine, ice contact, eolian and till deposits. Fine-grained silt, clay and varved clay basin deposits of the Abitibi Clay Belt are the most extensive surficial deposits in the area. Glacioluvial ice-contact esker complexes are confined to narrow north–south ridges, rising several tens of metres above the clay plains. These include the Munro Esker, which straddles the western boundary of the study area, and the Boundary Esker, which extends discontinuously along the Frecheville–Stoughton boundary and through the central part of Holloway Township.

Frecheville Township contains an unusual triangular configuration of alternating felsic to mafic metavolcanic rocks that mimic the topographic ridging in the area and are up to 90 m in height. This feature is interpreted to be a refolded fold along east-west and north-south synclinal axes. Another significant topographic feature includes the Ghost Range—an east-west trending ridge of felsic to ultramafic metavolcanic rocks up to 190 m in height.

Gold-bearing faults and shear zones are generally subparallel to stratigraphy and are displaced by numerous northeast and northwest trending faults. The PDFZ, a 1 km wide band of sheared rock that narrows to 200 m at the Ontario–Quebec border, is the most economically important structural feature in the area and is oriented parallel to the “look” direction of the Radarsat satellite.

ACQUISITION AND PROCESSING OF THE RADARSAT DATA SET

Landsat and SPOT satellites use the sun as their energy source and record the visible to near-infrared light spectrum (0.4 μm to 12 μm). They can effectively image the Earth’s surface during daylight hours but cannot penetrate cloud cover. Radar satellites, on the other hand, transmit their own signal and can produce Earth images both day and night. Because radar waves are approximately 20 000 times longer than visible light waves, they easily penetrate cloud cover and tend to accentuate the Earth’s topography. The ability to enhance landscape features makes Radarsat imagery an important data set for a variety of geological applications, including bedrock mapping, surficial mapping, mineral exploration and geological hazard identification.

Radar images record the portion of transmitted radar waves that return to the satellite, known as backscatter. The amount of backscatter is a function of both terrain parameters—the attitude or geometry of the surface, its roughness and its dielectric properties—and system parameters—radar wavelength, radar polarization, incidence or illumination angle and “look” direction of the radar beam. Radarsat has 2 fixed “look” directions—098° in ascending orbit and 278° in descending orbit.

The processed radar image is a single-channel display of grey-scale tones that range from black (no backscatter) to white (maximum backscatter) on a typical computer monitor. Calm water bodies or flat, smooth terrain (pavement) reflect nearly all of the incident energy away from the satellite and tend to produce dark grey to black tones. Because water has a relatively high dielectric constant, wet or swampy ground will generally return darker tones than dry ground. Topographic ridges that are manifestations of structural lineaments or faults will return light tones if facing the satellite “look” direction or dark tones if facing away from the satellite “look” direction. Any ridges parallel to the “look” direction will be less discernable.

Several satellite radar images of the study area were acquired by Radarsat in the summers of 1997 and 1998, using both fine beam (F1) and standard beam modes (S1 and S7). The fine beam and standard beam modes have a nominal ground resolution of 10 m and 30 m respectively. Signal data were processed at Radarsat International and the resulting path images were resampled to a pixel size of 12.5 m for the standard beam images and 6.25 m for the fine beam image and delivered to the OGS in LGSOWG format. The Radarsat radar signal operates at C-band (5.6 cm wavelength) and is HH polarized—transmitting and receiving horizontally polarized radar waves. With its “right-looking” configuration, the satellite is capable of scanning any area on Earth during both ascending and descending orbits. The viewing or incidence angle for the F1 beam varies from 37° in the near range to 40° in the far range, the shallowest of five incidence angles available for the fine beam mode. For standard beam modes, the S1 beam has the steepest incidence angle, at 20° to 27°, while the S7 beam has the shallowest incidence angle, at 45° to 49°. Originally, the S7 beam range was thought to be optimal for detecting subtle changes in relief, although recent studies indicate that the S1 beam is optimal for areas of moderate relief (Singhroy and St.-Jean 1997).

An airborne synthetic aperture radar (SAR) survey was flown by InterMap Technologies Ltd. in the late 1980s along east-west flight lines. The aircraft was equipped with an X-band 7-look sensor (2.4 cm to 3.8 cm radar wavelength), using HH polarized radar waves. Nominal resolution is 6 m in the azimuth direction and 12 m in the range direction. Incidence angles vary from 14° to 45°, with a southerly “look” direction. The standard across-
Figure 11.3. Geology of the Lightning River area (Jensen and Langford 1985).
track radiometric correction procedure was applied to compensate for variations in antenna power pattern.

Another raster data set used in this study was a digital elevation model (DEM) derived from Ontario Digital Topographic Data Base (ODTDB) vector contours. The ODTDB coverage for northern Ontario is at a scale of 1:20 000, with contour intervals every 10 m. The ODTDB includes all medium-scale topographic maps of the Ontario Basic Mapping (OBM) program. Nine OBM maps were combined to produce a partial DEM for the study area.

Northern Ontario OBM program maps are derived from 1:50 000 scale aerial photography. OBM program maps are divided into 10 km by 10 km quadrangles based on the Universal Transverse Mercator (UTM) projection and grid system, North American Datum 1927. Each map includes hydrographic features (streams, rivers, lakes, marshes, etc.), planimetric features (roads, railways, buildings, power lines, etc.), vegetation outlines and elevation in the form of contours or digital spot heights.

PRELIMINARY GEOLOGICAL INTERPRETATION

Three prominent geological features in the study area were examined in detail using the various remote sensing data sets. Three Radarsat beams—F1, S1 and S7—were evaluated to see which mode is optimal in identifying selected geological and structural features. In order to minimize variability, all three Radarsat images are descending mode or west-looking and all were processed using the “Gibson” adaptive speckle filter (Gibson 1997). In addition, they were compared with Landsat, airborne radar and DEM data. A brief discussion of this evaluation on a feature-by-feature basis follows.

Frecheville Triangular Structure (Figure 11.4)

This structural feature consists of alternating iron-rich to magnesium-rich, mafic to locally felsic, tholeiitic metavolcanics and related intrusives of the Kinojevis Group overlying minor calc-alkaline felsic metavolcanics. The triangular-shaped basin dips towards the centre, marked by Trollope Lake, at a relatively steep angle and is interpreted to be a refolded synclinal structure along east-west and north-south fold axes. Volcanic ridges rise approximately 90 m above the surrounding terrain, attaining a maximum elevation of 400 m.

Elements of this structure are visible on all three Radarsat images, being most pronounced with the S1 image. Because of their topographic prominence and orientation with respect to the Radarsat beam, the northeast limbs are visible in all three images. The inverted V-shaped northern fold closure is visible in all three Radarsat images as well as the airborne radar image, the DEM and, to a lesser degree, the Landsat Thematic Mapper (TM) data.

As expected, the east-west trending southern limb is most visible with airborne radar. It does, nonetheless, manifest itself somewhat in the S1 image and to a lesser degree in the F1 image. The southern limb is essentially invisible in the S7 image.

Although the airborne radar provided the most detailed image with respect to landform texture, it failed to outline the fold closure in the southeast corner visible in the F1, S1 and, to a lesser degree, S7 images. Similarly, the southeast fold closure is poorly outlined on the DEM because of its coarse ground resolution. These images support the geological interpretation of a refolded synclinal structure. It would appear that, even though the F1 image has a finer resolution than the S1 and S7 images, there is no concomitant increase in the geological information derived from it.

Not being constrained by a fixed “look” direction, the overall lay of the land is best developed with the DEM, although it lacks some of the detail found in the radar imagery. With Landsat TM, some of the ridging is manifested as narrow white lines, but the lack of shadow gives it a “flatter” appearance, making it more difficult to interpret geological structures.

Ghost Range (Figure 11.5)

The Ghost Range represents a basal sequence of the Stoughton–Roquemaure Group and is composed of thick ultramafic to mafic komatiitic subvolcanic intrusives and related lava flows that range from massive to pillow to locally spinifex-textured (Jensen and Langford 1985). It is a tightly folded east-west syncline with steeply dipping north and south limbs. Tops appear to point toward the fold axis (B. Berger, OGS, personal communication, 1998). The Ghost Range is the highest topographic feature in the study area, rising 190 m above the surrounding glaciolastic clay deposits to a maximum of 480 m above sea level.

Of the three Radarsat images, the S1 beam most clearly outlines the Ghost Range. The eastern and western extremities of the Ghost Range are moderately visible with the F1 beam and only very faintly visible with the S7 beam, with the central portion blending into the surrounding area. The form and orientation of the Ghost Range is not ideal for detection by Radarsat imagery because its radar beam is subparallel to the Ghost Range syncline. Conversely, the south-looking airborne radar beam is ideal for this type of structure. The “look” direction, combined with a more detailed ground resolution, resulted in the most clearly and continuously visible representation of the Ghost Range. The DEM also produced a good representation of this feature, mainly because of the height differential with the surrounding area.

The western fold closure is clearly visible in the airborne radar image and, to a slightly lesser degree, the S1 image. By contrast, it is not as apparent with the DEM and even less visible with the F1 beam mode. The western fold closure is essentially invisible with the S7 beam. The eastern fold closure can be traced out with airborne radar but is barely visible with the S1 beam and essentially invisible with all other data sets.
Figure 11.4. Airborne radar, Gibson-filtered Radarsat, Landsat and DEM images of the Frecheville Triangular Structure.
Figure 11.5. Airborne radar, Gibson-filtered Radarsat, Landsat and DEM images of the Ghost Range feature.
Figure 11.6. Airborne radar, Gibson-filtered Radarsat, Landsat and DEM images of the Kinojevis Volcanic Hills.
Crosscutting north-northwest and north-northeast lineaments are manifested by breaks in the northern and southern limbs of the syncline. These are again most clearly visible with airborne radar. The DEM is also able to define the same breaks, surprisingly similar to the airborne radar even though its ground resolution is an order of magnitude lower. Because the central portion of the Ghost Range was not defined in any of the Radarsat images, the north-northwest and north-northeast striking lineaments are not visible.

The Ghost Range appears as a very faint, dark grey trace on the Landsat TM image and, as with the Frecheville structure, very “flat” looking. Cultural features such as clear-cuts and other logging operations create significant noise for the geological interpretation of local bedrock.

**Kinojevis Volcanic Hills (Figure 11.6)**

The Kinojevis Volcanic Hills consist of predominantly alternating iron-rich and magnesium-rich mafic tholeiitic metavolcanics of the Kinojevis Group. The metavolcanics are massive to pillowed to variolitic with local amygdaloidal and feldspar porphyritic phases. Minor interflow sediments and hyaloclastites outcrop locally. This irregular outcrop area rises to a maximum elevation of 450 m, up to 140 m above the surrounding landscape. The Kinojevis Volcanic Hills consist of a series of elongate east-northeast ridges running parallel to the general stratigraphy in the area.

This outcrop area is again most visible with airborne radar and DEM. The airborne radar was able to delineate the individual east-northeast trending flows and/or flow units. The S1 and F1 images illustrate most of the outcrop ridges and, to a lesser degree, the individual flows and/or flow units. The S7 image was able to outline only some of the more prominent basaltic ridges. There is very little indication of these ridges with Landsat TM, unless one integrates Landsat TM with a different data set.

Several en échelon arcs, open to the west-southwest, are visible on both the airborne radar and the S1 images and, to a lesser degree, the F1 image. These arcs might be indicative of local recumbent folding, or ground mapping. These subtle features are not visible with either the DEM or the S7 image.

As with the Ghost Range, crosscutting northwest lineaments are well defined with the airborne radar image and the DEM, but essentially invisible with the Radarsat imagery.

**ACKNOWLEDGMENTS**

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**REFERENCES**


12. Project Unit 96–000. The Kidd–Munro Extension Project: Results of U-Pb Geochronology for Year 1

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INTRODUCTION

The recognition of unique, komatiite-high silica rhyolite footwall rocks for the Kidd Creek volcanogenic massive sulphide (VMS) deposit (>130 million tonnes) has led to new concepts in the genesis of giant VMS deposits: for example, a relatively deep, ultramafic sill is a plausible heat source to generate such a large deposit via sub-sea-floor hydrothermal convection in a single hydrothermal cell (Barrie, Cathles and Erendi, in press). Recent developments in the volcanology of komatiites have defined proximal, medial and distal facies with respect to source vents provide a regional framework for the komatiite-rich stratigraphic section at Kidd Creek, and for komatiite-hosted nickel deposits along strike within the Kidd–Munro Assemblage and time-stratigraphic equivalent assemblages. This project is designed to apply these new concepts and develop a framework for exploration for large VMS and nickel deposits in the 2.72 to 2.71 Ga ultramafic-rich terrane of the southern Abitibi subprovince. Specifically, the goals for this 5-year project are: 1) to provide a detailed architectural framework for the komatiite-rich volcanic terrane of the southern Abitibi subprovince, with emphasis on volcanic facies of komatiite flows, high magnesium basalts (contaminated komatiites), and rhyolite volcanic centres; 2) to correlate the Kidd Creek stratigraphic succession with other successions on scales of kilometres, tens of kilometres and ultimately across the southern Abitibi subprovince, using volcanology, structural geology, chronostratigraphy, and U-Pb geochronology; and 3) to develop exploration criteria specific to this terrane to help locate targets for large VMS and nickel deposits.

This report provides results from a five year research project on the controls of base metal mineralization in the Abitibi greenstone belt funded by an industry-government partnership (Barrie 1996). For Years 1 and 2 of the project, 5 preliminary maps, 189 research grade whole rock analyses, 12 U-Pb zircon/baddeleyite geochronology analyses, and 6 galena Pb isotope analyses have been conducted; and 2804 whole rock geochemistry samples have been collected from participating companies and considered in the production of the preliminary maps. In this summary report, the U-Pb geochronology for Year 1 is presented, and initial interpretations of the significance of these data are given. Final interpretations for these data and for the geochronology component for Year 2 will be given elsewhere.

The general geology for part of the western Abitibi Subprovince with the U-Pb ages determined for this project is given in Figure 12.1. In Figure 12.2, the geology for the Kidd Creek–Dundonald region is presented, highlighting the correlation between these two areas within the Kidd–Munro assemblage. In Figure 12.3, the geology with U-Pb ages for dacite units intercalated with komatiites and nickel deposits of northwest Langmuir Township is given.

GEOCHRONOLOGY

The oldest known strata in the Kidd–Munro assemblage are the Kidd Creek Footwall Mine Rhyolites and intercalated ultramafic flows. The U-Pb zircon age for monolithic rhyolite flow breccia from the East Outcrop is 2717 ± 2 Ma (Barrie and Davis 1990; using data from Nunes and Pyke 1981 and an additional abraded, concordant fraction at 2716.2 ± 1.6 Ma). A massive, flow banded rhyolite from within 50 m on the same outcrop has an age of 2714.3 ± 1.2 Ma, overlapping with the previously determined age at the 2σ level, but with a slightly younger age (Bleeker and Parrish 1996). In the hanging wall, one U-Pb zircon age for the Quartz Porphyry high silica rhyolite is 2711.5 ± 1.2 Ma (Bleeker and Parrish 1996), giving a maximum period of 2.3 to 8.7 my for the formation of the ore, considering the 2σ uncertainties of the ages. The Porcupine Group greywackes south of a sheared and unconformable contact with the Kidd Creek stratigraphic succession have detrital zircons that indicate deposition at least 2.69 Ga (Bleeker and Parrish 1996).

Generally, pre-Kenoran strata from the southern Abitibi Subprovince, includes the less-altered rocks from the Kidd Creek Volcanic Complex, have little evidence of zircon inheritance, and have primitive whole rock Nd, Hf and Os isotopic compositions, consistent with a derivation from the mantle at circa 2.7 Ga (Machado, Brooks and Hart 1986; Walker, Shirey and Steecher 1988; Barrie and Shirey 1991; Corfu and Noble 1992; Barrie, Cousens et al., in press). The areas underlain by relatively old crust as shown by U-Pb zircon ages greater than 150 my older than the Kidd–Munro assemblage are several hundred kilometres away, either to the west in the Wawa Subprovince (Turek, Smith and Schmus 1984), the southwest in the Wawa Gneiss Domain (Moser et al. 1996), or to the northeast in the Opatica Subprovince (Davis, Gariepy and Sawyer 1994).
Figure 12.1. U-Pb sample locations in the western Kidd–Munro assemblage and adjacent areas for Year 1 of the Kidd–Munro Extension Project.
The effects of post-Kenoran thermal events at circa 2.65 Ga are recorded in a variety of isotopic systems at Kidd Creek. Monazite, rutile and titanite U-Pb studies in and near the mine constrain the timing of post-mineralization thermal events, with ages from 2.62 Ga to 2.66 Ga (Schandl, Davis and Kroch 1990; Davis, Gariepy and Sawyer 1994). Regressions of Sm-Nd and Rb-Sr data for variably altered rhyolite yielded ages of 2.67 Ga and 2.58 Ga, respectively (Maas et al. 1986). $^{40}$Ar-$^{39}$Ar analyses of fuchsite taken from the “east-west” shear zone that cuts the North and Central orebodies established an average plateau age of 2618 ± 18 Ma (Smith, Schandl and York 1993). Lead isotope systematics have been used for an age estimate for the ore-host rock system at 2.64 ± 0.06 Ga (Bugnon, Tera and Brown 1979), which agrees well with a $^{207}$Pb/$^{204}$Pb to $^{206}$Pb/$^{204}$Pb regression age for 14 whole rock samples from the stratigraphic sequence of 2.65 ± 0.08 Ga (Barrie, Cathles and Erendi in press).

Results

The ages reported here are the preferred ages based on regressions of U-Pb data and on the most concordant data. The ages and a summary of the regression parameters are given in Table 12.1. The U-Pb data for individual fractions and Concordia diagrams will be given elsewhere.

MANN AREA

U-Pb zircon ages have been determined for three units in the Mann area: the large Mann mafic-ultramafic intrusion, a dacite fragmental unit in Little Township up-section from the intrusion and broadly along strike with the Jonsmith Zn-Cu occurrence, and a dacite tuff in Lamarche Township to the north (Figure 12.1).

Sample 96TB0085: Gabbro, Mann Mafic-ultramafic Intrusion

Like the Dundonald sill, the Mann mafic-ultramafic intrusion represents a sizable mafic-ultramafic intrusion with the potential to drive hydrothermal convection to form VMS deposits. It has been postulated that the Mann mafic-ultramafic intrusion may have represented a heat source for the Kidd Creek deposit 25 km to the southwest, which has since been separated from the Kidd Creek stratigraphy during regional faulting. This sill and adjacent felsic volcanic rocks were sampled for U-Pb geochronology to test whether the Mann mafic-ultramafic intrusion was

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**Figure 12.2.** Simplified geology of the area from Kidd Creek to Dundonald Township with U-Pb ages and sample locations.
emplaced into a coeval volcanic stratigraphic succession. Based on a regression of four analyses, an upper intercept age of 2706 ± 4 Ma, and a lower intercept age of 1440 Ma is calculated.

Figure 12.3. Geology of northwest Langmuir Township with U-Pb sample locations. The map to the left is 5 km across.

Table 12.1. Summary of U-Pb data for Year 1 of the Kidd–Munro Extension Project.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Unit</th>
<th>Assemblage, (sills in brackets)</th>
<th>Township</th>
<th>Age, in Ma</th>
<th>2-sigma error, in Ma</th>
<th># of fractions</th>
<th>Basis of age determination</th>
<th>Probability of fit in %</th>
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</thead>
<tbody>
<tr>
<td>96TB021</td>
<td>Dundonald MUM intrusion</td>
<td>(Kidd–Munro)</td>
<td>Dundonald</td>
<td>2707</td>
<td>+3/-2</td>
<td>4</td>
<td>Regression: Davis, 1983</td>
<td>91</td>
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<tr>
<td>96TB053</td>
<td>Dundonald dacite/rhyolite</td>
<td>Kidd–Munro</td>
<td>Dundonald</td>
<td>2717.3</td>
<td>+/-1.2</td>
<td>5</td>
<td>Regression: Davis, 1983</td>
<td>84</td>
</tr>
<tr>
<td>96TB082</td>
<td>Little township dacite, DCR</td>
<td>Duff–Coulson–Rand</td>
<td>Little</td>
<td>2709</td>
<td>+/-3</td>
<td>4</td>
<td>Regression: Davis, 1983</td>
<td>69</td>
</tr>
<tr>
<td>96TB085</td>
<td>Mann MUM intrusion</td>
<td>(Duff–Coulson–Rand)</td>
<td>Mann</td>
<td>2706</td>
<td>+7/-4</td>
<td>4</td>
<td>Regression: Davis, 1983</td>
<td>61</td>
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<tr>
<td>96TB050</td>
<td>Patton occurrence dacite, Scapa</td>
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<td>Stimson</td>
<td>2726</td>
<td>+7/-5</td>
<td>4</td>
<td>Regression: Davis, 1983</td>
<td>83</td>
</tr>
<tr>
<td>96TB091</td>
<td>Lamarche dacite tuff, Scapa</td>
<td>Scapa</td>
<td>Lamarche</td>
<td>2725</td>
<td>+7/-5</td>
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<td>Regression: Davis, 1983</td>
<td>83</td>
</tr>
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<td>96TB0079</td>
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<td>Langmuir</td>
<td>2708</td>
<td>+/-2</td>
<td>4</td>
<td>Regression: Davis, 1983</td>
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<tr>
<td>96TB0099</td>
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<td>Langmuir</td>
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<td>+/-1.5</td>
<td>4</td>
<td>Regression: Davis, 1983</td>
<td>58</td>
</tr>
</tbody>
</table>
Sample 96TB0082: Felsic Volcanic Breccia, Little Township

The sample come from upsection of the southern part of the Mann mafic-ultramafic intrusion. There was some difficulty in finding a suitable felsic volcanic rock. Much of the material available in the Timmins drillcore library from Duff, Little, and southwest Mann townships was tuffaceous-sedimentary, or very fine-grained and only sparsely plagioclase porphyritic, and thus not suitable for extracting zircon. This sample is from a relatively fresh, dacite breccia. A regression of the four analyses, including three nearly concordant ones yields an upper intercept age of 2709 ± 3 Ma. The lower intercept age is Grenvillian at 950 Ma. It is noted that the two nearly concordant points have an age that is effectively identical to the two least discordant points of the Mann mafic to ultramafic intrusion at approximately 2705 Ma. The Mann mafic-ultramafic intrusion and the Little Township dacites are coeval within error for their regressions, and this is supported by the close agreement for their least discordant concordant points.

Sample 96TB0091: Felsic Tuff, Lamarche Township

Regional airborne magnetic and EM surveys indicate that the Patton area rhyolites and strata probably continue to the west for several tens of kilometres. One sample of a rhyolite was taken from a Noranda drillhole in Lamarche Township to test if indeed the felsic rocks are the same age and composition as those found in the Patton occurrence. A regression based on four analyses yields an upper intercept at 2725 ± 7 Ma, coeval to the Patton occurrence felsic volcanic unit at 2726 ± 7 Ma.

DUNDONALD AREA

Ages have been determined for samples from the Dundonald sill and felsic rocks near the base of the volcanic stratigraphic succession (Figure 12.2).

Sample 96TB0053: Sub-pegmatitic Gabbro of the Dundonald Mafic-ultramafic Sill

The Dundonald sill sample (96TB0053) is from sub-pegmatitic material taken within 25 m of the upper contact. Zircon and baddeleyite have been analyzed for this sample. An age of 2707 ± 3 Ma defined by four concordant, or nearly concordant analyses.

Sample 96TB0021: Dundonald Felsic Flow

The Dundonald felsic flow sample (96TB0021) is from south of the west-facing Dundonald W-fold, upsection from the sill and underlying the komatiites that are host to the Ni and Zn-Cu mineralization. Because this unit and the overlying komatiites have peperitic textures involving the same graphitic argillite interfloow units, they are effectively coeval with the komatiites, and with the contained base metal mineralization. The zircon crystals comprise common prismatic types. Based on five analyses of abraded zircon and an average value of 2717.3 ± 1.2 Ma is calculated. This age is older by approximately 11 my than the Dundonald sill.

Fox-Stimson Area

Sample 96TB0050: "Patton Rhyolite", Coarse Felsic Fragmental

To determine the age of mineralization for the Patton occurrence, a dacite-rhyolite volcanic breccia adjacent to the mineralization was sampled (see Figure 13.1 in Barrie, this volume). The sample has undergone moderate hydrothermal alteration and metamorphism to the amphibolite facies, but the primary volcanic textures are preserved, and the contacts of the volcanic breccia from other rock types are discernable. In addition to zircon the sample contains abundant rutile and titanite. (These minerals could be useful as they may record the timing of thermal metamorphism and possibly of uplift in this region.) Three nearly concordant analyses and one moderately discordant analysis constrain a line with an upper intercept age of 2726 ± 3 Ma.

SHAW DOME AREA

Two samples were taken in Langmuir Township of the Shaw Dome area to determine the age of komatiite volcanism associated with the Langmuir nickel deposits, and for comparison with the komatiites in the Kidd–Munro assemblage to the north (Figure 12.3). Previous age determinations in the Eldorado Assemblage in the Shaw Dome area have yielded equivocal results. For a previous age determination on a dunite-associated gabbro, it was unclear whether the age was inherited from an older crust (Corfu et al. 1989).

Sample 96TB0079: Felsic Volcanic, Footwall to Langmuir #2

This sample contains a more complex zircon population than in the other cases. At least some of crystals reveal the presence of optically detectable cores with thin overgrowths. The presence of inherited component is confirmed by analysis #25 of one, apparently homogeneous, short-prismatic grain that yields an apparent 207Pb/206Pb age of 2724 ± 4 Ma. A line calculated through the latter data points and the discordant analysis #29 has a lower intercept age of 700 Ma and an upper intercept age of 2708 ± 3 Ma, which is interpreted to be the crystallization age of the sample.
Sample 96TB0099: Southeast Langmuir Felsic Fragmental

This sample was taken from a felsic volcanic rock apparently up-section, that is, across strike in a stratigraphic section with uniform facing directions, in an attempt to help "bracket" the timing of nickel-bearing komatite magmatism in this area. Note that there are areas of poor exposure between the location of this sample and that of the other Langmuir sample. Three concordant zircon analyses and one discordant analysis yields a regression with an upper intercept age of 2724.4 ± 1.5 Ma.

INITIAL INTERPRETATIONS

Mann Area

The most prominent feature of the Mann area is the Mann mafic-ultramafic intrusion. The U-Pb age of 2706 ± 2 Ma for the sill is approximately 10 Ma younger than the Kidd Creek VMS deposit 25 km to the southwest, and therefore cannot be a heat source for the Kidd Creek hydrothermal system. The less precise regression age for the Mann mafic-ultramafic intrusion is coeval within error to the Little Township dacite, and this is supported by the least discordant analyses for both samples which average approximately 2705 Ma. Furthermore these ages are identical within error to a U-Pb zircon age of 2706 ± 2 Ma for a nearby Duff-Coulsen-Rand assemblage rock (Ayer et al. 1997).

Given the coeval relationship between an apparently high level, large and hot mafic-ultramafic sill and overlying volcanic rocks, this area appears to be a favourable setting for VMS mineralization. Under favourable circumstances, a sill the size of the Mann mafic-ultramafic sill could produce multiple ore-generating hydrothermal cells and lead to orebodies on the order of 0.5 to 5 million tonnes (Barrie, Cathles et al., in press). Furthermore, in the absence of significant permeability variations, there should be a regular pattern to hydrothermal upwelling above a sill like the Mann that is approximately 3.8 times the distance to the paleo-seafloor (Barrie, Cathles et al., in press). The majority of the Duff-Coulsen-Rand rocks are tuffs and fragmental rocks which are suitable for trapping upwelling hydrothermal fluids. However, to date the only known base metal occurrence of significance is the Jonsmith occurrence, and with the exception of widespread, pervasive carbonate alteration in the region, there is no evidence for significant, concentrated hydrothermal alteration. Pillowed basalts are intercalated with the more felsic tuffs and fragmentals, so it would appear that the area was subaqueous at the time of volcanism and sill emplacement.

Dundonald Area

A correlation between the Kidd Creek footwall volcanic succession and the Dundonald volcanic succession is confirmed here on the basis of U-Pb geochronology, and the presence of the distinctive komatiite–low titanium basalt stratigraphic section in both locations (Figure 12.2). At Kidd Creek, the footwall stratigraphic succession of variably contaminated komatites, low Ti basalts and rhyolite flow and flow breccia units are 2717 ± 2 Ma to 2714.3 ± 1.2 Ma (Barrie and Davis 1990; Bleeker and Parrish 1996), identical within error to the Dundonald dacite age of 2716.8 ± 2.1 Ma reported here.

In Dundonald Township, the Terminus Zone Zn-Cu mineralization is upsection from the thick peridotite/dunite core of the Empire Flow in the thermal erosion channel. The U-Pb age for the Dundonald sill indicates that the sill is approximately 10 my younger than the strata that hosts the Zn-Cu mineralization, and this precludes the sill as a heat source. Thus the heat source for hydrothermal circulation was likely the thick ultramafic cumulate of the Empire Flow. A peridotite/dunite flow core has an extremely high contained enthalpy if it is allowed to crystallize slowly. Cooling and crystallization of a peridotitic komatite magma has approximately twice the contained enthalpy of a basalt, due to the very high latent heat of crystallization of olivine, as well as the higher emplacement temperature of a peridotitic komatite magma (Sparks, 1986). Although flow units are generally considered to cool too rapidly to transfer significant heat to a hydrothermal convection cell, the peridotite/dunite core of the Empire Flow would have been protected from the cooling effects of a collapsing hydrothermal cell by an impermeable serpentinite rind. The effects of hydrothermal fluid-rock interaction are manifested as strong chlorite alteration and quartz veining immediately outside the peridotite/dunite core in the adjacent heterolithic breccias.

Shaw Dome Area

There are at least two ages for volcanism in northwest Langmuir Township, and the Ni-bearing komatites are clearly younger than 2708 ± 2 Ma, so the program in this area has been successful in one of its main goals. The Langmuir #2 Ni mine footwall dacite is clearly older than the overlying komatite flows that are host to the Ni mineralization. The unit extends along strike for more than 1 km at or near the surface down-section from the Ni-bearing komatite flow (Green, 1978; Green and Naldrett, 1981). Outcrops of the dacite unit show that they are cut and partially melted by ultramafic dikes, and the development of ocellar, partial melt textures adjacent to the ultramafic dikes, an U-Pb age of 2708 ± 2 Ma for the dacite indicates that least some komatites in the Shaw Dome area are younger than 2708 Ma, and are therefore younger than the ultramafic flows in the Kidd Creek footwall which are intercalated with 2716 Ma rhyolites. The U-Pb sample taken from approximately 2 km south of the Langmuir #2 mine is from an area with relatively poor exposure. It was taken to provide a minimum age on the monoclinal stratigraphic section that contains the Langmuir #1 and #2 deposits. The older age of 2724.4 ± 1.5 Ma indicates that a reverse or thrust fault is present in between this sample and the Langmuir #2 deposit.
Implications for Kidd Creek

The identification of a Stoughton–Roquemaure Assemblage tholeiitic substrate beneath Kidd Creek at the time of its formation bears on the genesis of the host high silica rhyolites. High silica rhyolites, also referred to as “tholeiitic” rhyolites, or, in the southern Superior Province, as FII rhyolites (Lesher et al. 1986), are distinguished from other rhyolite types principally by their geochemistry, and have more than 73% SiO₂, high incompatible element contents, Rb/Sr greater than 1, Zr/Y less than 5, La/Nb/YbN less than 3.5, negative Eu anomalies, and less than 2000 ppm F. Zircon saturation temperatures calculated using the equations of Watson and Harrison (1983) for the Mine Rhyolite samples are 840 to 890°C, and are believed to represent minimum emplacement temperatures (Barrie, 1995). Zircon saturation temperatures for Quartz Porphyry samples are on average lower by approximately 30°C. Given the following: 1) a high silica rhyolite composition; 2) relatively high zircon saturation temperatures; 3) spatial and temporal associations with tholeiitic mafic and ultramafic rocks; and 4) that rhyolites represent less than 5% of an extremely bimodal stratigraphic succession, the Mine Rhyolite and the Quartz Porphyry are believed to have formed by partial melting of a mafic to intermediate, tholeiitic substrate at pressures of 1 to 3 kilobars, well below pressures for the garnet stability field under dehydration melting conditions (Beard and Lofgren 1991).

The presence of a tholeiitic substrate at Kidd Creek is consistent with the conclusions in Barrie, Cathles, and Erendi (in press), and is in contrast to those of Huston et al. (1996) who considered that Kidd Creek felsic volcanic rocks are transitional S-type (derived from partial melting of a sedimentary substrate), based on the oxygen isotopic composition of quartz phenocrysts at Kidd Creek. It has since been demonstrated that the oxygen isotopic composition of Kidd Creek quartz phenocrysts have been reset (King et al. 1997).

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13. Project Unit 96–000. The Kidd–Munro Extension Project: Geology of the Fox–Stimson Area

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C/O Kidd–Munro Extension Project, Geological Survey of Canada

INTRODUCTION

This report presents results from an ongoing five year research project on the controls of base metal mineralization in the Abitibi greenstone belt funded by an industry–government partnership (Barrie 1996). In year one mapping, geochemistry and U-Pb geochronology were undertaken in the Fox–Stimson area to determine: 1) whether the Kidd–Munro assemblage, known for its VMS and nickel deposits, is present as a structural outlier in this area; 2) to determine the nature of the crustal substrate beneath the Kidd–Munro assemblage at the time of its deposition; and 3) to more clearly document the stratigraphic succession that is host to the Patton Zn-Cu occurrence in Stimson Township to the north. The Fox–Stimson area comprises a 12 township block, including, from north to south: Fox, Stimson, Sweatman, Pyne, Mortimer, Sherrington, Aurora, Edwards, Wesley, Calvert, Teefy and Rickard townships (Figure 13.1); 50 km east-northeast of the Kidd Creek mine. The area has poor exposure except for parts of Calvert, Teefy, Mortimer and Edwards townships to the southeast, and central Rickard Township to the southeast. Limited drilling has occurred in the area except for the Patton occurrence. A geologic map at a 1:50 000 scale has been produced based on the geology of the available surface exposures, airborne geophysical surveys, and 132 geochemical samples provided by the industry participants of the Kidd–Munro Extension Project (see Acknowledgements). The maps and a complete data syntheses will be published at the close of the Ontario portion of the project.

GEOLOGY

The Fox–Stimson area includes parts of the Kidd–Munro, and Stoughton–Roquemaure assemblages as defined by Jackson and Fyon (1991), and possibly the Duff–Coulson–Rand assemblage also. In the northern three townships, exposures are limited to half dozen outcrops, and much of the following description is based on airborne magnetic survey patterns and drilling by Noranda Mining and Exploration Inc. There is a much higher background magnetic signature to the rocks in the northern half of Fox and Stimson townships and the northern two thirds of Sweatman Township, with a distinct change in magnetic intensity trending northwest across Sweatman Township. This northern area is interpreted to be meta-granodiorite at lower amphibolite facies. Lower amphibolite metasedimentary and metavolcanic rocks are exposed in central Fox Township. A cherty oxide-sulphide facies iron formation up to 20 m thick is present, with probable facing directions to the north, and steep dips. The iron formation is within fine-grained hornblende-plagioclase amphibolite, biotite-plagioclase-quartz schist. Aplite-pegmatite granitoid rocks with fine-grained red-orange (andradite?) garnet locally are present at the surface also; such garnet-bearing granitoid apophyses are common in upper greenschist to amphibolite grade rocks in or near metasedimentary terrane. A garnet-in metamorphic isograd can be broadly traced from central Fox Township to the southeast corner of Sweatman Township, parallel to the boundary between amphibolite/wacke gneiss basalt, clastic sedimentary rocks (see Figure 13.1).

In Stimson Township, the strata are predominantly metasedimentary rocks and meta felsic tuffs. The metasedimentary rocks are mostly meta-turbidites and argillites, with abundant biotite, muscovite, and muscovite retrogressed after sillimanite. Garnet (almandine?) is common in strata that apparently underwent early hydrothermal alteration. It occurs in irregular patches and short intervals within garnet-poor or garnet-free metasedimentary and meta-tuffaceous rocks, and is manifested as 0.2 to 2 cm clots and aggregates with textures that indicate growth prior to peak metamorphic conditions. Sillimanite pseudomorphs overprint other textures indicating growth after peak metamorphism; however it is always retrogressed to muscovite. Fine-grained, Bouma-like turbidite layers have apparent reverse grading due to the preferential growth of coarse sillimanite in the clay-rich layers. Facing directions are apparently to the south and southwest in the Patton zinc-lead occurrence area based on this metamorphic reverse grading.

The central townships are also poorly exposed. Published government reports and maps (e.g., Jackson and Fyon 1991) indicate felsic intrusive rocks underlie most of Pyne and Aurora townships. Outcrops adjacent to and to the east of the granitoid pluton are highly strained, at mid-greenschist facies, and probably reflect the contact strain effects of granitoid emplacement. Outcrops in central Mortimer and Edwards townships, previously mapped as metasedimentary rocks, are now interpreted as variably strained and metasomatized mafic metavolcanic rocks. High-strain basalt and pillow basalt units are present at the surface, with foliations parallel to the margin of the granitoid; the few stretching mineral lincations measured have variable plunges from moderate to the north to steep to the south.

Outcrops in Mortimer Township have ubiquitous, slight to moderate epidote alteration. One outcrop in a gravel pit in central Mortimer Township has pervasive,
Figure 13.1. Geology of the Fox-Stimson area, with geochemistry and U-Pb zircon geochronology sample locations.
moderate to strong epidote alteration. Within 2 km, a Noranda drillhole intersected basalt, graphitic argillite within fragmental rocks, and magnesian gabbro, with all rocks having moderate chlorite or chlorite-epidote-carbonate alteration. The graphitic argillite contains sphalerite locally, with the highest concentration at 1.6% Zn/1 m (Noranda Mining and Exploration Inc., internal report, 1995).

Volcanic strata in Calvert, Teefy and Rickard townships (see Figure 13.1.) are predominantly thick, massive and pillowd basalt units that face south with steep dips to the north. The basaltic strata are continuous for kilometres along a predominantly east-west strike, and most are slightly to moderately overturned, but are disrupted by a north-trending fault in eastern Teefy Township. They represent a panel of south-facing rocks in the region greater than 5 km thick that extend for tens of kilometres along strike. Similar large scale structures with north over south vergence are documented by seismic surveys in the northeastern Abitibi in Quebec (Calvert et al. 1995).

In northeastern Calvert Township, strong airborne magnetic anomalies correspond to peridotite in one drillhole, and peridotite sills continue to the west and northwest. A second area with ultramafic rock is in central Calvert Township, where three drillholes intersected sheared ultramafic rocks intercalated with graphitic, sulphide argillite with minor Zn mineralization, and minor dacite. This succession is similar to, but thinner than, that above the Dundonald sill in Dundonald Township 25 km to the southwest. To the northwest and southeast, up- and down-section, are massive and thick sequences of pillowd and massive calc-alkalic(?) basalts.

The stratigraphic succession in Rickard Township principally comprises southeast and south-facing calc-alkalic(?) basalts, with lesser dacites and interflow graphitic argillites. Exposures in central Rickard Township (near the Twindyke gold mine area: see Simony 1963) indicate bedding in massive and pillowd basalts is sub-vertical or steeply north-dipping and overturned. A north-northwest-trending fault is inferred between the rocks of central Rickard and Teefy townships based on airborne magnetic patterns and bedding attitudes.

**GEOCHRONOLOGY**

One U-Pb zircon age was determined from a rhyolite breccia with transitional calc-alkalic-high silica rhyolite geochemical affinities that is host to the Zn-Cu mineralization at the Patton occurrence in Stimson Township. The age is 2726 ± 3 Ma, based on a regression of 4 analyses. This is believed to approximate the age of the volcanogenic massive sulphide-style mineralization there, and also provides an age for the Scapa Assemblage in this region. This is reported in more detail in Barric and Corfu (this volume).

**Galena Pb Isotopic Study**

Thorpe (1990; and in press) determined that the lead isotopic signature of galena from most VMS deposits in the Abitibi subprovince and parts of the Wawa subprovince share a common source. The Abitibi–Wawa galena lead isotope model can be used for model age determinations with some confidence for reconnaissance stratiform galena. With the Abitibi Pb model in mind, two samples were taken from the Patton occurrence. Sample 96TB0050a is from stratified sulfide whereas sample 96TT0050c is from a crosscutting galena-tremolite vein. The analysis for 50a yields an identical value as the one grain measured for 50c, whereas two other grains for 50c

**Stoughton-Rouquemaure Assemblage, Calvert-Teefy-Rickard Townships**

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**Figure 13.2.** Stratigraphic column for the Calvert–Teefy township area. The strata are principally Stoughton–Rouquemaure tholeiitic basalts represented by samples 96TB064-72; see Figure 13.1 for sample location, and Figures 13.4 and 13.5 for geochemistry.
yield somewhat more radiogenic compositions. All the values deviate significantly from the Abitibi Pb-Pb linear array defined by Thorpe, and are believed to reflect radiogenic enrichment during a secondary remobilization event, sometime in the Late Archean or in the Proterozoic. Using the parameters in the Abitibi–Wawa Pb isotope linear (Thorpe, in press), model ages for these lead compositions range from 1922 Ma to 2002 Ma (values of 8.89; values of 3.65 to 3.85).

GEOCHEMISTRY

In the Fox–Stinson area, 23 less-altered samples were selected for research grade whole rock geochemistry at the Geological Survey of Canada Laboratories. These samples are located in Figure 13.1, and are presented in Figures 13.4 and 13.5. These are complemented by 132 in-kind geochemical samples provided by company participants of the Kidd–Munro Extension Project (see Acknowledgements).

A Jensen cation plot and a rare earth element ratio plot (see Figure 13.4) provide a summary of the major and trace element characteristics of rock types sampled. A thick, monoclinal (locally overturned) sequence of tholeiitic basalts that predominate in the southern half of the map area comprise part of the Stoughton–Roquemaure assemblage. Through Calvert, Teefy, Edwards and Rickard townships, they comprise a series of alternating high magnesium and high iron tholeiites. In Mortimer Township, the basalts are generally more magnesian. The higher iron varieties have higher rare earth element (REE) patterns and other incompatible trace elements (see Figure 13.5), suggesting a derivation from their more magnesian counterparts by tholeiitic (i.e., clinopyroxene-plagioclase) fractionation. These basalts have low (La/Sr) primitive mantle (PM) and (Gd/Yb)PM ratios indicating flat to slightly light rare earth element (LREE)-depleted REE patterns (see Figure 13.5). The REE and primitive mantle-normalized plots highlight the flat, slightly light rare earth element-depleted tholeiitic patterns. The primitive mantle-normalized patterns (see Figure 13.5) show slight high field strength element (HFSE) depletion consistent with the slight LREE depletion, no significant Ti, Nb or Ta anomaly, and suggests that the flood basalt-like tholeiites of the southern half of the map area were derived from a part of the mantle that had undergone a previous melting event.

The Scapa assemblage rocks in Stimson Township have a minor felsic volcanic component. The Patton felsic rocks are calc-alkaline rhyolites with a transitional affinity to “tholeiitic” or high silica rhyolites. The one exception is the sample taken from western Stimson Township, which is an adakite. Adakites are primitive calc-alkaline rocks, usually with abundant plagioclase, that are known to form in arcs by partial melting of a subducted slab of oceanic basalt within the garnet stability field. In general they are low temperature melts that are not associated with VMS deposits (Barrie, Ludden and Green 1993; Barrie 1995).

MINERAL DEPOSITS

In Rickard Township, volcanogenic massive sulphide-style Zn-Pb mineralization is present in interflow argillites within brecciated, hyaloclastics and moderately silicified basalt (see Figure 13.1). The best intersection reported is 3.2% Zn and 0.6% Pb over a core length of 3.2 m (Carmichael 1992: Kirkland Lake assessment files). Approximately 2 km north-northwest of this occurrence is the Twindyke Au mine (Simony 1963). The Twindyke mine is within east-west subvertical carbonate-quartz-sulphide veins in highly altered and sheared mafic volcanic rocks; altered and slightly mineralized felsic porphyry apophyses are found within the sheared rocks nearby. Gold values up to 1.2 oz./1.6m have been reported from vertical drillholes near the Twindyke shaft (Rio Tinto 1958: Kirkland Lake assessment files).

The Patton Occurrence

The Patton Zn-Pb occurrence was outlined by a ground magnetic and electromagnetic surveys in central Stimson Township conducted by Cross Lake Minerals Ltd. in 1989 and 1990. The stratigraphic succession (see Figure 13.3) contains up to 50% granitic dikes, gneissic granitic screens and diabase dikes that occlude the volcanic-sedimentary strata related to mineralization. The mineralization consists of semi-conformable sulphides up to 20% within meta-graphitic argillite and coarse fragmental felsic tuff, with minor granite and granitic pegmatite dikes cutting the strata. The highest grade intersection reported is 3.8% Zn and 2.5 Pb/6.8m, within altered coarse felsic tuff. Massive or gneissic granitic rocks are intercalated with biotite sericite plagioclase quartz graphite schist, greywacke, or felsic tuffs. The stratigraphic succession apparently faces to the south and southwest based on graded beds in turbiditic greywackes.

DISCUSSION: COMPARISON OF STOUGHTON–ROQUEMAURE ASSEMBLAGE AND FLOOD BASALT PROVINCES

In several ways the thick sequence of Stoughton–Roquemaure tholeiitic basalts in the southern part of the map area is comparable to flood basalts found in continental settings (e.g., Columbia River, Noril’sk), at passive rift margins (Parana, Entendeka, Red Sea) and in oceanic plateaus (Ontong Java, Kerguelen, Iceland: Mahoney and Coffin, 1997). Such flood basalts are characterized by a large volume (10^5 to 10^7 km^3) of predominantly tholeiitic basalt deposited over a relatively short period (10^6 to 10^7 years). Flood basin provinces commonly have 5 to 15 km thick stratigraphic sections, with mildly alkaline, light rare earth element-enriched basalts and picrites at or near the base, overlain by thick sequences of tholeiites characterized by variable but relatively flat REE and primitive mantle-normalized patterns.

Flood basalt volcanism and large igneous provinces are associated with lithospheric spreading and rifting.
above upwelling mantle diapirs within which extensive partial melting occurs. In the classic mantle plume theory that describes the genesis of flood basalt magmatism (e.g., Campbell and Griffiths, 1992), a mantle plume arises from the core-mantle boundary or the 670 km discontinuity, and has a large, mushroom-shaped plume head that entrains upper mantle material, and a hot, relatively narrow central conduit through which higher temperature melts rise from the source region. This theory has been used to account for komatites in the hotter Archean mantle, which are attributed to the hot plume tail (e.g., Campbell, Griffiths and Hill 1989). Thus classic mantle plume theory accounts for many of the features of both the Stoughton–Roquemaure assemblage and the komatites within the overlying Kidd-Munro assemblage. However, several geological features of the Stoughton–Roquemaure assemblage are inconsistent with their generation in mantle plumes. The erratic nature of the geochemistry of the Stoughton–Roquemaure basalts across section is dissimilar to the major flood basalts provinces, which generally have thick, monotonous sections of compositionally identical basalt, and smooth chemical variations across section. The presence of intercalated calc-alkaline dacites cannot be explained by mantle partial melting. It is noted however, that in the Cretaceous Columbian–Caribbean Igneous Province which is the only Cenozoic flood basalt province that is komatiite-bearing (e.g., Gorgona Island) contains calc-alkaline dacites and rhyolites in the Medellin area of Columbia (Kerr et al. 1997). The tectonic history of this province which has been structurally dismembered, locally accreted and otherwise scattered about the Caribbean and eastern Pacific may bear on the tectonics of the southern Abitibi.

**Figure 13.3.** Stratigraphic column for the Patton Zn-Cu occurrence area in Stinson Township. The strata are within the Scapa assemblage and are along strike with oxide-sulphide facies iron formation-bearing metasedimentary rocks at upper greenschist to lower amphibolite facies; see Figure 13.1.

**Figure 13.4.** Jensen plot and rare earth element ratio plot for samples in the Fox–Stinson area. Samples are divided according to chemical affinity and location. TH-1: primitive tholeiitic basalts with slight LREE depletion.
Figure 13.5. Chondrite-normalized rare earth element and primitive mantle-normalized plots for samples in the Fox–Stimson area. Samples are divided according to chemical affinity and location; see Figure 13.4 for divisions.
SUMMARY

1. The Fox-Stimson map area is predominantly underlain by a thick sequence of flood basalt-like tholeiites of the 2.72 Ga Stoughton-Roquemaure assemblage. The association of a thick sequence of tholeiitic basalts overlain by the komatiite-bearing 2.717 to 2.710 Ga Kidd–Munro assemblage invites comparisons with magmas derived in mantle plumes, such as the Cretaceous Columbian–Caribbean Igneous Province.

2. A U-Pb age for a Scapa assemblage transitional rhyolite that is host to the Patton Zn-Cu occurrence in Stimson Township is 2726 ± 4 Ma, the same age within error as the Lamarche Township felsic volcanic unit broadly along strike 30 km to the west (see Barrie and Corfu, this volume).

3. An upper greenschist garnet-in metamorphic isograd is present from central Fox Township to the southeast corner of Sweatman Township.

4. The Kidd–Munro assemblage is absent in this area, and there is little evidence for Duff–Coulson–Rand calc-alkalic rocks in the area.

ACKNOWLEDGMENTS

The Kidd–Munro Extension Project is designed to develop a detailed stratigraphic, structural, geochemical and geochronological framework to assist in the exploration for large VMS and Ni deposits in the 2.72 to 2.71 Ga ultramafic-rich terrane of the southern Abitibi subprovince. The following individuals and organizations are gratefully acknowledged for their participation in, and support of, the Kidd–Munro Extension Project: Paul Davis, Jarro Vesanto and Outokumpu Metals and Resources Canada; Ray Band, Scott McLean, and Falconbridge Exploration Limited; Tom Lane and Teck Exploration Limited; Michelle Mainville, Gino Roger, Jules Riopel, and Noranda Exploration Co., Ltd.; Tom Hart and Inco Exploration and Technical Services Inc.; the Timmins and Kirkland Lake Resident Geologist offices of the Ministry of Northern Development and Mines; John Ayer, Ben Berger, Phil Thurston, and the Ontario Geological Survey; Yuri Amelin and the Jack Satterley Geochronology Laboratory of the Royal Ontario Museum; Beth Hillary, Mark Hannington, Charlie Jefferson, and the Mineral Resources Division of the Geological Survey of Canada. Darryll Last provided excellent drafting for the first two years for this project. This is GSC contribution #1998114.

REFERENCES


14. Special Project. Ore Deposit Descriptions of the Timmins Area

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The Timmins District has had a long history of continuous mineral production from over 60 mines in the past 90 years. Excellent descriptions for many of the deposits are available in the literature. However, the scope of the Ore Deposit Descriptions project is to provide a description of the nature and setting of the mineralization from selected

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**Figure 14.1.** Location map of deposits included in the special project
deposits for which little information is available, with an emphasis on the economic aspects of the mineralization. The target audience is principally geologists and prospectors who are engaged in the search for new mineral deposits with characteristics similar to Timmins-style deposits, however those geologists and researchers in academia, various government bureaus, and mines should also find the content to be of interest. The project is structured to provide an overview only, as space, time, and information limitations preclude development of a full and in-depth description of each deposit. The intent is to stimulate thought and discussion, and to highlight areas requiring more research. The format of the project will initially be in the Open File Report format with a target release date of March, 1999, however it is intended that a full-colour digital version of the report will be released at a later date.

The report is structured such that each deposit is described using the same format as much as possible. The deposits are described on the basis of their location, ownership, and claims, regional geological setting, property scale geological setting, structural geology, and economic geology. Macroscopic description of the deposits include principle rock types, alteration and mineral assemblages, structural features and tabulated litho-geochemical data. Sections describing the production history, amount of development and diamond drilling to-date, a brief summary of the mining methods employed, and a detailed description of the various orebodies encountered in the mine are included in the economic geology section. The location of the deposits are shown in their regional geological context in Figure 14.1.

During the course of completing the description of Kinross Gold Corporation's Hoyle Pond Mine, it became apparent that very little information was available describing the alteration style and element associations of the 1060 Zone quartz veins. In order to address this gap in information, a separate study was commissioned through a co-operative effort by Kinross Gold Corporation and the author. This study will incorporate quantitative and petrographic data to characterize the alteration minerals associated with the 1060 Zone style of mineralization, and will be released separately.

The preliminary conclusions of the Ore Deposits Description report result from observations of the gold deposits located in southern Hoyle Township along what Kinross Gold Corporation has termed their "New Mine Trends". Here, ample evidence is present to document a history of polyphase folding and multiple generations of gold mineralization. These observations, along with the nature of the host lithologies lends support to the widely held premise that the rocks of the "New Mines Trends" are the folded northern equivalents of the Tisdale Group as defined by Pyke (1982). Should this view be correct, this raises the possibility that the other bands of mafic-ultramafic rocks located further north (e.g., central Murphy Township, and Gowan–Matheson townships) also are the folded, time-stratigraphic equivalents of the Tisdale Group units. Indeed, the presence of several known gold occurrences (the Nickel Offsets Deposit: 370 650 tons grading 0.174 ounces per ton gold (Murray and Pickens, 1996), the Texmont Deposit: 114 000 tons grading 0.22 ounces per ton gold (Pearson, 1989), and the Frankfield Deposit: 310 000 tons grading 0.21 ounces per ton gold (Pearson, 1989), although hosted by the older Kidd–Munro assemblage, demonstrate that gold mineralization is present well to the north of the traditionally recognized favourable stratigraphy. This has significant implications for exploration because it identifies a number of new, essentially untested areas as having high potential for hosting economic quantities of gold mineralization.

A second conclusion from the project is that the gold mineralization at the Detour Lake Mine is principally related to pyrrhotite-chalcopyrite-pyrite veins and stringers that cross-cut all host rock types, quartz veins and potassic (biotite) alteration zones. Although native gold hosted within quartz veins is observed on occasion, it's contribution to the mine's production is minimal (R. Hill, 1998, personal communication). The Detour Lake Mine is clearly an atypical example of the style of gold mineralization in the region, and as such the controls on, and genesis of the deposit remains enigmatic. Further work will be required to document in detail the alteration assemblages, mineral paragenesis, deformation history, and distribution of gold in an effort to understand and explain the genesis of the deposit. This understanding is required in order to effectively target future exploration programs elsewhere in the northwestern part of the Abitibi subprovince.

REFERENCES


15. Project Unit 94–096. Detailed Geology of the Northern Eagle Property Barite Occurrence, Hemlo Area

T. L. Muir
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INTRODUCTION

The discovery of the main mineralized zone of the Hemlo gold deposit in 1981 was tied, in part, to an exploration model that invoked a syngenic origin for the deposit. The observed association of barite with the deposit and the presence of apparently bedded barite deposits, such as at the Northern Eagle property (Patterson 1984), about 21 km west of the Hemlo gold deposit, appeared to be consistent with the syngenic model. Sulphur isotope studies of barite and pyrite within the Hemlo gold deposit and within barite showings to the west of the deposit were consistent with a syngenic origin, although it was acknowledged that the results could not distinguish between a syngenic and an epigenetic hydrothermal origin (Cameron and Hattori 1985; Hattori and Cameron 1987). Presently, although there is no consensus as to the origin of the Hemlo gold deposit, more favoured scenarios centre around structural control involving magmatic hydrothermal systems, possibly including porphyry intrusions. Nonetheless, many geologists associated with studies of the Hemlo gold deposit presently seem to favour a syngenic origin for the barite.

When examined in the field though, there are several aspects which are consistent with an epigenetic origin for the barite. No primary bedding features are found in the metasedimentary and barite units of the Hemlo gold deposit, where strain and recrystallization are notable. Barite in the deposit varies from massive to layered and may be wispy to disseminated, or form veins in, or the matrix to, brecciated ore (Kuhns et al. 1986; Walford et al. 1986). Barium is also present in many of the ganne minerals, including microcline and green muscovite (Harris 1989). Harris (1989) concluded that barite in the Hemlo gold deposit formed later than the ore minerals but was part of an evolving hydrothermal system. At the Northern Eagle property, Gliddon (1985) inferred that barite was sedimentary in origin although he noted a spatial association of barite with shear zones. Muir (1991) noted isoclinal folds as well as transposition and differentiation of layering and concluded that the layering ought not to be assumed to be bedding. Additional stripping and hydraulic cleaning of the area during 1995 improved the exposure and revealed several features which clarified or brought to light important geological relationships. With the importance of tying the geology of the Hemlo gold deposit to the Hemlo greenstone belt (e.g., Muir 1997, Jackson 1998, Beakhouse 1998, Davis et al. 1998), detailed mapping of the exposed area was undertaken this field season. All observations presented here are based on this work.

GENERAL SETTING

The Northern Eagle property is located about 600 m north of Highway 17, about 1 km west of the Black River, 14 km east of Marathon (Figure 15.1). The outcrop area mapped this field season (Figure 15.2) is centred at the approximate UTM co-ordinates of 558850E and 5393050N. The baritic units occur near an east-striking contact between mafic metavolcanic (?) rocks and metasedimentary rocks. All units are moderately to steeply south dipping. Footwall metasedimentary rocks consist of turbiditic units, locally with rip-up clasts and/or graded crystal detritus. The units are overturned and face north, based on inferred normal grading, as deduced by Gliddon (1985). Baritic rocks form tabular bodies within the structurally lower quarter of a 20 m thick zone of variably altered rocks, which locally display relict sedimentary characteristics. This zone is termed here, for convenience, the Northern Eagle alteration zone. The baritic rocks range from massive to weakly and thinly layered, although no primary sedimentary connotation is implied. The main unit appears to terminate at the eastern end of the exposed area, although altered rocks continue to the east. A swarm of feldspar porphyry dikes has intruded altered units to the south of the baritic rocks. Strain is moderate to intense and appears to increase, from the footwall, towards the zone of altered rocks. Hanging wall units are insufficiently exposed to evaluate strain intensity relationships.

LITHOLOGIC UNITS

There are several main types of rocks exposed at the Northern Eagle property showing. An attempt was made to depict those variations that would best illustrate spatial relationships among units, based on primary and secondary features (strain, alteration), particularly in a temporal context. Much of the western third of the exposed area is heavily oxidized, thereby reducing the clarity of distinction among various units. Gliddon (1985) provided more detailed descriptions of many of the units, although many of the units are defined and interpreted somewhat differently than herein. For the following descriptions, refer to Figures 15.3a, 15.3b and 15.3c.
Unaltered to Weakly Altered Rocks

MAFIC ROCKS (UNIT 1)

Fine-grained, dark green amphibolite (unit 1a) with a weakly developed gneissosity and locally developed uncharacterized alteration occur in the hanging wall to the Northern Eagle alteration zone. No primary volcanic features were observed. At the eastern end of the exposed area, a narrow zone of lighter green, altered, foliated, mafic rock (unit 1b) forms the southernmost limit to the Northern Eagle alteration zone.

METASEDIMENTARY ROCKS (UNITS 2 AND 3)

Two general types of sedimentary rocks are present, namely fine-grained turbidites displaying wacke-siltstone couplets (unit 2) and black, siliceous mudstone (unit 3). The turbidites have been subdivided into less-strained rocks (unit 2a), in which rip-up clasts and grading of feldspar and quartz crystals are well defined; and more

Figure 15.1. General location of the Northern Eagle property in the Hemlo greenstone belt.
Figure 15.2. Simplified geology of the Northern Eagle property barite showing based on Figure 15.3.
a) NORTHERN EAGLE PROPERTY BARITE OCCURRENCE (WEST PART)

Dikes
- 8a Aphanitic felsic dike
- 8b Felsic, foliated, plagioclase porphyry dike
- 8c Aphanitic intermediate dike
- 8d Foliated biotite lamprophyre dike

Intrusive Contact

Altered Rocks and Emplaced Products*
- 7a Barite-rich rock
- 7b Barite-pyrite-rich rock
- 7c Barite-rich rock ± siliceous and/or feldspatic alteration
- 7d Siliceous-feldspatic rock ± barite
- 7e Feldspatic-siliceous rock ± barite

- 6a Cleaved, green-mica-bearing phyllitic schist (derived, in part, from unit 3a)
- 6b Cleaved, lenticular or fragmental, green-mica-bearing phyllitic schist (derived, in part, from unit 3d)
- 6c Cleaved, green-mica-pyrite-bearing phyllitic schist (likely derived from units 2, 3, 4, and/or 5)

- 5a Transposed, cleaved ± laminated feldspatic rock
- 5b Transposed, cleaved, lenticular (fragmental?) feldspatic rock

- 4a Cleaved, layered, partly altered (bleached, pyrite) sedimentary rock
- 4b Cleaved, layered and laminated phyllitic schist (derived from units 2a, 2b)
- 4c Cleaved, laminated phyllitic schist (feldspathized, pyritized) (likely derived, in part, from unit 2b)

*Refers to probable dilational infilling by barite etc.

(Legend continues on Figure 3B)

Figure 15.3. a) Detailed geology of the Northern Eagle property barite showing, western part.
b) NORTHERN EAGLE PROPERTY BARITE OCCURRENCE (CENTRAL PART)

Unaltered to Weakly Altered Rocks

3a Cleaved siliceous black mudstone
3b Cleaved siliceous black mudstone with pyrite nodules
3c Cleaved and laminated siliceous black mudstone
3d Cleaved, siliceous black mudstone with coarse clasts (fragmental)
2a Turbiditic fine sandstone-siltstone couplets ± grading ± rip-up clasts
2b Layered ± cleaved, fine sandstone-siltstone sedimentary rock (likely turbidites)
2c Coarse clastic sedimentary rock with intermediate to mafic matrix
1a Gneissic amphibolite
1b Altered, foliated mafic rock (derived from unit 1a)

(Legend continues on Figure 3C) Centre of outcrop area approximately 558850 E, 5393050 N

Figure 15.3. b) Detailed geology of the Northern Eagle property barite showing, central part.
Figure 15.3. c) Detailed geology of the Northern Eagle property barite showing, eastern part.
highly strained, weakly to moderately cleaved rocks (unit 2b), which display compositional layering similar to the less strained turbidites, but with no recognizable grading. In unit 2b, rare, compositionally distinct lenses may represent rip-up clasts. Both subunits locally display particularly dark layers (charcoal grey), compared to the predominant medium to light grey colour. As across-strike exposure is not continuous, it is inferred that a change from turbidites to inferred turbidites occurs over less than 3 m (e.g., from the inset in Figure 15.3b to the main view). The change in characteristics of the metasedimentary rocks is attributed largely to strain (see section on Structural Geology). Subunit 2b forms the immediate footwall to the Northern Eagle alteration zone and the hanging wall unit to the gneissic amphibolite, adjacent to the hanging wall of the Northern Eagle alteration zone. A third, volumetrically very minor, enigmatic unit (2c) consists of aphanitic grey clasts within an intermediate to mafic matrix. Similar units were observed in several localities within stripped trenches on the Golden Sceptre (Battle Mountain) property, west of the Hemlo gold deposit (Muir 1997).

The mudstone unit is subdivided into:

a) Cleaved, siliceous black mudstone with no detectable layering (unit 3a);

b) Similar mudstone containing 0.5 to 2 cm diameter pyrite nodules (unit 3b);

c) Cleaved and variably transposed, laminated to layered, siliceous, black to very dark grey mudstone (unit 3c);

d) Cleaved siliceous black mudstone with variably strained, locally angular, light-coloured lenses interpreted to be primary clasts (unit 3d).

The mudstone units occur structurally below the mafic rocks, and within the Northern Eagle alteration zone, almost exclusively spatially associated with the swarm of feldspar porphyry dikes. The layered/laminated mudstone occurs structurally above altered feldspathic rocks (unit 4). Many of the fragments or clasts appear to be primary, however, some of the "lenses" are secondary in character and appear to be the result of incipient, uncharacterized alteration that has occurred along fractures. These fractures have subsequently been variably strained. The pyrite nodules range from spherical to slightly ellipsoidal. Many are zoned with a massive fine-grained core, followed by one to several, well- to poorly defined, concentric "layers" of fine- to medium-grained radiating pyrite crystals. Locally, some of these nodules have partly coalesced, producing what appears to be a botryoidal texture.

**Variably Altered Rocks**

The remainder of the supracrustal rocks has been subdivided into 5 types, some of which have a reasonably evident protolith.

**FELDSPATHIC ROCKS (UNITS 4 AND 5)**

Three subdivisions of unit 4 include a spectrum of rocks as follows. A zone of cleaved rocks (unit 4a) that have undergone bleaching with the introduction of pyrite lies adjacent to unaltered sedimentary rocks (unit 2b) in the footwall to the barite mineralization. The zone is gradational with the underlying unit 2b over a thickness of 30 to 80 cm and appears to be continuous within the exposed area.

Within the eastern half of the map, structurally above the barite mineralization, is a unit of apparently feldsparitized, cleaved, layered and laminated phyllic schist (unit 4b), in which remnant zones or lenses of metasedimentary rocks (unit 2b) occur. This suggests that the protolith for unit 4b is at least, in part, unit 2b. Some layers display numerous lenses, which may represent rip-up clasts. The unit typically displays a weakly to moderately rusty weathering surface. Unit 4b appears to pinch out to the west.

The third subdivision (unit 4c) consists of more intensely but variably feldsparitized and pyritized, variably laminated phyllic schist with well-developed spaced cleavage (3 to 10 mm apart). This unit essentially surrounds the barite mineralization. Where observed, unit 4c appears to grade into unit 2b, suggesting that the rocks also may have been derived from metasedimentary rocks. Cleavages anastomose in 2 dimensions and commonly contain pyrite from hairline thick to 2 mm thick. The weathered surface is locally heavily oxidized, particularly in the western part, thereby precluding a more precise definition and estimate of the protolith. Locally, massive pyrite occurs in ill-defined zones up to 2.5 cm thick. Pyrite crystals range from equigranular and fine grained, to coarse grained. The crystals are not preferentially oriented.

Unit 5 is enigmatic and a minor constituent of the outcrop area. Both a disrupted laminated variety (unit 5a) and a lenticular variety (unit 5b) are well cleaved and appear to consist of transposed rocks. The lenticular type may represent a primary fragmental unit, as various textures were noted amongst the lenses. Both types are grey on weathered surfaces, apparently devoid of pyrite, and appear to be feldsparitized.

**GREEN-MUSCOVITE-BEARING ROCKS (UNIT 6)**

Of the 3 subdivisions of green-muscovite-bearing rocks, 2 appear to be derived from types of black mudstone (unit 3), based on spatial association, textures, and abruptly gradational contacts. A non-laminated, well-cleaved phyllic schist (unit 6a) is spatially associated with non-laminated black mudstone (unit 3a), and a well-cleaved, lenticular to fragmental appearing variety (depending on degree of strain) (unit 6b) is spatially associated with fragmental mudstone (unit 3d). These green-muscovite-bearing rocks generally appear to be more highly strained (flattened) than the black mudstone, possibly because the type and degree of alteration facilitated preferential incorporation of strain in the muscovitic altered products. Lenses in the green-muscovite-bearing rocks, where present, typically have aspect ratios ranging from 5:1 to 20:1, depending on the localization of strain (e.g., relative to competent dikes). Locally, there are numerous, uniden-
tified, nearly equidimensional, very dark green to black crystals (<1 mm) with bright green "pressure shadows".

A third subdivision consists of pyritiferous green-muscovite-bearing phyllitic schist (unit 6c) which tends to be spatially associated with feldspatized rocks (unit 4c), particularly in the western half of the map area. This rock is also well cleaved and strained. The presence of pyrite has resulted in a locally heavily oxidized surface (west half). In the eastern end of the exposed area, this unit is less oxidized and appears to be similar in character to unit 4c other than the presence of green muscovite and pyrite. The protolith(s) for this unit is (are) unclear, but may include units 2, 3, 4 and/or 5.

**Highly Altered Rocks or Emplaced Products**

**BARITIC, SILICIC AND FELDSPATHIC ROCKS (UNIT 7)**

Most of the baritic, silicic and feldspathic rocks occupy a continuous tabular zone, ranging from about 1 to 3.5 m thick, near the structural footwall of the Northern Eagle alteration zone. Some of the rocks appear to represent replacement of pre-existing rocks, whereas others may represent material, such as barite, quartz and feldspar, that was hydrothermally introduced into dilatant structures, hence the term "emplaced products". Barite-rich rocks (estimated barite content >50%; units 7a, 7b) were generally clearly distinguishable from other rocks by field-estimated density. These rocks are variably pyritiferous, oxidized on weathered surfaces, white on fresh surfaces, and have a fine-grained sugary texture. Pyrite may be very fine-grained and disseminated (<1% to 2%; unit 7a) or about 1 mm in size (5 to 25%), and partly disseminated, or may define a weakly developed layering (unit 7b).

Two narrow, dike- or vein-like barite units occur structurally above the main baritic zone, surrounded by cleaved, feldspathic, pyritiferous rocks (units 4c) (Figure 15.3b). One unit appears to terminate by pinching out within unit 4c (Figure 15.3a). The two, barite units are separated by green-muscovite-bearing phyllitic schist (unit 6).

Unit 7c consists of rocks that are less dense and spatially associated with, and similar in appearance to, unit 7a. These white, fine-grained, sugary textured rocks have an estimated 10 to 50% barite. Units 7d and 7e consist of white rocks, which contain an estimated 0 to 10% barite. Locally, unit 7c displays thick (2 to 6 cm), discontinuous white "layers" of barite-rich material, separated by cleaved, recessively weathering material; thin (<1 cm) layering defined by fine-grained, grey to white "layers" and granular rusty weathering "layers"; and grey irregularly shaped domains and lenses separated by less than 2 cm thick, irregularly shaped, vein-like white material (barite?). The latter form is reminiscent of fracture-controlled or "stockwork"-like alteration in the North Zone of the Hemlo gold deposit.

Siliceous and feldspathic rocks occur on the fringes of, or within, the main baritic zone (combined units 7a, b, c). Siliceous rocks (unit 7d), likely with some feldspar and possibly locally barite, form 2 lenses. One lens lies near the hanging wall contact of the main baritic unit in the central part of Figure 15.3b. It is white weathering, displays a sugary texture, is devoid of pyrite, and is cleaved to coarsely layered. The other, larger lens lies at the footwall contact of the main baritic zone, straddling Figures 15.3a and 15.3b. This lens consists of white-weathering, sugary-textured, layered material, with 1 to 2 mm thick "seams" of pyrite separating the 1 to 5 cm thick white "layers". The layering ranges from poorly defined and relatively discontinuous to contorted.

Predominantly feldspathic rocks (unit 7e), with or without quartz and, possibly, locally barite, occur as thin ill-defined lenses within and at the hanging wall contact of the main baritic zone (Figure 15.3b). These are fine-to medium-grained granular rocks, which differentially weather to define some form of weakly developed layering, partly evident as differences in the degree of oxidation.

**Intrusive Rocks**

**DIKES (UNIT 8)**

Four types of dikes were recognized within the outcrop area. The most common type, felsic, feldspar (plagioclase) porphyry (unit 8b), forms a swarm of about 10 dikes that intruded into a zone at least 20 m wide (based on exposure limits). These dikes are generally uniform in texture, contain numerous (35%) medium-grained phenocrysts, and are foliated. Inclusions are common, particularly in the thicker dikes, although some of the thinner dikes are inclusion-free. Inclusions consist of unaltered, foliated, sedimentary wall rock, altered, cleaved, green-muscovite-bearing rocks, and exotic (magmatically related?) material of intermediate composition with rare, fine-grained feldspar crystals (phenocrysts?). Glidson (1985) has also previously reported altered inclusions for the Northern Eagle property occurrence. All inclusions are aligned parallel to the predominant fabric and dike contacts, and have aspect ratios ranging from 5:1 to 10:1. Cleavage in the inclusions, where present, is parallel to the maximum dimension of the inclusions.

Other dikes include an aphanitic, aphyric, felsic dike (unit 8a), an aphanitic, aphyric, intermediate dike (unit 8c), and a couple of well-foliated biotite lamprophyre dikes (unit 8d).

**QUARTZ VEINS**

Quartz veins occur mostly as northeast to east-northeast-striking veins, and locally east-striking, narrow, gash fracture infillings, irregular to complexly shaped bodies, and small, oval or lenticular pods. The most continuous veins occur in the footwall sedimentary rocks, whereas the pods and irregularly shaped masses tend to occur in the feldspathic units (4b, 4c). The pods are also spatially associated with minor "Z"-shaped folds, small-scale,
northeast-striking dextral slips, and boudin necks in dikes. The feldspar porphyry dikes locally contain elongate veins or pods. No quartz veins occur within the baritic units.

**STRUCTURAL GEOLOGY**

**Folds**

Isoclinal, east- and west-closing, “U”-shaped folds are present in the footwall and hanging wall metasedimentary rocks of the Northern Eagle alteration zone (see Figures 15.3a, 15.3c, 15.4a, 15.4e), and open “S”- and “M”-shaped folds are present in the cleaved, altered “feldspathic” rocks within the zone (Figures 15.3b, 15.3c, 15.4g). Many of the folds within unit 3c are disrupted by various stages of transposition of layering (Figure 15.4b, 15.4f) or by dislocations along the predominant cleavage, which also affect layering within unit 2b (Figure 15.4e). Locally the abruptly gradational contacts between units 3a and 6a appear to be folded (“S”-shaped) or to have replaced pre-existing folds. Some folded and transposed layering was too complicated to sketch, but nonetheless indicated the considerable strain recorded within mudstone units 3a and 3d. Plunge of fold axes could not be determined from the outcrop surfaces.

Locally, small “Z”-shaped folds, which have deformed the predominant planar fabric, have developed commonly in spatial association with quartz veins. In some cases, these folds have an associated axial planar cleavage oriented counterclockwise to the main trend of the layering in the outcrop. Plunges of fold axes could not be determined from the outcrop surfaces.

A sinistral kink fold with a northwest-striking fold envelope is present in the western part of the map area (Figure 15.3a). The fold envelope is slightly curved and terminates in the exposed area towards the southeast. The fold affects previously folded layering, cleavages, quartz veins and dikes.

**Planar Structural Elements**

External to the Northern Eagle alteration zone the predominant fabric is essentially parallel to the south-dipping bedding and consists of a preferred dimensional alignment of, most commonly, biotite, as well as an alignment of flattened clasts (rip-ups) and crystal detritus. Towards the Northern Eagle alteration zone (footwall), particularly within about 3 m, a roughly spaced, layer-parallel (except in fold noses) cleavage, has developed. Within biotite-rich layers in the hanging wall and footwall metasediments there is also a north-dipping, penetrative, preferred dimensional alignment of biotite, oriented counterclockwise to the layering and predominant cleavage (see Figures 15.3c, 15.4a, 15.4e).

A pronounced, layer- and laminae-parallel, spaced cleavage (3 to 10 mm) is present within many of the rocks of the Northern Eagle alteration zone. There are some indications of 2 “sets” of anastomosing cleavages, which have strikes differing by less than 4°. In the pyritiferous subunits, the pyrite was commonly distributed along the cleavage planes. Some of the green-muscovite-bearing rocks, with or without a lenticular character, display mylonitic fabrics on a hand lens scale. In such cases, highly flattened, grey (quartz?), white (feldspar?), and green (muscovite) lenses, with greater than 10:1 aspect ratios, occur on a submillimetre scale. Locally developed boudinage of laminations, layers and cleavages is evident within the more feldspathic rocks as are low-angle to layering/cleavage faults.

**Linear Structural Elements**

A well-developed elongation and intersection lineation is evident on dip surfaces of cleavage planes, particularly in the phyllitic schist units. The lineation plunges moderately to the west-southwest (see Figures 15.3b, 15.3c) and is best defined by stretched clasts, and groups of biotite or green muscovite crystals. It appears to be parallel to the intersection of the 2 “sets” of anastomosing cleavages where present and/or the intersection of a cleavage with layering.

**Faults**

Several north-northwest-striking faults with dextral sense displacement occur across the map area. Only the main ones are depicted in Figure 15.3. A parallel set of fractures is also present and may be related to the faulting. Maximum displacement is inferred to be less than 1.5 m but most faults involve a displacement of less than 2 cm.

Several examples of discordant warped cleavages indicate that cleavage subparallel to parallel displacement involving lenses or wedges of units has occurred (Figures 15.3b, 15.3c). This is likely a larger-scale manifestation of the discordancies noted in Figures 15.4b, 15.4e, and 15.4f. The sense of displacement, where determined (rarely), is dextral.

Small-scale, dextral displacement was locally noted in short (less than 0.5 m long), northeast-oriented curvilinear faults, which tend to dissipate along strike parallel to layering and/or cleavage. Quartz vein pods are commonly spatially associated with these faults.

Two examples of layer-parallel cataclastic breccias occur within unit 4b (see Figures 15.3b, 15.3c). Neither of these faults could be traced for more than 2.5 m. Fragments within the less than 1 mm to 2 cm wide zones are angular to subangular.

**SYNTHESIS AND DISCUSSION**

A comparison between rocks of the Hemlo gold deposit area and the Northern Eagle property is limited by the relatively small outcrop area at the Northern Eagle property, which offers little insight into other rock types in the vicinity. In the Hemlo gold deposit area, distal turbidite deposits abound. However, black mudstone is uncommon, is not in the immediate vicinity of the deposit and, hence, is not associated with barite.

Some gneissic, feldspathic units in the Hemlo gold deposit area are similar in appearance to the layered and
Figure 15.4. Sketches of various features of the Northern Eagle property barite showing. Sketches are keyed to locations indicated in Figure 3c.
laminated altered feldspathic rocks at the Northern Eagle property, although strain and metamorphic grade are higher in the Hemlo gold deposit area. Dikes, particularly feldspar porphyries, are common to both areas.

Structure

A stripped area the size of the Northern Eagle property occurrence is unlikely to reveal sufficient features to deduce a comprehensive structural history of the surrounding area. However, the relationships that are evident, coupled with the polyphase structural history of the Hemlo area (Muir and Elliott 1987, Muir 1997), form a reasonable basis for deciphering the relationships.

Bedding (S_0) and layering with no primary bedding characteristics (S_1) presently define isoclinal to tight "S"-, "M"- and "U"-shaped folds, consistent with an F_2 generation. No fabric appears to be folded by F_2. Predominant-cleavage-parallel dislocations of fold limbs, juxtaposition of slightly discordant wedges of cleaved, layered-laminated rocks, and boudinage of the cleavage and layering/laminations are consistent with inferred D_2 structures in the Hemlo gold deposit area. The small "Z"-shaped folds with northeast-oriented axial planar cleavage are consistent with F_3 folds and an S_3 cleavage. The S_3 fabric is depicted in Figures 15.4a and 15.4c. Dextral sense displacement and deformation around many quartz veins, particularly those oriented in a northeasterly direction, is similar to D_3 strain features for the Hemlo gold deposit area. The layer- and/or cleavage-parallel cataclastic breccias are also similar to the late D_3 cataclasites in the Hemlo gold deposit area. The sinistral northwest-oriented kink fold is consistent with an F_4 generation fold. Small-displacement, northwest-striking dextral faults and fracture sets are also present in the Hemlo gold deposit area.

A few differences exist between the Hemlo gold deposit area and the Northern Eagle property area. The inferred S_3 fabric does not clearly crenulate an earlier fabric although there are few examples on which to base observations. The metamorphic grade is lower at the Northern Eagle property, hence there is a lower degree of recrystallization and a general lack of true schists.

With respect to the pyrite nodules, the somewhat intricate textures and degree of sphericity likely indicate that development of these nodules postdates much of the strain and regional metamorphism, and hence these nodules would not represent primary features.

Alteration

The limit of visually obvious alteration is remarkably well defined on the footwall (north side). Similarly well-defined limits are evident, for example in the hanging wall (north side) of the Williams A Zone of the Hemlo gold deposit (Muir 1997, Figure 28). Alteration occurs on both sides of the main baritic unit. Some of the complex textures, such as those found in unit 7c (see above) appear to represent replacement features. Others are reminiscent of fracture-controlled or "stockwork"-like alteration in the North Zone of the Hemlo gold deposit, because of highly bleached rocks along a network of intersecting planar zones. The main baritic zone appears to be discordant (Figure 15.3b) to some of the structurally overlying units.

Gliddon (1985) referred to the presence of minor amounts of albite, along with quartz and/or pyrite and/or carbonate, in a number of cases, with respect to layering in baritic rocks and adjacent units. Albite is a major constituent of highly altered rocks, mainly within the western to northwestern part of the Hemlo gold deposit, where gold grades are generally much lower than in the main ore zones, which are predominantly associated with potassic alteration.

The timing of alteration relative to intrusions and strain also appears to be similar to the Hemlo gold deposit area in that the main alteration zone:

a) is not apparently folded by F_2;

b) overprints F_2 folds having similar styles, in rocks which presently have different competencies because of their initial properties and those imposed by alteration;

c) appears to be controlled, in part, by a pre-existing cleavage but has been further deformed largely due to flattening; and

d) predates dike emplacement.

Green muscovite is present in the Hemlo gold deposit and is related to some of the complex alteration processes. However, there are some differences in trace element contents, particularly among chromium, vanadium and barium, between green micas of the Hemlo gold deposit and the Northern Eagle property (see Pan and Fleet 1991). Other obvious differences include a lack of gold and molybdenum mineralization and associated elements, such as Ag, Au and Sb, at the Northern Eagle property, as compared to the Hemlo gold deposit.

Even though the feldspar porphyry dikes are, overall, spatially associated with many of the altered green-muscovite-bearing rocks, in detail, the dike does not appear to be directly related to alteration as both black mudstone and altered green-muscovite-bearing equivalents are in direct contact with the dikes. Rather, it is considered likely that the dikes tended to intrude into a zone of least resistance, namely the most fissile, pre-existing altered rocks.

The shape, size, and orientations of quartz veins are generally similar to those of the Hemlo gold deposit area. For an example comparison, see the detailed map of the Heritage outcrops in Figure 27 of Muir (1997).

Intrusive Relationships

The feldspar porphyry dikes demonstrate various critical relationships, which are consistent with relationships in the Hemlo area (Muir 1997):

a) The dike is oriented parallel to subparallel to transposed layering and the predominant cleavage, which is parallel to the axial planes of the "S"-, "M"- and "U"-shaped folds that are defined by layering.

b) The dike postdates alteration and notable strain (flattening) as evident in contact-cleavage relation-
ships (see Figures 15.4c, 15.4d) and the presence of altered rocks as inclusions. The matrix of the dikes is fine grained and moderately foliated but the pheno-
crys ts do not appear to be deformed. The well-defined alignment and aspect ratios of the inclusions, reported above, the lack of folded (i.e., shortened) cleavage in inclusions, and the lack of notable evidence for shortening of dikes contacts at a high angle to the cleavage (Figure 15.4c) suggests that much of the
alignment was flow induced during dike emplacement.

c) The moderately foliated dike matrixes, coupled with boudinage in some dikes (see Figures 15.3b, 15.3c) indicates that the dikes were subjected to layer-
dike-parallel flattening with orthogonal extension. The most highly strained green-muscovite-bearing lenticular rocks occur spatially with the largest boudinage structure associated with a feldspar por-
phyry dike (see Figure 15.3c).

d) At least some of the quartz veins postdated intrusion of the dikes. Some vein development was synchronous with or postdated development of dike boudins.

OVERVIEW AND IMPLICATIONS

A critical analysis of all of the above observations and deductions would probably lead one to conclude that there is no unequivocal evidence for either syngenetic or epigenetic barite deposition. For instance, it could be argued that the siliceous rock with contorted layering could be a deformed primary chert unit, and the locally occurring, small-scale folds in the apparent contact between the black mudstone and the altered green muscovite equivalent could indicate that alteration pre-
dated all folding, and hence occurred very early in the depositional history. Both of these features might be considered to be consistent with a syngenetic deposition of barite and silica in a sedimentary basin.

Alternatively, the initial interpretation of bedded barite and the premise of the association of large-scale basin deposition of barite, over tens of kilometres, contemporaneously associated with the development of the Hemlo gold deposit, seem suspect. There are no unequivocal bedding features in the baritic rocks (see also Gliddon 1985). Moreover, some of the features appear to represent replacement of pre-existing rocks. Isoclinal folds in layering occur above, below and near the baritic units. Similar-appearing isoclinal folds also occur within the zone of altered rocks, which appear to replace sedimentary rocks. Given that the folded rocks presently represent a range in competencies, the distribution of these folds suggests that alteration was superposed on previously folded rocks. Furthermore, there are no folds in the layered barite, only a slight, broad curvature at the overall outcrop scale (see Figure 15.2) as is shown locally by most other units. Alteration occurs structurally above and below the baritic and silicic units. Alteration products locally include green, Ba-Cr-(Ti) muscovite which corresponds, in part, with the green Cr-V-Ba muscovite of the Hemlo gold deposit (see Pan and Fleet 1991).

The syngenetic origin for the Hemlo gold deposit is essentially no longer considered viable (see discussions in Kuhns et al. 1994; Pan and Fleet 1995, Muir 1997). Given that so many of the temporal and structural relationships between the Hemlo gold deposit and the Northern Eagle property are similar, it would appear that the proposed syngenetic link and origin for the Northern Eagle property is also no longer appropriate. The results of various isotope studies, which cannot distinguish between syngenetic and hydrothermal barite (Hattori and Cameron, 1987), are equivocal. The detailed examination of the Northern Eagle property, described above, coupled with detailed work in the Hemlo gold deposit area, provides an opportunity to envisage the deposition of epigenetic barite in a larger-
scale perspective involving the Hemlo greenstone belt.

Corfu and Muir (1989b) postulated a major structural greenstone-belt-scale discontinuity, likely related to the Hemlo fault zone with which the Hemlo gold deposit is spatially associated (see also Williams et al. 1991). Recent, regional-scale, structural and metamorphic analy-

ses indicate that the western end of the Hemlo greenstone belt is at a higher structural level and corresponding lower metamorphic grade than the eastern part (Jackson 1998). Furthermore, Jackson (1998) ties the existence of such a large-scale discontinuity not only to potential sites for mineralization, such as the Hemlo gold deposit, but also to style, that is, brittle ductile and vein related versus ductile and disseminated.

Specifically with respect to barite, the above approach appears to have validity. For example, gold-molybdenum mineralization in the Hemlo gold deposit, which is, in places, spatially associated with various forms of barite, is disseminated and spatially associated with high-strain zones such as the Lake Superior shear zone and Moose Lake fault zone. These zones occur near and parallel to the Hemlo fault zone (see Hugon 1986; Burk et al. 1986; Kuhns et al. 1994; Muir 1997). At least one high-strain zone, also associated with alteration, occurs several kilometres to the west-northwest of, and on strike from, the Hemlo gold deposit (Muir 1997). Minor vein-related barite and crudely spatially associated minor gold and molybde-
um mineralization occur in well-strained, east-northeast-striking fissile rocks of Heron Bay adjacent to Lake Superior, at the west end of the Hemlo greenstone belt (i.e., Peckongay property; Schnieders and Smyk 1991). The Northern Eagle property occurrence is also associated with highly strained rocks (see also Gliddon 1985). Here, the characteristics and configuration of both the main baritic zone and the structurally overlying dike-like subsidiary barite zones, as exposed (see Figures 15.2 and 15.3b), is consistent with post-regional folding (F2), large-scale dilational infilling and replacement.

Fyon et al. (1992) noted that there are 2 generalized orientations of gold-associated regional deformation zones in major gold camps in Ontario, namely east-north-
east to northeast and east-southeast to southeast. In the Hemlo greenstone belt, the inferred main "break" and possible "splays" are somewhat sinuous. At the Hemlo gold deposit, the strike of high-strain zones is approxi-
mately east-southeast. At the Peckongay property, the
strike is east-northeast. At the Northern Eagle property where there is essentially no gold or molybdenum, the strike is approximately east. In this context, the Northern Eagle property may represent an intermediate stage between the Hemlo gold deposit and Heron Bay sites controlled, in part, by structural level and orientation. In addition, the presence of minor albite and not potassium feldspar is consistent with low to nil gold values at the Northern Eagle property. The penultimate question then, is what are the fundamental controls which govern why gold occurs where it does, irrespective of the presence of large-scale, structurally controlled alteration. The orientation of the Northern Eagle alteration zone and the lack of potassic alteration may be the fundamental reasons for the lack of gold at the Northern Eagle property.

Combining the importance of the delineation of high-strain zones, related alteration and associated rock types is important in targeting greenstone belts worth pursuing for gold exploration. In this case, the occurrence of barite may be a variable that relates to the large-scale picture involving hydrothermal fluids tapped by crustal weaknesses during specific stages of magmatism.

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16. Project Unit 96–097. Pattern of Total Strain in Western Parts of Birch–Uchi (Confederation) Greenstone Belt, Northwestern Ontario

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The western Birch–Uchi greenstone belt on Woman Lake and vicinity (Prisak and assistants 1969, 1971; Thurston 1985; Fyon and Lane 1986; Stott and Corfu 1991; van Staal 1998) is composed of three tectonic assemblages (Figure 16.1), which have been strained heterogeneously. Study of the pattern of total strain (a net result of ductile deformation) permits the structural testing of tectonic hypotheses accounting for the structure of greenstone belts in general and the Birch–Uchi belt in particular. In a previously published summary, Crews and Schwerdtner (1997) referred to the northward transportation hypothesis (Stott and Corfu 1991), according to which a lobe of Confederation assemblage rocks moved northward with respect to adjacent rock masses of assemblages I and II (see Figure 16.1). Results of shear-sense analysis (Crews and Schwerdtner 1997) are compatible with this hypothesis, although the Woman–Confederation assemblage boundary may be slightly farther east than originally proposed (C.R. van Staal, Geological Survey of Canada, personal communication, 1998). For example, it may coincide with the stratigraphic base of a metasedimentary unit mainly composed of polymictic metaconglomerate (Figure 16.2). This unit is thrown into a steeply plunging S-fold, whose short limb is buckled and whose axial plane parallels the quasi-pervasive mineral schistosity in most rock units of the 3 tectonic assemblages (see Figures 16.2 and 16.3).

![Figure 16.1. Red Lake–Birch–Uchi greenstone belts and vicinity; B1, B2 granitoid batholiths. T = centre of triple junction infoliation (schistosity) pattern and the strain field; W = Washagomis Lake.](image)
Figure 16.2. Lithological map, central and northern Woman Lake.
Moslehi is estimating the principal-strain ratios in metaconglomerate, and will be able to determine the magnitude of total shear strain parallel to the Woman–Confederation assemblage boundary.

Individual schistosity trajectories, constructed from field measurements, transect both assemblage boundaries on Woman and Spot lakes (Crews and Schwerdtner 1997), and this may imply that ductile deformation on a metre scale started everywhere at the onset of S-folding. Schistosity diminishes at a major triple junction in Woman assemblage rocks (see Figure 16.1), where the steep strain fabric (stretched lapilli, mafic clots in metagabbro, etc.) is dominantly linear. Triple junctions in the pattern of schistosity trajectories are known from a variety of structures (Brun 1983), and therefore provide no direct evidence for Ramberg (1981) type diapirism of large granitoid plutons. In ideal diapiric models of greenstone belts (Figure 16.4), the triple junction has a central position, within the youngest stratigraphic units. This contrasts with the situation in the western Birch–Uchi belt (see Figure 16.1), where the triple junction is in Woman assemblage rocks, near the Trout Lake batholith, and where schistosity attitudes seem to be unaffected by the assemblage boundaries (Fyon and Lane 1986). East of northern Woman Lake, strain on a metre scale diminishes toward Washagomis Lake (see Figure 16.1).
Figure 16.5. Structural map of Washagomis Lake area.
Mafic metavolcanics on Washagomis Lake (see Figures 16.1 and 16.5) are replete with pillow structures that are virtually devoid of conspicuous strain on a metre scale. These Confederation assemblage rocks have been thrown into northeast-southwest trending large folds with narrow, highly deformed hinge zones that are generally not exposed. Such folds attest to strain on a kilometre scale (Schwerdtner, Crews and Moslehi 1998), which may predate the schistosity of rocks exposed on Woman Lake and vicinity.

In summary, the strain pattern of the western Birch–Uchi belt is incompatible with Ramberg-type granite diapirism. At least two deformation pulses may have affected the metavolcanics and associated metasediments, but the tectonic style needs to be further elucidated by studying the Swain Lake deformation zone and its wall rocks. Such a study is to be undertaken in 1999-2000.

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17. Project 95–034. Western Superior NATMAP: Progress in Understanding the Evolution of Continental and Oceanic Blocks


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INTRODUCTION

The Western Superior NATMAP (National Mapping Program) project was initiated in 1997 with the principal objective of updating the tectonic framework for greenstone–granite subprovinces of the western Superior Province through establishment of relationships between older (pre-2.8 Ga) and younger (ca. 2.7 Ga) crustal blocks. The existing framework, which regards the subprovinces of the western Superior Province as its fundamental tectonic building blocks, accreted at ca. 2.7 Ga (Williams et al. 1992), does not adequately explain the widespread presence of 3.0 to 2.8 Ga crust. For example, the Wabigoon subprovince has been considered as a single entity despite the presence of greenstones and granites with ages ranging from 3 to 2.7 Ga. Hence, the main focus of the project are greenstone belts adjacent to the 3.07 to 2.93 Ga central Wabigoon region, and belts surrounding the 3 Ga crust in the North Caribou terrane (Thurston, Osmani and Stone 1991). Relationships are being elucidated through examination of the age, geochemistry and structural history of

- greenstone belts adjoining known old crustal blocks, such as the Sturgeon–Savant (Sanborn-Barrie, Skulski and Whalen 1998), Obonga (Tomlinson et al. 1996) and Onaman–Tashota (Stott, Davis and Parker 1998) belts of the Wabigoon subprovince, Confederation (van Staal 1998) and Wallace Lake belts of the Uchi subprovince, and Island Lake (Lin, Cameron et al., in press), Stull Lake (Corkery, Skulski and Whalen 1997; Stone, Halle, Lange and Chaloux, this volume) and Knee Lake (Syme et al., in press) belts on the northern margin of the North Caribou terrane;

- granitoid regions of the central Wabigoon subprovince (Percival 1998; Stone, Halle and Chaloux, this volume) and northern flank of the North Caribou terrane (Stone, Halle, Lange and Chaloux, this volume); and

- two metallogenetically important deposit types: volcanogenic massive sulphide (VMS) and gold in old and young crustal blocks (Parker, this volume).

In estimating the extent of oceanic and continental crust at 2.7 Ga, the NATMAP project will provide an enhanced framework that should serve to focus mineral exploration in terranes with the highest potential.

WABIGOON SUBPROVINCE

The central Wabigoon region (“Wabigoon diapiric axis” of Edwards and Sutcliffe (1980)) has been postulated to represent 2.93 to 3.07 Ga (Davis, Sutcliffe and Trowell 1988) basement to adjacent supracrustal belts. Work in 1997 and 1998 focussed on the central Wabigoon region in the Sturgeon–Obonga lakes corridor (Percival 1998; Percival et al., in press; K. Tomlinson, Geological Survey of Canada, unpublished data, 1998) and flanking greenstone belts to the west (Sturgeon–Savant belt; Sanborn-Barrie, Skulski and Whalen 1998; Sanborn-Barrie and Skulski, in press) and east (Onaman–Tashota belt; Stott, Davis and Parker 1998; Stott and Straub, this volume).

A first-order subdivision of the central Wabigoon region can be made into gneissic and homogeneous granitoid rocks (Percival 1998; Stone, Halle and Chaloux, this volume). The gneisses carry an S1 layering and common F2 folds, both older than the main homogeneous tonalites and granodiorites, which are characterized by a penetrative S3 foliation and gently east-plunging F4 folds. Tonalitic gneiss and S1 fabrics appear older than 2774 ± 2 Ma tonalite (Davis 1989). An unconformity may be present between structurally complex tonalite gneisses and quartz-rich sedimentary units in the Harmon Lake area (Percival et al., in press). Most of the dated pre-2.9 Ga granitoid units carry simple foliations and are thus difficult to distinguish from younger plutonic rocks. The younger granitoid rocks occur as sheets up to several kilometres thick, enclosing lenses of older gneissic and supracrustal rocks, including a 3 km thick sandstone unit east of Sturgeon Lake. These amphibolite-facies sedimentary
rocks contain detrital zircons in the range 2709 to 2890 Ma and are cut by granite-diorite dated at 2709 ± 4 Ma (V. McNicoll, Geological Survey of Canada, unpublished data, 1998). The map pattern of the granitoïd region is dominated by gently east-plunging, upright F4 folds (Percival 1998). Two sets of ductile shear zones accompanied by syntectonic granite cut the map area and include dextral-slip, kilometre-scale, D3, east-striking shear zones, and north-northeast striking, sinistral D4 shear zones (L > S).

Concordant S1 fabrics along the eastern margin of the Sturgeon–Savant belt correlate with S3 foliation in adjacent granites. Similarly, east-trending D2 structures in the Sturgeon–Savant belt correlate with D4 features in the granitoïd complex. However, D2 structures have steep hinges in the greenstones whereas D4 folds have shallow plunges in the granitoïd domain, suggesting that D1 fabrics were upright in the supracrustal belt and D3 structures were recumbent in the granitoïds.

The Savant–Sturgeon belt occurs at the interface between old (ca. 3 Ga) rocks of the central Wabigoon region and juvenile oceanic terranes (2.77 to 2.72 Ga) to the west. Most of the belt has been mapped in detail (Trowell 1983), with the area surrounding the 2.735 Ga Mattabi deposit in the greatest detail (Franklin and Morton, in press). The Juten Group at the eastern margin of the belt comprises a thin clastic wedge overlain by basalt and appears to represent a tilted, volcanic rift margin on greater than 2.9 Ga continental crust. Quartz-rich clastic rocks with detrital zircon ages of 2.91 to 3.3 Ga extend from Savant to Vista Lake and in northeast Savant Lake, rest unconformably on an older sequence of supracrustal rocks (Skulska, Sanborn-Barrie and Stern 1998; Sanborn-Barrie and Skulska, in press).

The central and western parts of the belt represent a conformable, ca. 60 million year sequence of episodic volcanism. Oceanic crust at the base of the pile is capped by 2775 Ma felsic tuffs between Fourbay and Couture lakes (T. Skulska, Geological Survey of Canada, unpublished data, 1998). These rocks appear conformably overlain by the Six Mile Lake cycle, probably correlative with the 2744 Ma Beckington West cycle and 2745 Ma Handy Lake group, in turn conformably overlain by the 2718 Ma central Sturgeon Lake cycle that represents an arc rift sequence.

The younger than 2704 Ma Savant sedimentary group is interpreted to have formed in a foredeep separating 2775 to 2704 Ma oceanic volcanic rocks in the west from a ca. 2710 Ma continental arc built on a less than 2900 Ma volcanic rift margin in the east (Sanborn-Barrie and Skulska, in press). Collision between these two terranes is recorded by two main episodes of deformation. In the north, D3 involved northwest-striking F1 folds and axial planar cleavage (S1). This is overprinted by east-northeast trending, steeply plunging F2 folds and associated axial planar foliation, S2. In the south, the dominant fabric is a north-striking S1 that is axial planar to F1 folds. Overprinting D2 fabrics affect rocks north of Sturgeon Lake but are nonpenetrative to the south.

To the east, the Onaman–Tashota greenstone belt may mark a continuous transition in greenstones from a ca. 2.7 Ga continental margin on the south to supracrustal rocks of about 3 Ga in the north. Variations in assemblage types and metallogeny will provide insight into relationships between the central Wabigoon region and ca. 2.7 Ga continental margin. The belt comprises strata ranging from 3.056 Ga, the oldest volcanic rock known in the Superior Province, to 2.739 Ga (Stott, Davis and Parker 1998). Massive to pillowed basaltic flows are nonvesicular and locally accompanied by iron formation. Thick porphyritic rhyolite flows with abundant autobreccia and hyaloclastite east of Gzowski Lake lie unconformably on basaltic basement. Clastic sediments in the central part of the belt are derived from felsic volcanic rocks, indicating that only the felsic edifices were emergent. Felsic volcanic rocks, locally 2739 Ma, were built on pre-2.9 Ga crust in the northern half of the belt. The greatest volume of pyroclastic rhyolite and VMS-related alteration (Fe-enrichment) occurs in the vicinity of Marshall Lake on the northern (continental) margin of the belt. The tectonic history across the region includes: 1) D3 structures preserved in the basaltic flows east of Gzowski Lake; 2) growth of emergent rhyolite volcanoes and erosion into flanking epiclastic basins; 3) D2 deformation across the central part of the belt; 4) D2 strain aareules imposed on surrounding rocks during emplacement of 2.698 Ga plutons; and 5) D4 dextral high-strain zones.

At the southwest end of the central Wabigoon region, several years of 150 000 scale mapping (e.g., Stone, Halle and Chaloux, this volume) have revealed that the Irene–Eltrut lakes complex consists of biotite tonalite, tonalite gneiss and thin greenstone sills intruded by granite. To the east are the ca. 2.93 Ga Finlayson and 3.0 Ga Lumby Lake belts, as well as the 3.0 Ga Marmion tonalite (Davis and Jackson 1988). Small greenstone belts to the north represent extensions of the ca. 2.74 Ga Kakagi Lake belt. Thus the region straddles an interface between domains of Mesoarchean and Neoarchean age. Despite high strain levels in the Raleigh Lake greenstone belt, major rock types identified include pillow basalts, iron formation and gabbro; dacitic to rhyolitic flows, tuffs and breccias; and sandstone-siltstone sequences. Felsic plutonic rocks are subdivided in chronological order into: 1) tonalite gneisses; 2) moderately foliated biotite tonalite that locally cuts greenstones; 3) two-mica granite, locally mineralized with rare metals; 4) hornblende-bearing tonalite to granite; 5) leucocratic biotite granite (e.g., 2685 ± 2 Ma White Otter batholith; D. Davis, personal communication, 1996); and 6) diorite-monzodiorite-monzonite plutons (sanukitoid suite). The youngest intrusive rocks carry weak magmatic foliations, whereas older rock types have well-developed, planar and locally linear fabrics. Steep shear zones are developed through the area, particularly in supracrustal belts. Structural data indicate that late intrusions are sheet-like, with margins at least locally above supracrustal and possibly older crustal material; significant crustal growth was accomplished by magmatic intraplateing (Stone, Halle and Chaloux, this volume). In this interpretation, old crust is more extensive at depth than is indicated at surface.

**UCHI SUBPROVINCE**

In the Uchi subprovince, the Confederation Lake belt consists of the ca. 2960 Ma Balmer, ca. 2840 Ma Woman and ca. 2740 Ma Confederation assemblages (Nunes and Thurston 1980). The relationship between these units is not adequately explained in current tectonic models, provid-
ing the stimulus to re-examine contact relationships in this low-strain area (van Staal 1998; Rogers, van Staal and McNicoll, in press; Schwerdtner et al., this volume). A simple structural history includes two major folding events that affected all assemblages. Marker horizons in the Confederation assemblage define large folds, including a large, doubly plunging, tight to isoclinal, north-south trending D1 antecessional fold, folded into an “S” by northeast-southwest trending D2 folds. No significant displacement across the Balmer–Woman–Confederation boundary is indicated by a stitching, syn-Woman (2.84 Ga) gabbro (V. McNicoll, Geological Survey of Canada, unpublished data, 1998). The Woman–Confederation boundary, redefined in light of new geochemical data, coincides with the Rowe Lake shear zone. This is not a major fault because: 1) the shear zone is a narrow (<5 m) ductile structure; 2) it cuts the youngest phase of the Trout Lake batholith (ca. 2.73 Ga), which intrudes both the Balmer–Woman and Woman–Confederation boundaries; and 3) the shear zone is one of a series of postfolding, north-northeast–southwest trending sinistral shears. In the north, the Woman–Confederation assemblage boundary is defined by clastic sedimentary rocks. Petrographic and geochemical (K. Tomlinson, Geological Survey of Canada, unpublished data, 1998) characteristics of the Balmer have ophiolitic affinities, indicating that the Balmer, and conformably overlying Woman and Confederation assemblages, formed in an oceanic setting.

To the west in the Wallace Lake area it may be possible to better constrain the relationships between the ca. 3 Ga North Caribou terrane and Uchi supracrustal rocks. At Wallace Lake, a >2.92 Ga sequence of quartzite, carbonate and iron formation (Conley Formation) is likely unconformable on 3 Ga basement of the North Caribou terrane. Volcanic rocks of the Big Island Formation, including spinifex-textured komatiite, conformably overlie Conley sediments, and are in turn overlain by volcano-sedimentary units of the Siderock Lake Formation (Sasseville and Tomlinson, in press). These are separated from 2.85 to 2.70 Ga supracrustal packages of the Rice Lake and Garner Lake belts by late ductile faults. The structural chronology involves early (D1) map-scale folds and accompanying axial-planar foliation, variably overprinted by a steep, east-west trending S2 foliation.

**SACHIGO SUBPROVINCE**

In the Sachigo supraregion, supracrustal rocks in the Edmund–Stull–Knee Lake corridor are flanked by felsic granitoid rocks (Corkery, Skulski and Whalen 1997; Stone, Halle, Lange and Chaloux, this volume). In the Edmund–Stull Lake belt, significant new observations include the following: 1) Hayes River Group basalts form a thick sequence with mid-oceanic-ridge basalt (MORB)-like geochemistry (Corkery and Skulski, in press); 2) basalts north of the D1 Wolf Bay shear zone have geochemical affinities to the Oxford Lake Group; 3) the White House tonalite (2734 ± 2 Ma) provides a minimum age for the Hayes River Group and indicates that rocks of Oxford Lake age intruded through Hayes River rocks (Corkery and Heaman, in press; Corkery and Skulski, in press); and 4) four generations of deformation affected the belt, although D1 is recognized only in the mafic complex (Jiang and Corkery, in press). To the east, supracrustal belts are narrower and strain higher, although similar stratigraphic assemblages are recognized (Stone, Halle, Lange and Chaloux, this volume). In the Knee Lake belt, a regional stratigraphy for the 2830 Ma, basalt-dominated Hayes River Group was established and an early fault (thrust?) identified that potentially juxtaposes distinct lithotectonic sequences (Syme et al., in press). The Hayes River Group was deformed prior to deposition of the 2706 Ma Oxford Lake Group, which comprises alkaline-cal-alkaline volcanic and volcanioclastic rocks and clastic sedimentary rocks (Syme et al., in press). Two generations of folds, as well as numerous faults and shear zones, transect the belt, including the dextral transpressive southern Knee Lake shear zone (Lin, Jiang et al., in press). Similar stratigraphic units characterize the Island Lake belt to the south, where the 2861 Ma Hayes River Group is exposed in a series of fault-bounded panels and unconformably overlain by sediments of the Island Lake Group. Two gold mineralizing events are associated with shear zone development (Lin, Cameron et al., in press).

**REGIONAL METALLOGENY**

A multiyear metallogenic study in the northwestern Superior Province is focused on VMS and gold mineralization. The aims are to document the distribution and features of VMS mineralization as a means of comparing the metallogeny of old (pre-2.74 Ga) and young (2.7 to 2.74 Ga) assemblages. Preliminary observations (Parker, this volume) indicate that Zn-Pb-Ag-Cu mineralization in some pre-2.74 Ga rocks, such as the Balmer assemblage in Red Lake and the North Rim assemblage at Eayapimikama Lake, is associated with quartz–magnetite iron formation and ferruginous chert. This style of mineralization is not restricted to older assemblages based on similar associations in the 2.72 Ga Manitouwadge belt. Common synvolcanic, chlorite or amphibole–magnetite–garnet alteration assemblages are associated with VMS mineralization in old and young metavolcanic assemblages in the Sachigo (Eayapimikama, Setting Net lakes) and Uchi (Maskooch, Discovery lakes) subprovinces. This alteration occurs in a network of intersecting veins and is commonly folded by D1. The presence of orthoamphibole-garnet-cordierite alteration in felsic metavolcanic rocks at Shrimp Lake in the North Spirit Lake belt (Wood 1988) may represent amphibolite grade, seafloor alteration associated with VMS mineralization.

Initial results of the Western Superior NATMAP project indicate significant variation in relationships between Mesozoic and Neoproterozoic greenstone belts within the Wabigoon, Uchi and Sachigo subprovinces. The north-trending tectonic interface between juvenile Neoarchean belts of the western Wabigoon (Henry, Stevenson and Gariety 1998) and a continental margin to the east appears to be preserved in the Sturgeon–Savant belt (Sanborn-Barrie and Skulski, in press). A similar transition may exist in the Uchi subprovince, where >2.92 Ga
supracrustal rocks are autochthonous on 3 Ga basement in the Wallace Lake belt in the west, whereas in the east, units of similar age were built on oceanic crust at Confederation Lake. In the north, the youngest supracrustal rocks (Oxford Lake Group) were deposited on older crust; the relationship between the older (2.86 to 2.83 Ga) Hayes River Group and 3 Ga crust of the North Caribou terrane is under active study. Subprovince boundaries are late features perhaps only indirectly related to accretionary structures.

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18. Project Unit 97–07. Tectonic Hypotheses Regarding the Late Orogenic History of the Birch–Uchi Greenstone Belt

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INTRODUCTION

The Birch–Uchi Archean greenstone belt (Figures 18.1, 18.2) is a large (about 80 km width in map view), arcuate, tectonically complex portion of the Uchi Subprovince (Stott and Corfu 1991). As part of the ongoing Western Uchi NATMAP Project, sedimentological studies of selected areas of the Birch–Uchi belt (e.g., Devaney 1997, 1998) are intended to provide data that will assist in solving the puzzle of the multi-stage volcanic and tectonic evolution of this belt. (No field work was conducted on this project this year.)

Such a tectono-stratigraphic analysis, in combination with ongoing volcanological, structural, geochemical and geochronological work (van Staal 1998), will also clarify understanding of the stages of formation and deformation of the mineral deposits (e.g., Cu-Zn, Au) in the Birch–Uchi greenstone belt.

Although somewhat speculative and perhaps premature, the two hypotheses presented below may help to explain the area’s late orogenic history. Much structural and geochronological work will be required to test these hypotheses (future work of this project and that of van Staal 1998), but the results should be particularly applicable to exploration for tectonically late lode gold deposits.

INDENTER/PROMONTORY HYPOTHESIS

Supracrustal assemblage ages in the Uchi Subprovince (Stott and Corfu 1991, Figure 6.3) suggest a structural offlapping of assemblages and/or early orogenic, broadly north-verging thrust-stacking (Williams et al. 1992, Figure 25.10b), followed by late orogenic wrench faulting (Stott and Corfu 1991, Corfu et al. 1995). Evidence for the supposed thrust stacking and other early orogenic deformation stages is poor (or subtle?), perhaps the result of strong overprinting by later tectonic events.

The large-scale map patterns of geological units in and marginal to the Birch–Uchi greenstone belt are shown in Figures 18.1 and 18.2 (see also: Stott and Corfu 1991; Beakhouse 1989, Figures 5, 6). In the west-southwest part of the belt, structurally offlapping assemblages young away from the Trout Lake Batholith (Stott and Corfu 1991, Figure 6.5) and lacin sinistral-sense structural features are well preserved (Fyon and Lane 1985, Fyon and O’Donnell 1986, Crews et al. 1997, van Staal 1998). As outlined below, it is suggested that the large-scale map patterns faithfully record the late orogenic, approximately northwest-southeast compression (present geographic coordinates used herein) of a previously tectonized and heterogeneously structured area.

As shown in Figure 18.2, the Jeanette Lake granitoid complex (Beakhouse 1989, Figure 6; or “Bamaji–Blackstone batholith” of Stott and Corfu 1991, Figure 6.2) is envisioned as a structurally competent microcontinental block (see below); the arcuate/protrusional outline of folded strata northwest of the Jeanette Lake granitoid complex suggests that this supposedly competent block extends below wacke metasedimentary rocks. It is thought that this block was forced northward, causing either broad folding of the Birch–Uchi greenstone belt around an incoming indenter (the Jeanette Lake granitoid complex) or increased compression of the Birch–Uchi greenstone belt around an already adjacent basement promontory (the Jeanette Lake granitoid complex). Within the Birch–Uchi greenstone belt, many faults and other structures appear to directly reflect this regional geometry (see Figure 18.2). In the west to southwest parts of the belt, late faulting is predominantly sinistral (Fyon and Lane 1985, Fyon and O’Donnell 1986, Crews et al. 1997, van Staal 1998). In the north to northeast parts of the belt, the best preserved (and likely latest) faulting is predominantly dextral (i.e., the northeast part is a “mirror image” of the faulting in the southwest part), as evidenced by outcrop-scale folds and offsets (Fraser 1998) and the presence of dextrally sigmoidal fault-bounded conglomeratic pull-apart basins (less than 10 km long; analogous to large-scale tension gashes: Devaney 1997, Figure 18.1; Devaney 1998) filled partly with locally derived polymict conglomerate. Such sedimentological evidence (Devaney 1997, 1998) indicates that these are late (post-Conformation Assemblage), “Timiskaming-type” pull-apart basins (“late, fault-bounded and unconformable basins” of Williams et al. 1992, Table 25.2) which may be correlative with the conglomeratic Billett assemblage (Stott and Corfu 1991, Stott 1996) in the central Uchi Subprovince. In the northwest part of the Birch–Uchi greenstone belt there was presumably more orthogonal compression of approximately northeast-trending stratal packages. Note the paucity of late granitic plutons in the northwest part of the belt (a convergent zone) versus the greater abundance of such plutons in the supposedly more extensional or transtensional zones (e.g., areas of strike-slip in the west/southwest parts of the belt; cuspate areas of structural divergence, such as near the Mink Lake stock in Figure 18.2).
In the northwest part of the belt, the Western Peninsula sub-area (Beakhouse 1989; Devaney 1997, Figure 18.1) is considered to be a small-scale block of relatively competent felsic metavolcanic rocks both bounded and internally divided by deformation zones. It is suggested herein that as a result of northwest-southeast compression, this structurally competent felsic unit acted as a “scholle” (structural flake or block, termed the “Western Peninsula Scoll’e”; “WPS” in Figure 18.2b) which experienced tectonic wedging or lateral escape to the northeast, from a zone of

Figure 18.1. Shaded relief total field aeromagnetic pattern of the Birch–Uchi greenstone belt (from OGS 1997); light tones have higher magnetic susceptibility (e.g., metabasalts), dark tones have the least magnetic susceptibility (e.g., metawackes, granitoids); compare with Figure 18.2.
Figure 18.2. Geological map (a; after Stone 1998, Figure 5) and interpretive cartoon map (b) of late orogenic structural components of the Birch–Uchi greenstone belt and surrounding areas. In b: large arrows indicate northwest-southeast regional compression/transpression, WPS = Western Peninsula Scholle, JLGC = Jeanette Lake Granitoid Complex, TLB = Trout Lake Batholith, and stippled areas are conglomeratic dextral pull-apart basins (basin outlines are, from upper right to lower left: vague, sigmoidal, partial sigmoid, coalesced sigmoids; see also Devaney 1997, Figure 1).
convergence and compression to an adjacent zone of extension and structural divergence along lithological contacts and faults (the Swain Lake Deformation Zone of Beakhouse 1989).

The map pattern of the proposed indenter or promontory flanked by symmetric and opposing sinistral versus dextral fault systems (Figure 18.2b) is similar to larger-scale indenter map patterns seen in some younger orogens (e.g., Matte 1986, Nijman and Savage 1989, Ratschbacher et al. 1991a,b, Laubscher 1992) and some Archean orogens (e.g., Krapez and Barley 1987, Stott and Corfu 1991, Williams et al. 1992, Corfu et al. 1995). In contrast to the interpretation above, Good (1988) and Beakhouse (1989) stressed the role of belt-marginal granitoid intrusions in the late, ‘D2’ deformation of the northern Birch–Uchi greenstone belt.

During late orogenic stages of compression or transpression, likely with local areas of transtension (e.g., pull-apart basins, releasing bend plutons), orogens can be segmented or partitioned into blocks or sub-units (e.g., Hill 1982, “rhombohedral blocks” of Rothery and Drury 1984, “orogenic sublids” of Laubscher 1992). Great structural complexity will result from intermittent strike-slip, oblique-slip and/or dip-slip faulting along the margins of intra-orogenic blocks of various shapes (Holdsworth et al. 1998); such blocks are likely to “jostle” against each other (like ice floes), sticking and then slipping, and the blocks may break into smaller sub-blocks (via more local extension or compression; for example, Dewey and Sengor 1979, Rothery and Drury 1984). Examples of tectonic wedging or lateral escape of scholles and related fault patterns are discussed by Dewey and Sengor (1979), Sylvester (1988), and Ratschbacher et al. (1991a,b). The concept of the proposed “Western Peninsula Scholle” should be tested by detailed structural mapping; this scholle should have evidence of dextral faulting along its southeast margin and sinistral faulting along its northwest margin.

Such structurally complex blocks of various scales may be present within and around Archean greenstone belts, including the Birch–Uchi greenstone belt. In the western Superior Province, some block margins are likely to have been determined by the regionally transpressive (north-northwest to south-southeast) conjugate fault patterns of east-trending dextral faults and northeast-trending sinistral faults (Card et al. 1989, Williams et al. 1992); in the Uchi Subprovince, southeast-trending dextral faults are an additional factor (Stott and Corfu 1991).

Implications for Gold Exploration

The Birch–Uchi greenstone belt has numerous gold occurrences (Parker and Atkinson 1992), including the past-producing Jason Mine (Casumplit Lake area; Beakhouse 1994) and the Springpole Lake prospect (Barron 1996; Devaney 1997, Figure 1). The “Western Peninsula Scholle” (see above) has gold occurrences around its margin. Given that gold was typically deposited very late in the tectonic history of greenstone belts (e.g., Colvine et al. 1988, Card et al. 1989, Williams et al. 1992), structural analysis using the concepts outlined above (e.g., fault systems flanking indenters and scholles, Timiskaming-type basins) could be relevant to insightful exploration for mesothermal lode gold deposits.

The Springpole gold deposit has been interpreted as part of an alkalic (trachyte, lamprophyre) “high-level porphyry and breccia pipe complex” with “both intrusion-related and epithermal types” of gold mineralization (Barron 1996, p.237). Barron discounted a Timiskaming-type pull-apart basin setting for this 2757 to 2715 Ma alkalic volcanic complex (one which includes a bedded tuff maar deposit: Barron 1996) and the adjacent polymict conglomerate unit, but the conglomerate in the Springpole area appears identical to the other polymict conglomerate units (pull-apart basins) in the vicinity (Devaney 1997, Figure 1). Despite Barron’s interpretation of epithermal, rather than mesothermal, mineralization, the geology of the Springpole area rocks conforms with much of the fault-bounded pull-apart basin scenario typical of late orogenic Timiskaming-type assemblages (e.g., Card et al. 1989, Wyman and Kerrich 1989). Also, geochronological work on the belt’s volcanic and plutonic rocks by Beakhouse et al. (1998) suggests that the young end of the Springpole rock age range (ca. 2715 Ma) should be a late tectonic age.

Testing of the Jeanette Lake Granitoid Complex

Indenter/Promontory Hypothesis

Testing of the indenter interpretation in which the undated Jeanette Lake granitoid complex is a microcratic block around which the Birch–Uchi greenstone belt was compressed requires dating of the Jeanette Lake granitoid complex. Wallace (1983) mapped tonalitic gneisses and undeformed granitoids in the Jeanette Lake granitoid complex. Tonalitic gneisses dated at 2.8 to 2.9 Ga are present to the east along regional strike in the central Uchi Subprovince (Pembina, Quarry gneisses: Stott and Corfu 1991, Stott 1996) and to the west (Trout Lake Batholith: Noble 1989). Based on the sporadic presence of such gneissic areas in the southern half of the Uchi Subprovince (gneissic areas interpreted herein as microcratic blocks which formed nuclei for subsequent granitic batholithic complexes), it is suggested that any older parts of the Jeanette Lake granitoid complex may be correlative or similar to these 2.8 to 2.9 Ga gneisses.

DOUBLY VERGENT OROGEN HYPOTHESIS

Following regional geological summaries (Stott and Corfu 1991, Williams et al. 1992, Corfu et al. 1995, Stone 1998), the orogenic model of Willett et al. (1993), and selected orogenic belt case studies (e.g., van Kranendonk and Wardle 1996), it is interpreted that the Uchi and English River subprovinces together form a doubly vergent orogen.

In this view (Figure 18.3), various supracrustal sequences of Andean-type continental marginal arc origin
DOUBLY VERGENT OROGEN

English River Subprovince

Uchi Subprovince

S - vergent thrusts, N - younging strata

\texttimes

\textbullet

N - vergent thrusts, S - younging strata

S

Winnipeg River Subprovince

Berens River Subprovince

\(J.R.\) Devaney

Figure 18.3. Highly simplified cross-section of the paired Uchi and English River Subprovinces modeled as a doubly vergent orogen between cratonic blocks; microcratonic blocks within the Uchi Subprovince are not shown. The dextral Sydney Lake Fault (note the circular ‘into page’ and ‘out of page’ symbols) separates the English River and Uchi Subprovinces.

(Uchi Subprovince) and accretionary prism origin (western English River Subprovince) were compressed between the cratonic blocks of the Winnipeg River and Berens River Subprovinces, ideally producing south-verging thrust packages with predominantly north-younging beds in the English River Subprovince versus north-verging, predominantly south-younging thrust packages of the Uchi Subprovince.

Within such a doubly vergent orogen, the simplistic view of various fault-bounded supracrustal packages or sequences would obviously be complicated by: 1) the presence of any microcontinental blocks within the Uchi Subprovince (e.g., 50 km wide exposed areas of 2.8 to 2.9 Ga tonalitic gneisses; see above); 2) various scales of folding; 3) any earlier orogenic phases; 4) late Archean (2.7 Ga) intrusions; and 5) late orogenic regional strike-slip (regionally transpressional) faulting.

Testing of the doubly vergent orogen hypothesis is a broad aspect of many ongoing studies of northwest Superior Province (e.g., the NATMAP and LITHOPROBE projects). When testing the model, local variability is to be expected. For example, work by Hynes (1997) in the Lac Seul area of the English River Subprovince revealed early structures verging to the west or northwest, in contrast with the model prediction of south-verging structures (see Figure 18.3), but this can be accounted for by the backthrusting that occurs in many accretionary prisms (Silver and Reed 1988, Willett et al. 1993) or by the presence of a sub-basin (such as the trench-slope basins that typically overlie prism units) with a deformational style different from that in underlying prism units. In the Uchi Subprovince, late orogenic, regionally transpressional wrench faulting and deformation around the indenting promontories of any microcontinental blocks are likely to have produced a complex mosaic of various scales of fault-bounded blocks, local extensional or transtensional zones (pull-apart basins, releasing bend plutons), and local compressional/transpressional zones, all superimposed on the structures of an earlier fold-and-thrust stage. Exploration for mineral deposits in regions with such complex histories will require sophisticated and detailed structural analysis.

REFERENCES


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BACKGROUND

The Wabigoon Subprovince forms one of the major superterrane belts of the Archean Superior Province in Ontario. Although considerable work has been done to map and interpret the western half of this subprovince (e.g., Blackburn et al. 1991), comparatively little interpretive work had been done east of Lake Nipigon (e.g., Blackburn and Johns 1988). The Onaman–Tashota greenstone belt (Figure 19.1) dominates the eastern Wabigoon Subprovince but the relationship of this Archean greenstone-granite area to the rest of the Wabigoon Subprovince was unknown. There might be a correlation between at least part of the Obonga greenstone belt and the central part of the Onaman–Tashota greenstone belt, just north of Humboldt Bay, eastern Lake Nipigon. This is part of the question as to whether regional-scale correlation of age-comparable stratigraphic packages could be made between separate greenstone belts on opposite sides of Lake Nipigon. Such correlation is encouraged from an analysis of the Uchi Subprovince (Stott and Corfu 1991) in northwestern Ontario where several tectonic assemblages could be traced for several hundred kilometres along the length of that subprovince.

Several problems were identified and provided a rationale for this project:

1. Could this greenstone belt, which straddles the breadth of the Wabigoon Subprovince, provide insight on the construction of a superterrane that borders a young (2.7 Ga) postulated accretionary prism to the south and includes volcanic and intrusive rocks more than 3 Ga in age on its northern margin near Armstrong?

2. Can we trace a geological connection between the Obonga greenstone belt (which lies on strike with the Sturgeon Lake greenstone belt containing the Mattabi VMS [volcanogenic massive sulphide] deposits) and the central part of the Onaman–Tashota greenstone belt?

3. Are there major unconformable or tectonic boundaries between young and old volcanic sequences in this region?

4. Can we identify a history of deformation that predates the younger (2.7 Ga) felsic volcanic rocks and sediments?

5. Are all elastic sedimentary sequences equivalent in their tectonic role and timing?

6. What is the metallogenic and alteration history, and mineral potential of this greenstone belt?

This report, based on a 1:50 000 scale geological survey (Figure 19.2) of the central part of the Onaman–Tashota greenstone belt, summarizes the main observations of rock units and structures and addresses some of the questions above. This investigation is part of a 5-year geological survey of the Onaman–Tashota greenstone belt that concludes in 1999. It advances previous detailed mapping of the area (Amukun 1977, 1979, 1980, 1987).

GEOLOGY

Mafic Volcanic Rocks

FIELD DESCRIPTIONS

The massive to pillowsed basaltic flows form the major component of the supracrust in the map area. The rocks are typically very fine grained but coarser grained flows (1 to 2 mm grain size), especially pillowsed facies, form significant mappable units in the vicinity of Willet Lake. The coarser massive flows are easily confused with sill-like sheets of gabbrro that also occur widely in this greenstone belt. The basaltic flows throughout the area display an almost universal absence of amygdules and vesicles. They are only locally accompanied by oxide or sulphide facies iron formation. These flows typically display relatively limited alteration; some sections of basaltic flows display weak to moderate calcite alteration, generally along fractures. This is exemplified south of Gledhill Lake, near the Onaman pluton where calcite occupies pervasive networks of brittle fractures that disappear close to the pluton. Elsewhere, albite-epidote alteration pods occur in some pillowsed sequences but are typically weakly to moderately developed. On a newly constructed logging road, just west of Emily Lake (west of Gzowski Lake), some iron carbonate alteration accompanies disseminated fine grained pyrite in several outcrops of basalt, interbedded sediment, and felsic dikes related to the Gzowski stock nearby. Previously mapped mafic tuffaceous rocks (Amukun 1979) are generally found to be intensely sheared portions of massive basalt or gabbrro (e.g., along the Gzowski Fault northwest of Gzowski Lake). Basaltic
rocks show well-preserved pillows in some areas with useful top indicators especially southwest and west of Gzowski Lake, and in the vicinity of Toronto–Willett lakes. The 4 km thick Willett lakes section of basalt, interlayered with sheets of gabbro, faces and dips steeply north-northeast. This sequence, which contains a quartz porphyrytic rhyolitic tuff on Toronto Lake dated at 2923 Ma (Stott et al. 1998), is deflected around the Deeds stock to the east.

**FIELD RELATIONSHIPS**

One of the objectives of this investigation is to identify distribution of older, pre-2900 Ma rocks in the northern half of the greenstone belt and their field relation to the large areas of felsic volcanic rocks that dominate portions of the map area, especially east of Gzowski Lake. A high angle of intersection occurs between the east trending

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**Figure 19.1.** Location of the Gledhill–Willett lakes map area in the Onaman–Tasbota greenstone belt, east Wabigoon Subprovince. Previously completed map areas, with their map numbers, are also outlined.
basaltic flows and the northeast trending contact with broad rhyolite units near Gledhill Lake and near Venus Lake, just north of the CN Railway tracks (see Figure 19.2). Otherwise, there is little difference in the map area between bedding and foliation trends of either basaltic or rhyolitic flows.

**Intermediate to Felsic Volcanic Rocks**

**FIELD DESCRIPTIONS**

Most of the map area east of the Gzowski stock is occupied by the Venus–Oboshkegan felsic volcanic field (see Figure 19.2) comprising several volcanic centres of quartz and/or feldspar porphyritic rhyolite flows and abundant interlayered zones of autobreccia to hyaloclastic breccia. Breccia zones are interlayered with massive flows and represent hydrothermally brecciated zones within flows and some talus breccia. Debris flows and pyroclastic deposits are much less common. Some sections of feldspar porphyry (e.g., north of Nigig Lake) are so coarse as to suggest possible intrusive origin although these rocks grade upwards to less coarse quartz-feldspar rhyolite with clear flow textures, increasing quartz phenocrysts and capped by breccia. The felsic flows of the Venus–Oboshkegan field are characterized by the prevalence of feldspar phenocrysts in the Lake Ste. Marie volcanic centre and quartz phenocrysts in the Metcalfe–Oboshkegan centre (Stott and Parker 1996). However, in both centres, local, well-exposed sections of rhyolitic flows range in phenocryst composition from dominantly feldspar to quartz + feldspar to quartz. Quartz phenocrysts are commonly round to euhedral. Autobreccia locally shows clear gradation to massive rhyolite, evident southeast of Omakaki Lake and southwest of Gledhill Lake where a broad zone of autobreccia, more than 6 m wide, showing jigsaw-fit texture, grades into massive feldspar porphyritic rhyolite. Varying degrees of iron-enrichment alteration of the breccia matrix is observed. Hornblende (garnet alteration is most typical (Photo 19.1a).

**FIELD RELATIONSHIPS**

In the Venus–Oboshkegan field, the map pattern of felsic volcanic rocks trends northeastwards at a high angle locally to the bedding and foliation trend of basaltic rocks. This suggests that the rhyolite flows extruded unconformably upon an older basement of basaltic flows. In addition, clastic sedimentary rocks (described below) are associated with and derived exclusively from rhyolite and show no evidence of input from mafic rocks although mafic components might have been preserved in the argillaceous units. This is generally consistent with rhyolite deposited above and not interbedded with basalt. The widespread presence of amphibole-garnet-rich matrices in autoclastic

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**Figure 19.2.** Geology of the Gledhill–Willet lakes area.
rhyolite is also a characteristic feature of the Marshall Lake felsic volcanic centre, observed by the authors at Marshall Lake and along the southern flank of the Marshall assemblage, north of Willet Lake.

**Sedimentary Rocks**

**FIELD DESCRIPTIONS**

Sedimentary rocks in the map area are grouped into 3 associations:

1. Thin interflow siltstone and iron formation in basaltic sequences, including an erosional unconformity on Toronto Lake (Berger 1992) underlying a rhyolitic unit dated at 2923 Ma (Stott et al 1998);

2. Gledhill Lake sequence of felsic volcanic-derived arenaceous sandstone to siltstone and argillite, locally interbedded with felsic tuff and resedimented tuff and deposited on the flanks of volcanoes in the Venus–Oboshkegan rhyolite field.

3. Albert Lake sequence of intermediate to felsic volcanic-derived wacke, argillite and arenaceous rocks deposited on the southern flank of the Marshall Lake volcanic centre.

### Interflow Units

Very thinly bedded siltstone to fine grained wacke occur interbedded with basalt in only a few locations, including exposures on the newly constructed logging road west of Gzowski Lake, and accompanied by magnetite-chert iron formation east of Suni Station on the CN Railway tracks.

### Gledhill Lake Clastic Sedimentary Sequence

A sequence of felsic volcanogenic sedimentary deposits is associated with the Venus–Oboshkegan rhyolite field. This assemblage is composed of arenaceous sandstone to siltstone, wacke and argillite. Virtually all sandy to silty bedded units appear to be derived from felsic rocks; grains of coarse sand size or larger are consistently composed of rhyolite (Photos 19.1b and 19.1c). Intermediate or mafic sources are rarely apparent at the north end of and just west of Gledhill Lake where biotite-garnet-rich wacke to argillaceous wacke occur. In those cases the volume of synvolcanic iron enrichment alteration observed in the rhyolitic breccias could provide a mafic source. Otherwise, lithic arenite to lithic wacke dominate this clastic assemblage. Argillaceous beds become more significant in the northern part of Gledhill Lake (locally containing nodular pyrite), and further north where the bedding is

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**Photo 19.1.** Photographs of selected localities in the Gledhill–Willet lakes map area. a) Dark hornblende-garnet matrix in subangular, autobrecciated rhyolite, north of Gledhill Lake. b) Crossbedded lithic arenite derived from a felsic volcanic source with trough of gravel at top of photo, composed of pebbles of rhyolite. Nixon Lake, 4 km northeast of Gledhill Lake. c) Dismembered, transposed beds of arenite and siltstone from Nixon Lake, just north of photo b. d) Monozogranite of Gzowski Pluton intruded by coeval gabbro dike, which is intruded by a granite pegmatite dike and other thin granite dikes.
complexly transposed, and especially along the northern part of the assemblage close to the CN Railway tracks. In the latter area, there is an unusually thick sequence dominated by argillite and argillaceous siltstone. This volume of argillaceous rock is comparable to that observed in the Albert assemblage, on and north of Albert Lake.

**Albert Lake Clastic Sedimentary Sequence**

A sequence of clastic sedimentary rocks occurs on and northwest of Albert Lake, just east of Willet Lake. This sequence is mainly composed of thinly bedded siltstone to silty wacke and argillaceous siltstone. The silty sandstone units are derived from less consistently felsic and more intermediate sources than those in the Gledhill Lake area. Argillaceous units locally contain beds rich in staurolite, which are unaffected by deformation and reflect a late thermal event. Most of the argillaceous rocks lie at the base of the assemblage and comprise schistose dark mudstone with some rusty weathed sections and interbeds of siltstone and chert. As in the case of the Gledhill Lake sequence, these argillaceous rocks form thick sections (> 400 m), which are atypical for Archean greenstone belts in northwestern Ontario.

The Albert Lake volcanogenic sedimentary rocks lie stratigraphically above (south of) the main portion of the Marshall assemblage of rhyolite and andesite (north of) a felsic pyroclastic sequence that appears to terminate the Marshall assemblage volcanism. Some beds of intermediate to felsic tuff or re-deposited tuff (tuffaceous sandstone) occur within the Albert assemblage. Consequently, this sedimentary sequence is coeval with and forms an epiclastic component within the Marshall assemblage. Its intermediate composition indicates derivation by erosion of this felsic to intermediate volcanic pile. The basal argillaceous rocks reflect the initial formation of a muddy basin flanking the paleo-topographically high volcanic edifice. The mudstone was succeeded upwards (southwards) by silty sandstone with increasing tuffaceous interbeds at the top of the sequence where it is gradationally overlain by intermediate tuff and re-deposited tuff that close the history of Marshall Lake felsic volcanism, dated at 2739 ± 1 Ma (U-Pb zircon, Stott, Davis and Parker 1998).

**Mafic Intrusive Rocks**

Gabbroic intrusions occur throughout the map area, generally as apparently synvolcanic, tabular sheets. Significant volumes of gabbroic sheets are interlayered in the basaltic strata of the Toronto–Willet lakes section. However, the most prominent gabbroic intrusion in the area is a late tectonic vertical sheet approximately 400 to 700 m thick that follows the north side of the Gzowski Fault between the Gzowski and Deeds plutons and further southeast. The gabbroic intrusion was marginally affected by the fault. It grades locally to melanocratic gabbro, north and east of Gzowski Pluton.

**Felsic Intrusive Rocks**

**GZOWSKI PLUTON**

The Gzowski pluton is a massive, homogeneous body of potassium-feldspar porphyritic, hornblende granodiorite to monzogranite. Hornblende clots are concentrated locally in some parts of the intrusion. Granitic dikes are not common. At the south end of Gzowski Lake, coarse hornblende-pyroxene gabbro dikes cross cut the magmatic foliation of the granodiorite. These dikes in turn were cross cut by the granodiorite and by thin granite pegmatite dikes (Photo 19.1d). The plutonic fabric is unmetamorphosed and conformable to the shape of the pluton, post-dating any east trending tectonic foliation in the region. However, the late tectonic southeast trending Gzowski Fault shears the pluton at the north end of Gzowski Lake.

**DEEDS PLUTON**

The Deeds pluton, along its southern margin, is bordered by hornblende monzonitic to gabbroic phases. The main portion of the intrusion comprises a hornblende-bearing potassium feldspar porphyritic monzogranite to granodiorite. The reddish-brown potassium feldspar is uniformly mantled by white plagioclase, similar to “rapakivi texture”. This euhedral rapakivi phase appears to be widespread in the pluton and is generally characterized by widely spaced joints. The dominant joints are subhorizontal and typically spaced 1 to 3 m apart. The intrusion is generally free of hornblende clots, inclusions and dikes. These features are of interest to investigators of dimension stone resources. Along part of its northwestern margin, the Deeds pluton is bordered by a tabular lens of tonalite gneiss that appears to have been dragged up from the lower crust during emplacement of the pluton. It appears to provide a window of lower crust directly underlying the greenstone belt.

**GLEDHILL STOCK**

The Gledhill Stock is composed of potassium-feldspar megacrystic, biotite to hornblende-biotite monzogranite to granodiorite. It contains widespread small clots of hornblende or biotitized hornblende crystals. Locally there are widespread larger inclusions of gabbro to diorite, and marginal to the pluton there is a local gabbro phase. The stock is homogeneous, typically equigranular but with variably distributed potassium-feldspar megacrysts overprinting the rock fabric. The intrusion is a late tectonic body, similar to the Gzowski pluton; it imposed a contact thermal aureole on adjacent garnetiferous wacke near Gledhill Lake. The felsic mineralogy is unmetamorphosed and the foliation and lineation are interpreted as primary magmatic fabrics that conform to the shape of the intrusion and show no relation to the external, tectonic fabric in the volcanic rocks. These fabric observations apply also to the Gzowski and Deeds plutons.
STRUCTURAL GEOLOGY

General Facing Directions of Basaltic Strata

South and east of Gledhill Lake, the basaltic strata appear to face southwards, comparable to facing directions on strike in the Metcalfe Lake area (Stott and Parker 1996). West and southwest of Gzowski Lake, stratigraphic top directions face southwards but become consistently north facing towards Deeds Lake pluton and further north in the Toronto–Willet lakes area. The southern flank of the Marshall Lake felsic volcanic assemblage (see Figure 19.2) faces south, based on facing directions in this assemblage east of Willet Lake and a southward (upward) increase in hornblende-garnet alteration to a ferruginous chert horizon, which provided a hydrothermal cap to the underlying alteration system. The Toronto–Willet lakes section of basalt includes a unit of felsic tuff dated at 2923 Ma (Stott, Davis and Parker 1998) and faces northwards toward the south-facing 2739 Ma Marshall Lake assemblage.

Bedding and foliation in basaltic flows throughout the map area typically dip 50° to 80° northwards. Tectonic strain in basalt west of Gzowski Lake and north of Willet Lake is relatively weak to moderate. Deformation in the rhyolitic flows is likewise relatively weak with localized brittle fracturing a common response to shortening across the belt.

The following episodes of deformation pertain to structures and lithologic relationships observed in the Gledhill–Willet lakes area; they are not yet correlated with the structures observed elsewhere in the greenstone belt.

D1 Structures

East of Gledhill Lake, the foliation in the basaltic rocks strike west-northwest and dip moderately to steeply northwards, contrasting sharply with the southwest striking foliation and bedding of sedimentary and felsic volcanic rocks on and west of Gledhill Lake. There is virtually no evidence of basaltic source for the sedimentary rocks and the regional pattern of contacts of rhyolite and epilastic rocks transects the trend of basalt flows. This, plus the contrast in foliation and bedding trends between the basalt and the rhyolite and associated epilastic rocks is interpreted to reflect the presence of a deformation of the basaltic rocks that predates the deposition and erosion of the Venus–Oboshkegan rhyolite. D1 is interpreted to be a regional north-south shortening event that affected the basaltic rocks. The evidence for D1 is indirect; no unconformities were observed. The presence of pre-2.9 Ga volcanic rocks in the northern half of the Onaman–Tashota greenstone belt (Stott, Davis and Parker 1998) indicates that there should be a tectonic history that predates the 2739 Ma Marshall Lake volcanic rocks and similar rhyolitic sequences in the Venus–Oboshkegan field. However, overprinting by the more pervasive D2 fabric obscures the evidence for similar-oriented pre-D2 structures.

D2 Structures

In many areas, there is a close correspondence in foliation trend in both basalt and rhyolite near their mutual southwest trending contacts indicating that subsequent D2 shortening affected all supracrustal rocks in the region. The sedimentary rocks on Gledhill Lake form a southwest trending syncline that closes to the southwest and corresponds to the dominant southwest trending D2 foliation of the felsic volcanic rocks. This syncline becomes a more complex structure further north where structural facing directions vary locally and there is evidence of overturned folds. The sedimentary rocks display open to tightly transposed folds even across individual outcrops. The D2 foliation dips moderately to steeply northwest in the Gledhill–Gzowski lakes area, and in the Willet Lake area it dips consistently north to north-northeast.

D3 Structures

The Gzowski pluton has been dated at 2698 ± 1 Ma (U–Pb zircon age, Stott, Davis and Parker 1998) and is interpreted to be coeval with other late tectonic felsic plutons in this area. Each of these late plutons has produced a strain aureole upon the enveloping supracrustal rocks. The deflection of the regional foliation and mineral and/or shape lineation near these plutons marks the outer extent of the penetrative strain aureole. This locally imposed deformation during pluton emplacement represents the D3 episode of structural development in the greenstone belt.

D4 Structures

FAULT ZONES

Two major faults have been observed that split from a major fault; the Gzowski Fault and Robinson Fault (see Figure 19.2) trend southeast, parallel to each other. They bifurcate from a fault that continues northwestwards into and beyond Toronto Lake and joining the northwest-trending, right-handed transcurrent Pushkokogan Fault (see figure 7.3 of Breaks 1991) along the subprovince boundary. The Gzowski Fault follows closely along the southern flank of a major, tabular gabbroic intrusion that locally forms steep cliffs. The effect of this fault on adjacent rocks is rather limited. Apart from sheared gabbro on the logging road west of Gzowski Lake, only local shearing and kink banding accompanies the fault in argillite, for example south of the CN Railway tracks.

The Robinson fault zone has been mapped in the Robinson pluton where it is readily accessible. The fault zone is over 400 m wide on Willet Lake and trends southeast towards the south end of Gzowski Lake where it deflects to the east-southeast (see Figure 19.2). This fault zone is well developed in the Robinson pluton, showing clear dextral, transcurrent shear sense indicators, including asymmetrical extensional shear bands and wings on feldspars similar to indicators described by Passchier and Trouw (1995). The fault continues into the greenstone belt where its presence is apparent on a Landsat image of the region. Its presence near Emily Lake, west of Gzowski
Lake is also suggested by the marked local increase in fine-grained, disseminated pyrite and iron carbonate alteration affecting several rock types exposed on a new logging road.

**ECONOMIC GEOLOGY**

Substantial hydrothermal alteration of the Marshall Lake felsic volcanic pile is evident not only at the main deposits of volcanic-hosted massive sulphides west of Marshall Lake but also along the southern margin of the rhyolitic flows. Hornblende-garnet alteration increased upwards (southwards) in the rhyolite and is most pronounced under a cap of iron formation, north of Willet Lake. Felsic tuff overlying the ferruginous chert shows little evidence of this alteration. This part of the Marshall Lake felsic volcanic assemblage lies higher in the stratigraphy than the original VMS discoveries. A reconsideration of the Marshall felsic volcanic assemblage as host for additional VMS deposits might benefit from a geochemical survey along the southern and eastern portions of this assemblage where stratigraphically continuous cap rocks of chert have been observed. Part of the southermmost flank of the Marshall assemblage will be accessible by a logging road system in the future as evident from the 1997 to 2017 Forest Management Plan for Domtar Forest Products Limited in Red Rock. The road system, presently leading to the Deeds pluton, is under construction by Buchanan Forest Products Limited in Jellicoe.

East of Gzowski Lake, some sulphide occurrences have been reported (e.g., Royal Crown occurrence where grab samples, reported by Noranda Exploration Co. Ltd., returned up to 9.99% Zn and 1.91% Pb with typical assays averaging 0.5%: AFRI File 42L06SW0007, Assessment Files, Resident Geologist’s office, Thunder Bay). Not far from these occurrences, sulphide facies iron formation is interbedded with felsic volcanic flows and pyroclastic rocks and forms discontinuous marker horizons.

Relatively few areas display iron carbonate-sulphide enrichment associated with high strain zones. One area of interest has been recently uncovered and staked during logging road construction by employees of Buchanan Forest Products Ltd. This area lies west of Emily Lake (west of Gzowski Lake) where several outcrops of basalt and some thinly bedded interflow sediment contain iron carbonate and disseminated pyrite. Feldspar porphyry dikes, apparently related to the Gzowski pluton nearby also contain disseminated pyrite and iron carbonate alteration. This section of the road, with widespread iron carbonate alteration, lies on strike with the northwest trending Robinson fault (see Figure 19.2) that can be traced into the Robinson pluton to the west. Although staking of this zone of alteration has taken place along the new road, it should be noted that this area of the greenstone belt, west of Gzowski Lake and north of Tashiota Lake, contains relatively little record in the assessment files of previous exploration. With a significant fault zone trending through the area accompanied by newly discovered alteration, exploration for gold could test the less accessible extent of the Robinson fault zone, northwestward towards the western margin of the belt near Dwight Lake.

Owing to the development of a logging road system over the next few years that will access a significant part of the Deeds pluton, this intrusion should be investigated for its building stone potential. Large parts of the intrusion are homogeneous with distinctive euhedral, rapakivi-like feldspars (not commonly found in northwestern Ontario), widely spaced joints, and few if any blemishes and dikes.

**TECTONIC INTERPRETATION**

U-Pb zircon age determinations from the northern half of the Onaman–Tashiota greenstone belt (Figure 1 from Stott, Davis and Parker 1998) show that the belt includes volcanic sequences ranging in age from 3056 Ma to 2923 Ma. It appears that a substantial part of the northern half of the greenstone belt formed an older, dominantly basaltic, continental base upon which rhyolitic volcanoes formed (e.g., 2739 my old Marshall assemblage). The earliest structures are fociations preserved in basalt east of Gledhill Lake. A field of rhyolite flows shows northeast trending contacts against east trending basaltic flows that suggest the D1 tectonic fabric in the basalt predates the growth of rhyolite volcanoes and erosion of rhyolite into epiclastic basins. Given the volume and generally fine grain size of clastic sediment deposited on the flanks of the rhyolite domes, and the lack of coarse products of mass wasting processes (talus and debris avalanche deposits), the felsic volcanic rocks are probably shallow-water to emergent (McPhee, Doyle and Allen 1993, p.97). The disconformable contact between 2739 million-year-old felsic to intermediate volcanic rocks of the Marshall assemblage and 2923 million-year-old mafic flows of the Toronto–Willet lakes sequence provides further evidence of a long and episodic tectonic history. A subsequent D2 period of northwest to north-directed shortening deformed felsic volcanic flows and sedimentary rock in the Venus–Oboshkegan area and reactivated D1 structures in the underlying basalt. The D2 deformation is the most widely recorded across the belt. It is superimposed by more intensely flattened and steeply lineated strain aureoles in the supracrustal rocks that envelop late tectonic intrusions such as the 2698 Ma Gzowski pluton and the larger Deeds pluton. This episode of D3 deformation, produced during emplacement of the plutons, is followed by southeast trending, right-handed transcurrent faults that merge in the vicinity of Willet Lake and apparently join with the Pashkokogan Fault along the northern margin of the greenstone belt. These fault zones are probably a late response to the northwest directed transpressive deformation that produced well-developed folds and stretching lineations plunging moderately (30°) eastward throughout the Marshall Lake area in the northern part of the belt.

**REFERENCES**


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20. Project Unit 95–14. Geology of the Ignace and Pekagoning Lake Areas, Central Wabigoon Subprovince

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BACKGROUND

The Wabigoon Subprovince is a complex, large-scale part of the western Superior Province. Composed essentially of greenstone belts and 77\% felsic plutonic rocks (Goodwin 1976), the Wabigoon Subprovince extends as a broad, belt-like domain 1000 km easterly from the Williston Basin in Manitoba to the Hudson Bay Lowlands. The bulk of Wabigoon greenstone and plutonic suites have ages spanning a relatively brief period from 2730 to 2685 Ma (e.g., Davis and Edwards 1986), which is broadly consistent with interpretations (e.g., Williams 1990) that the Wabigoon Subprovince rapidly developed as a magmatic arc and accreted to the Superior craton. The accumulation of geochronologic data (e.g., Davis and Jackson 1988; Davis, Sutcliffe and Trowell 1988; Davis and Moore 1991 and Tomlinson et al. 1997) has shown, however, that enclaves of older material are present and that Wabigoon Subprovince geology and tectonic evolution are complex. For example, pre-2900 Ma plutonic and supracrustal rocks occur at Marmion Lake, near Obonga Lake, and at Caribou Lake. These fragments of Mesoproterozoic crust are interspersed with Neoarchean greenstone belts and intrusions and are scattered diagonally across the Wabigoon Subprovince from Atikokan to Armstrong—the Wabigoon diapiric axis of Edwards and Sutcliffe (1980).

The pattern of old and young Archean crustal domains is an important but poorly understood aspect of Wabigoon Subprovince geology. The crustal elements are important because their distribution and relationships provide clues as to how the Wabigoon Subprovince formed and evolved. Further, definition of the old and young crustal domains can constrain exploration for mineral commodities such as base and precious metals, many of which show a strong affinity to rocks of a certain age and tectonic environment (Fyon et al. 1992). In the same context, the crustal elements of the Wabigoon Subprovince are poorly understood because their extent and boundaries are not well established. It is also unclear whether the domains of older rocks represent discrete fragments of early crust or are, for example, exposed parts of a large and extensive old rock mass possibly overlain by younger material.

In 1995, regional mapping at a scale of 1:50 000 was initiated to address the relationships between old and young crustal elements of the south-central Wabigoon Subprovince. Mapping began near Atikokan (Figure 20.1) and has subsequently generated 1 or 2 maps annually with the Ignace and Pekagoning Lake areas marking the 1998 contributions. Geology in the area of Stormy Lake and Upper Manitou Lake is compiled from Blackburn et al. (1991).

Although greenstone sequences between Mine Centre and Atikokan have been extensively studied (see summary in Stone, Kamineni and Jackson 1992), northern parts of the area have received less attention. Previous mapping in the Ignace and Pekagoning Lake areas consists largely of reconnaissance surveys and compilations by Tanton (1938), Pye and Fenwick (1963), Davies (1965), Sage et al. (1974), Schwerdtner (1976) and McWilliams (1998).

Recently, the old–young theme in Superior Province geology has become the focus of a National Mapping Program (NATMAP) of the Geological Survey of Canada (GSC) (Percival et al., this volume). The present study, integrated with research by GSC staff in the Savant Lake–Obonga Lake area, provides an opportunity to elucidate the major crustal elements in a large part of the central Wabigoon Subprovince. The federal and provincial projects also provide a geologic framework for interpretation of the LITHOPROBE seismic-refraction lines that were acquired in 1996 along a north-south corridor transecting the Wabigoon Subprovince east of the present area.

REGIONAL GEOLOGIC SETTING

The combined mapping of the present project (1995 to 1998) is broadly centred on the Irene–Eltrut lakes complex (Schwerdtner 1976) at the southwest end of the Wabigoon diapiric axis. The Irene–Eltrut lakes complex consists essentially of biotite tonalite, tonalite gneiss and thin greenstone slivers intruded by biotite granite, such as in the White Otter batholith (see Stone and Hallé 1996 and Figure 20.1). Southward, the tonalite and gneiss occupy concave embayments into 2727 to 2722 Ma greenstone sequences (Davis, Poulsen and Kamo 1989) extending from Rainy Lake through Mine Centre to Calm Lake at the south boundary of the Wabigoon Subprovince. These greenstone sequences are locally juxtaposed across the Quetico fault with metasedimentary rocks of the Quetico Subprovince.
East of Atikokan, greenstone belts wrap around a large biotite tonalite batholith at Marmion Lake. The Marmion Lake batholith and greenstone belts at Finlayson and Lumby lakes (see Figure 20.1) have pre-2.9 Ga ages and are the main elements of the Mesoarchean domain in the south-central Wabigoon Subprovince.

North of the Irene–Eltrut lakes complex, such as in the Ignace and Pekagoning lakes area (see Figure 20.1), felsic plutonic rocks are interfingered with variably curved, tapered and bifurcated Archean greenstone belts and slivers. This greenstone material forms distal extensions of the post-2775 Ma Kakagi Lake–Savant Lake volcanic belt—an extensive network of supracrustal sequences in the northern Wabigoon Subprovince (Davis, Sutcliffe and Trowell 1988; Blackburn et al. 1991).

In addition to biotite tonalite and biotite granite, suites of felsic plutonic rocks, including the hornblende suite, sanukitoid suite and peraluminous granite suite, are mapped and described further below. The latter felsic plutonic suites typically underlie about 10% of the combined map area (see Figure 20.1).

GEOLOGY—IGNACE AND PEKAGONING LAKE

Supracrustal Rocks

Supracrustal rocks are concentrated in the Raleigh Lake greenstone belt, which attains a width of 10 km to the west of Ignace (Figure 20.2). The Raleigh Lake greenstone belt is intruded by several oval plutons of biotite granite and curves eastward tapering to 1 km in width. A second arm of the belt curves abruptly west and probably joins with greenstone sequences at Upper Manitou Lake (see Figure 20.1). Greenstone belts are much less common in the Pekagoning Lake area (Figure 20.3), consisting of thin (<1 km wide) units in gneisses and at the margins of sanukitoid intrusions.

Figure 20.1. Geology of the south-central Wabigoon Subprovince. The Ignace and Pekagoning Lake areas are outlined.
Mafic metavolcanic rocks that probably originated as basaltic submarine lava flows are the most abundant rock in greenstone belts—particularly thin parts of belts. At Raleigh Lake, pillows are commonly present, whereas higher levels of strain and metamorphism have obliterated primary structures in narrow parts of belts in the Ignace and Pekagoning Lake areas. Thin, pyritic iron formation units are typically present with mafic flows. Ultramafic flows and intrusions appear to be absent, whereas gabbroic sills and stocks are identified at scattered localities in mafic parts of greenstone belts.

Intermediate to felsic metavolcanic rocks comprising dacitic to rhyolitic flows, tuffs and breccias occur in the area of Raleigh Lake (see Figure 20.2). South and southeast of Raleigh Lake, intermediate to felsic metavol-

Figure 20.2. Geology of the Ignace area.
canic rocks are gneissose due to moderate levels of strain, but appear to be mainly flows and tuffs. Coarse breccias are concentrated west of Raleigh Lake. Epiclastic rocks, including thin-bedded sandstone, are commonly associated with flows and tuffs.

Metasedimentary rocks are concentrated at Balmoral Lake (see Figure 20.2) and Smirch Lake (see Figure 20.3). Although highly migmatized, metasediments at Balmoral Lake appear to be interbedded sandstone-siltstone sequences possibly deposited in deep basins by turbidity currents. The thin sandstone unit at Smirch Lake is probably the southern extension of the Stormy Lake Group (Blackburn et al. 1991) that occurs north of the Pekagoning Lake area.

**Felsic Plutonic Rocks**

Six suites of felsic plutonic rocks are mapped in the Ignace and Pekagoning Lake areas. Crosscutting relations indicate that tonalite gneisses of the gneissic suite are among the oldest felsic plutonic rocks. These occur as irregular to belt-like units mainly in the area between Eltrut and Pekagoning lakes (see Figure 20.3). Gneisses are heterogeneous layered rocks locally varying in composition from tonalite and granodiorite through diorite to basalt. At the

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**Figure 20.3.** Geology of the Pekagoning Lake area.
felsic end of their compositional spectrum, gneisses are
gradational to biotite tonalite and at the mafic end of their
spectrum, gneisses grade to amphibolites of volcanic
origin. Gneisses show complex textures. For example,
gneissic layers are commonly folded and can be strongly
attenuated and boudinaged (straight gneisses) or linedated
(pencil gneisses) and can exhibit mylonitic textures.

Biotite tonalite is typically a coarse-grained, grey and
moderately foliated rock comprising about 20% of the
study areas. At Electra Lake (see Figure 20.3), biotite
tonalite occurs in oval to curved and elongated units that
are conformable with enveloping gneisses and possibly
represent deformed plutons. Elsewhere, such as in the
northwest corner of Figure 20.3, biotite tonalite forms
extensive intrusive bodies. Biotite tonalite appears to have
intruded and metamorphosed greenstone sequences, par-

ticularly narrow mafic metavolcanic units in the northwest
corner of Figure 20.3.

Two-mica granite of the peraluminous or S-type
granite suite is typically very coarse grained to pegmatitic,
white and contains one or both of biotite and muscovite
as well as accessory garnet, tourmaline and apatite. Two-
mica granite occurs as thin units associated with metasedi-
mentary rocks at Balmoral Lake and at the margin of the
Revell batholith (see Figure 20.2). Dikes of two-mica
granite in mafic metavolcanic rocks west of Raleigh Lake
are mineralized with rare metals (Breaks 1993).

Rocks of the hornblende suite are coarse, granular and
mesocratic, grading compositionally from biotite-hornblende
tonalite to granite. Hornblende granites are distin-
guished by large (0.01 to 0.05 m) blocky potassium-feld-
spar megacrysts. The hornblende suite is found as inclusions in biotite granite batholiths, such as the Indian
Lake batholith (see Figure 20.2), and as narrow, commonly
strongly foliated units at the margin of the White Otter
batholith. Rocks of the hornblende suite appear to be
intermediate in age between those of the biotite tonalite
suite and the biotite granite suite.

Biotite granite is widespread, underlying approxi-
mately half of the area (see Figures 20.2 and 20.3). Biotite
granite is typically coarse grained, inequigranular and
leucocratic, containing a few percent biotite and subequal
proportions of quartz, plagioclase and potassium-feldspar.
It occurs in forms ranging from dikes and small bodies
intruding most other rocks to large batholiths such as the
White Otter and Indian Lake batholiths. With the probable
exception of the sanukitoid suite, biotite granite is the
youngest major intrusive rock of the area. The White Otter
batholith has a U-Pb zircon age of 2685 ± 2 Ma (D. Davis,

The sanukitoid suite is represented by the Entwine
stock and an unnamed intrusion, both of which are partly
mapped in the Pekagoning Lake area (see Figure 20.3).
The intrusions are composed of a very coarse-grained,
quartz-undersaturated rock compositionally variable from
diorite and monzodiorite to monzonite. From 10 to 30% biotite, hornblende and clinoptyroxene are typically pres-

ten. Ultramafic rocks that appear on the basis of field
observation to be hornblendites and pyroxenites occur as
inclusions and deformed dikes in sanukitoid intrusions and
as rare dikes in country rocks near sanukitoid intrusions.
An ultramafic plug of approximately 2 km diameter
intrudes the Entwine stock (see Figure 20.3).

A few diabase dikes, provisionally correlated with the
1100 Ma Keweenaunawan swarm, transect the area northwes-
terly.

Structure

The youngest intrusive rocks, such as those of the biotite
granite and sanukitoid suites, show weak foliations
defined by alignment of primary igneous minerals such as
feldspars, biotite and amphibole. These probably magmat-
ic foliations tend to define concentric zoned patterns
within oval intrusions. Most other rock types, including
supracrustal sequences, biotite tonalite and gneisses, show
more intensely developed, predominantly planar and
locally strongly linedated fabric. In these rocks, the fabric
is typically defined by alignment of metamorphic minerals
and recrystallized aggregates of minerals and is probably
tectonic in origin. For the most part, the dominant planar
fabric elements follow complex curved trajectories, main-
taining parallelism with nearby geologic contacts.

The structural data provide evidence that late intru-
sions locally overlie supracrustal rocks. For example, fabric in the southern Raleigh Lake greenstone belt,
including the metasediments at Balmoral Lake, dips south
beneath the White Otter batholith (see Figure 20.2). North
of Pekagoning Lake (see Figure 20.3), foliations in
gneisses and greenstone slivers dip north at the south
contact of a sanukitoid intrusion. These observations
imply that the late intrusions are sheet-like in form, with
margins at least locally overlapping supracrustal country
rocks.

Generally high levels of strain have obliterated primary structures and preclude a meaningful assessment of
the structure of greenstone belts. A few, mainly
southward younging directions were derived from pillow
shapes in less deformed parts of the Raleigh Lake belt. The
highly curved contacts, such as between intermediate to
felsic and mafic volcanics, imply, however, that green-
stone strata of the Raleigh Lake belt may be complexly
folded.

A variety of shear zones and faults are identified. For
example, greenstone strata northwest of Raleigh Lake (see
Figure 20.2) show strong southeast-striking, steep south-
west-dipping foliations. This well-foliated zone may mark
the extension of a broad shear zone identified by Parker
(1989) in greenstone sequences northwest of the present
area. Rocks at the northern contact of the metasedimentary
unit at Balmoral Lake are also well foliated and intruded by
two-mica granite. This implies that the sedimentary
sequence is in fault contact with metavolcanic rocks of the
Raleigh Lake greenstone belt.

The oval ultramafic plug in the Entwine stock (see Figure 20.3) is cut by narrow faults associated with
northeasterly striking lineaments. The faults show subho-
rizontal lineations and represent late brittle stages of
deforation of the area.
Metamorphism

Supracrustal rocks show field evidence of moderate to high metamorphism characteristic of amphibole-bearing mineral assemblages in basaltic rocks. An assemblage of garnet + amphibolite + feldspar ± biotite is fairly widespread in the major part of the greenstone belt at Raleigh Lake. Amphibolites in narrow arms of greenstone belts and in gneisses locally show textural evidence of garnet converted to feldspar. An assemblage of garnet + biotite + sillimanite + feldspar + quartz is observed in metasedimentary migmatites in the area of Balmoral Lake.

Economic Geology

Aside from periodic extraction of grey granite from the Butler Quarry west of Ignace (see Figure 20.2), no mineral production is known from the present area. Satterly (1941) briefly described molybdenite occurrences in narrow extensions of the Raleigh Lake greenstone belt south of Ignace. Blackburn and Hinz (1996) reviewed recent mineral exploration, which is concentrated mainly in the Raleigh Lake area. Breaks (1993) described a rare-element pegmatite field (1.5 km wide by 4 km long) hosted by mafic and intermediate metavolcanic rocks west of Raleigh Lake. Within the pegmatite field, low to intermediate dip, white muscovite potassic pegmatite dikes are mineralized with beryl and spodumene and accessory columbite, tantalite, microtite, cleavelandite and holmquistite (Breaks 1993).

During the present survey, grab samples were collected and assayed for gold, base metals and platinum group elements. Sample sites in the Ignace area (see Figure 20.2) mainly comprise rusty zones, which probably represent metamorphosed iron formation units in metavolcanic rocks. The assays (Table 20.1) returned low values of gold, copper and zinc. In the Pekagoning Lake area (see Figure 20.3), ultramafic inclusions, dikes and the ultramafic stock associated with sanukitoid intrusions were sampled and assayed for platinum group elements. These samples (Table 20.2) returned negligible values.

Despite low values obtained from grab samples, the shear zone extending northwest of Raleigh Lake appears to be a favourable site for gold mineralization. Evidence for greenschist alteration associated with fluid movement through the zone may provide guidance in focussing exploration (see Parker 1989).

Coarse intermediate to felsic pyroclastic rocks west of Raleigh Lake show fairly widespread development of garnet and amphibole, which may indicate premetamorphic hydrothermal alteration. Following the recommendation of Blackburn and Hinz (1996), this area appears to show the best potential for base metal mineralization.

New occurrences of two-mica granite were identified at the northeast side of the Revelst Batholith and in metasedimentary rocks at Balmoral Lake (see Figure 20.2). These localities may provide an environment favourable to rare-metal mineralization. The ultramafic plug in the Entwine stock (see Figure 20.3) should be prospected for sulphide zones and further evaluated for platinum mineralization.

SUMMARY AND REGIONAL IMPLICATIONS

Rocks representative of both the old and young domains of the Wabigoon Subprovince are present in the combined multiyear map area (see Figure 20.1). The biotite tonalite batholith at Marmion Lake, greenstone sequences at Lumby Lake and Finlayson Lake, and tonalites and gneisses northwest of Finlayson Lake comprise a large and fairly coherent domain of pre-2.9 Ga crust whose eastern extent is not yet defined. In contrast, greenstone sequences west of Calm Lake and in the Ignace and Pekagoning Lake area are thought to be post-2775 Ma in age either by direct dating or by virtue of their continuity with dated parts of the Kakagi Lake–Savant Lake volcanic belt. Hence, the presence of both Mesoarchean and Neoarchean domains naturally leads to the problem of where to establish their mutual boundary.

A simple interpretation would hold that biotite tonalite intrusions and tonalite gneisses northwest of Finlayson Lake are representative of the entire Irene–Eltrut lakes complex. In this case, the Irene–Eltrut lakes complex (with exception of the biotite granite and sanukitoid suites) constitutes a fairly extensive domain of old plutonic rocks. Potentially young greenstone sequences to the west would have been deposited on top of or tectonically emplaced onto the Irene–Eltrut lakes complex.

Crosscutting relations suggest that biotite tonalite in the Pekagoning Lake area postdates greenstone sequences and alludes to a more complex range of ages for tonalite plutons than in the above scenario. Further, geochronologic studies (Corfu and Stone, in press) showed that multiple generations of biotite tonalite spanning up to 300 Ma are present in the Berens River area of northwest Ontario. By analogy, the Irene–Eltrut lakes area may have had a protracted magmatic history, possibly resulting in a complex geometric distribution of old and young biotite tonalite intrusions and gneissic belts. Hence, geochronology combined with regional mapping will be required to establish the extent of the old crustal domain in tonalitic and gneissic areas of the Irene–Eltrut lakes complex.

Perhaps of greater importance than the problem of establishing boundaries of old domains are more fundamental questions regarding the overall form of the pre-2.9 Ga crust. For example, do each of the old domains represent discrete and separate blocks or are they merely the exposed parts of a larger continuum? Tectonic models, which incorporate both old and young components of the Wabigoon Subprovince (e.g., Card 1990), have depicted the old domains as microcontinental fragments within otherwise juvenile arc material. In this context, the old domains represent large, fairly discrete blocks. Old material within the blocks would have been intruded and locally redistributed by young magmas and tectonism, but otherwise would be enveloped by juvenile material both at surface and at depth.
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<th>Cu (ppm)</th>
<th>Pb (ppm)</th>
<th>Zn (ppm)</th>
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An alternative model to the block-like distribution of old material arises from structural evidence that the youngest intrusions overlie greenstone sequences. For example, margins of the White Otter batholith appear to be sheet-like in form and to have been locally emplaced over top of the Raleigh Lake belt. This leads to speculation that intrusions such as the White Otter batholith may be overall sheet-like bodies on top of older crustal material. A process of crustal growth whereby progressively younger felsic magmas punctured and overplated or intraplated old material may have been operative. In this scenario, old crust may be more extensive at depth than is indicated by surface exposures.

A variety of tools can be used to assess the distribution of old crust. For example, the presence of inherited zircons in felsic rocks may indicate melting of older crust during magmatism and the LITHOPROBE seismic profiles may be capable of imaging old and young crustal segments. The size range and distribution of old material may also provide insight on whether old crust occurs as blocks or as a substratum. If early material is broadly distributed at depth in the southern Wabigoon Subprovince, one would expect to see both large and small-scale fragments scattered fairly widely at surface. For example, old material might occur as xenoliths in young batholiths or at the interstices of younger intrusions. Such a broad distribution and wide range size of old fragments might not be expected if the old material occurs as discrete microcontinental fragments. In any case, a combination of regional mapping and other analytical tools may be required to resolve fundamental problems of Wabigoon Subprovince geology.

REFERENCES


INTRODUCTION

Geologic mapping in the Ellard Lake and Pasquatchai River areas marks the continuation of a multiyear survey of the northern Superior Province of Ontario. The combined project area is situated about 350 km north of Red Lake, Ontario and includes an area of approximately 13,000 km$^2$ mapped since 1995. The Stull Lake area was mapped at the onset of the project and was revisited in subsequent field seasons with the result that several important additions and revisions to Stull Lake geology are described here.

The northern Superior Province lying adjacent to the Hudson Bay Lowlands is a remote area of the province. Hindered by difficult and expensive access and lack of services, the area has had only limited geologic mapping (Downie 1937; Meen 1937; Satterly 1937; Bennett and Riley 1969) and little attention from exploration companies and prospectors. Hence, a major goal of this project is to provide a new set of maps in aid of mineral exploration and an improved understanding of the regional geology and tectonic evolution. In the latter instance, the study area lies north of the North Caribou Terrane — a domain of rocks older than 2.9 Ga that may represent a cratonic core of the Superior Province. Various subprovinces are thought to have rapidly accreted to the south side of the North Caribou Terrane at about 2.7 Ga (e.g., Thurston, Osmani and Stone 1991); however, much less is known about evolution of the northern Superior Province.

The current project has generated from 1 to 3 maps at a scale of 1:50,000 in each of the past 3 years. Many of the map areas abut the Manitoba border with the result that field work and cross-border geologic compilation in these areas is done in conjunction with surveys of Manitoba Energy and Mines (e.g., Corkery, Skulski and Whalen 1997).

The project is also integrated with the Western Superior NATMAP Project of the Geological Survey of Canada (see Percival et al., this volume). J. Whalen of the GSC is currently engaged in geochronologic and isotopic analysis of plutonic rock samples from the area.

The study area (Figure 21.1) is elongate northerly over a distance of 120 km and provides a geologic transect of four east-southeast-striking greenstone belts at Sachigo Lake, Stull Lake, Ellard Lake and Yelling Lake, as well as intervening felsic plutonic domains. An overview of the regional geology, including a description of the greenstone belts and plutonic suites, is given by Stone and Hallé (1997). Mapping in the current season has traced eastern extensions of the greenstone belts at Ellard and Yelling Lakes and provided revisions to geology of the Stull Lake greenstone belt.

EELLARD LAKE AND PASQUATCHAI RIVER AREAS

Geology

The greenstone belt at Ellard Lake (Figure 21.2) is part of a discontinuous chain of greenstone belts extending 300 km west-northwesterly at the north margin of the Superior Province and locally attaining a width of 10 km. Westward at Little Stull Lake (see Figure 21.1) the Ellard Lake greenstone belt merges with the Stull Lake greenstone belt. A much narrower (1 to 2 km wide) greenstone belt is traced east-southeasterly through Yelling Lake and appears to curve eastward and bifurcate into several narrow arms in the area of the Thorne River (see Figure 21.2; see also Bennett and Riley 1967).

Mafic metavolcanic rocks are the main component of the Ellard Lake greenstone belt and constitute nearly all the Yelling Lake greenstone belt. Pillows are widely developed and imply that the mafic metavolcanic rocks originated as submarine basalt flows. Mafic metavolcanic rocks have been transformed into schists where affected by faults and into amphibole gneisses by combined strain and metamorphism in places such as at the margins of greenstone belts. Mafic metavolcanic sequences are commonly intruded by gabbro dikes and sills. A possible ultramafic flow is identified on the basis of field characteristics (soft dark dense rock) about midway between Lacey and Sherman Lakes (see Figure 21.2).

Intermediate to felsic metavolcanic rocks that typically show fragmental textures occur south of Ellard Lake and in a tapered and bifurcated unit east of Lacey Lake (see Figure 21.2). The latter unit includes sill-like intrusions of quartz-feldspar porphyry. The fragmental metavolcanic units con-
tain thin horizons (up to 100 m wide) of epiclastic rocks including sandstone-siltstone sequences and rare conglomerate. South of Ellard Lake, fragmental metavolcanic rocks are in contact with an extensive (1 km wide) sandstone-siltstone sequence. Conglomerates with a wide variety of plutonic and supracrustal clasts are exposed about 3 km west of Ellard Lake. It is unclear on the basis of field characteristics whether the conglomerates are a coarse, proximal facies of the sandstone-siltstone sequence or a younger Timiskaming-type sequence.

Felsic plutonic areas of the northern Superior Province can be divided into essentially the same suites of felsic plutonic rocks that are found in the Wabigoon Subprovince (see Stone et al., this volume). These include the biotite tonalite, tonalite gneiss, hornblende tonalite to granite, biotite granite and monzodiorite (sanukitoid) suites. The suites have characteristics such as texture, mineralogy, composition and relative ages that are common to both areas. Distinctions occur mainly in the overall form, relative proportions and possibly absolute ages of various suites in the northern and southern areas, respectively.

Crosscutting relations indicate that biotite tonalite of the tonalite suite is among the older plutonic rocks and is fairly widespread, occurring as broad units of overall west northwest trend (see Figure 21.2). Biotite tonalite is heterogeneous, varying from white to grey and from massive through

![Figure 21.1 Geology of the northern Superior Province, Sachigo, Stull and Yelling lakes area, Ontario and Manitoba.](image-url)
foliated to weakly gneissic varieties. Compositions range from tonalite to granodiorite usually with 5 to 10% biotite. Weakly gneissic varieties of biotite tonalite grade perceptibly to tonalite gneiss that occurs as small, irregular units of lesser area here than in the Wabigoon Subprovince. Tonalite gneisses are heterogeneous layered rocks typically more mafic than biotite tonalite. Gneisses at Rieder Lake (see Figure 21.2) contain up to 50% bands of amphibolite and are cut by conformable and disconformable dikes of granite.

Intermediate-age plutonic rocks are hornblende tonalite to granite of the hornblende suite that occur as west-northwesterly, elongate units grossly interlayered with tonalite and greenstones. Rocks of the hornblende suite are coarse and mesocratic (5 to 20% hornblende and biotite) and range compositionally from tonalite to granite. Granitic varieties are distinguished by large (up to .05 m) potassium-feldspar megacrysts that are strongly aligned in areas affected by broad shear zones.

Biotite granite occurs widely along the north margin of the Pasquatchai River area (see Figure 21.2) and as dikes and irregular masses intruding most other rocks. Biotite granite is typically a massive to weakly foliated, white to pink,
coarse-grained to inequigranular and leucocratic rock composed of subequal quartz, plagioclase and potassium feldspar with a few percent biotite. Inclusions of tonalite and tonalite gneiss occur in larger bodies of biotite granite.

Heterogeneous rocks of the sanukitoid suite occur as distinct oval intrusions in the Ellard Lake greenstone belt and at the headwaters of the Pasquatchai River, and appear to postdate most other plutonic rocks. Sanukitoid intrusions are zoned, with mesocratic diorite, monzodiorite and monzonite in the rim areas grading inward to leucocratic granodiorite and granite at the core. Zonation is pronounced in large intrusions whereas smaller bodies tend to be composed of the more mafic and quartz-undersaturated rocks. Syenite of the small sanukitoid intrusion north of Yelling Lake (see Figure 21.2) is exposed at one locality on the Echoing River.

Archean plutonic and supracrustal rocks are cut by rare mafic dikes of north to northwesterly trend. On the basis of orientation, the dikes are thought to be members of the northerly trending 1883 million-year-old Molson swarm and the northwesterly trending 1267 million-year-old MacKenzie swarm respectively (Osmani 1991). An unexposed intrusion of carbonatite north of Ellard Lake (see Figures 21.1 and 21.2) is dated at 1826 ± 97 Ma (Carb Lake carbonatite of Sage 1987) using the K-Ar method on biotite obtained from exploration drill core.

Tan fragmental limestone, tentatively correlated with the Ordovician Bad Cache Rapids Group (Sanford, Norris and Bostock 1967), is exposed at one locality on the Thorne River and is inferred to form a thin veneer on Archean rocks in the northeast part of the Pasquatchai River area (see Figure 21.2). Limestone boulders are widespread and limestone detritus typically constitutes half of beach sand deposits on Archean rocks of the area.

Structure and Metamorphism

Nearly all outcrops of plutonic and supracrustal rocks show a mesoscopic structural fabric. In late intrusions, such as those of the biotite granite and sanukitoid suites, elongated and aligned quartz, feldspar and mafic-mineral grains define a weak foliation. Regionally, these foliations strike overall west northwesterly and maintain general parallelism with large-scale structural elements such as geologic contacts. oval intrusions of the sanukitoid suite, such as at Ellard Lake, are exceptional in that foliations define an overall concentric zoned pattern.

Early plutonic rocks, such as those of the biotite tonalite and hornblende suites as well as supracrustal rocks, have a variable, but generally more intensely developed fabric than late plutonic rocks. The mesoscopic fabric comprises mineral foliations, gneissosity and long axes of inclusions most of which strike north northwesterly. Where measured, the dip of fabric elements tends to be northerly at intermediate to steep angles more so than southerly.

Younging directions are derived from pillow shapes and graded beds in less deformed parts of greenstone belts. In places in the Ellard Lake greenstone belt, such as at Sherman Lake (see Figure 21.2), pillows young inward from opposite sides and imply an overall synclinal form to the belt.

Overturnd, south-younging, north-dipping pillows are observed at one locality in the eastern Yelling Lake greenstone belt.

The area north of the Ellard Lake greenstone belt (see Figure 21.2) is extensively faulted. The North Kenyon fault zone and South Kenyon fault zone are regional-scale structures traced over 500 km west-northwesterly along the north margin of the Superior Province on the basis of aeromagnetic anomalies (Osmani and Stott 1988). In the study area, these faults lie along the trace of the Yelling Lake greenstone belt and approximately midway between the Ellard Lake and Yelling Lake greenstone belts, respectively. Splays are developed on the North Kenyon Fault Zone in the areas of Yelling Lake and the Thorne River.

The North Kenyon Fault Zone is marked by a zone of mylonite in felsic plutonic rocks and schist in greenstones. Planar fabric within the mylonite dips northerly and mineral lineations, defined by stretched quartz grains, are mainly subhorizontal although steep, down-dip lineations are locally present. Plutonic rocks in the area where the North Kenyon Fault Zone crosses the Thorne River show extensive evidence of brittle deformation, including epidote-filled fractures and narrow cataclastic zones.

The South Kenyon Fault Zone and a possible splay north of Sherman Lake (see Figure 21.2) are distinguished by broad zones of strongly foliated to protomylonitic plutonic rocks. Although unexposed west of the Echoing River, the South Kenyon fault zone appears to lie on strike with the Carb Lake carbonatite stock and may have controlled the emplacement of this Proterozoic intrusion.

An assemblage of amphibole + plagioclase is observed in mafic metavolcanic rocks of the Yelling Lake and Ellard Lake greenstone belts. Faulted parts of the Yelling Lake greenstone belt contain phyllic minerals including chlorite, sericite and biotite. Intermediate metavolcanic rocks east of Lacey Lake (see Figure 21.2) locally have an assemblage of sericite + quartz + carbonate, possibly indicating green-schist-facies metamorphism in the centre of the widest part of the Ellard Lake greenstone belt.

Economic Geology

Previous mineral production in the study area came from the Sachigo River Mine, situated about 1 km north of Sherman Lake (see Figure 21.2). From 1938 to 1941, the mine produced 1634 826 g of gold and 190 543 g of silver from 71,995 t of ore (Edwards 1944). The ore was derived from a discontinuous narrow (0.3 m wide) variably smokey grey to clear glassy quartz vein containing sparse pyrite and finely disseminated gold. The quartz vein strikes 279° dipping steeply to the north, and is conformable with regional fabric of the Ellard Lake greenstone belt, but is not exposed at surface. The auriferous quartz vein was developed where a "chocolate brown feldsparporphyry dike" intruded between volcanic flows and was subsequently sheared parallel to the dike (Edwards 1944).

Following closure of the Sachigo River Mine, the Sherman Lake area was explored sporadically (assessment files, Red Lake Resident Geologist's Office) without success at delineating significant gold mineralization. Felsic frag-
mental rocks south of Ellard Lake and parts of the Ellard Lake greenstone belt west of Ellard Lake were explored for base metals in the early 1970s.

Narrow high-grade quartz veins such as those mined at the Sachigo River locality represent extremely difficult exploration targets, largely due to poor exposure in the area. Parallelism of the auriferous sheared quartz vein at the Sachigo River Mine and regional faults suggests, however, that gold mineralization may be related to the regional faulting. Regional- and local-scale faults of overall north-northwesterly strike can possibly be delineated geophysically or through detailed mapping, and explored for gold in areas where they transect lithologically diverse parts of the greenstone belt. The area of Sellen Lake (not shown in Figure 21.2) at the headwaters of the Thorne River and Lacey Lake is recommended for gold and base metal exploration. This area is underlain by previously unmapped intermediate to felsic fragmental rocks, quartz-feldspar porphyry intrusions and metasediments. Locally, these rocks are schistose, cut by quartz veins and contain a few percent sulphides.

STULL LAKE AREA

Geology

The Stull Lake greenstone belt represents a wide (12 km), well-exposed and lithologically diverse greenstone sequence and affords a good locality for resolving the geology of greenstone belts in the northern Superior Province. Stone and Pufahl (1995) divided the Stull Lake greenstone belt into 4 lithologic panels. Panel 1 in the north was composed of north-younging, pillowed, maﬁc metavolcanic rocks and sandstone-siltstone sequences. Panel 2 was made up of south-younging, intermediate to felsic, metavolcanic flows and fragmental rocks with associated clastic metasediments and rare maﬁc ﬂows. Panel 3 consisted of south-younging, alluvial fan-ﬂuvial deposits of boulder conglomerate, cross-bedded sandstone and massive sandstone. These have been correlated by Downie (1937) with late sedimentary sequences at Oxford Lake, Manitoba (Oxford Lake Group). Panel 4, like panel 1, was made up of north-younging, pillowed, maﬁc ﬂows and deep-water metasediments. The boundaries of panels are marked by intrusive contacts with adjacent plutonic rocks and internally by faults.

Since 1995, helicopter-supported mapping has reﬁned the distribution of plutonic suites in areas adjacent to the Stull Lake greenstone belt (see for example, the southern half of Figure 21.3). Cross-border compilation and mapping with staff of Manitoba Energy and Mines has succeeded in joining various panels of the Stull Lake greenstone belt with greenstone stratigraphy in the Twin Lakes and Little Stull Lake areas (see Figures 1, 3; T. Corkery, Manitoba Energy and Mines, personal communications, 1998).

Recent mapping has reﬁned several major faults that transect the Stull Lake greenstone belt. The west-northwesterly striking fault, which marks the south margin of panel 3 at Wynne Bay of Stull Lake (Wynne Bay deformation zone of Atkinson, Parker and Storey 1990) and 2 splays north of Wynne Bay have been traced eastward where they merge and deﬁne a broad zone of strongly schistose rocks. The fault zone at Wynne Bay is possibly correlatable with the Stull Lake–Wunnemin Lake fault zone of Osmani and Stott (1988) and therefore constitutes a regional-scale structure.

Preliminary analysis of geochemical data indicates that volcanic rocks in the southern part of panel 2 are alkaline and are classiﬁed as trachybasalt and trachyte using the scheme of Le Maitre (1989, Figure 13). In outcrop, the alkaline rocks are ﬁne grained, hard and grey with distinct feldspar laths and show a variety of ﬂow and fragmental textures. This contrasts with pillow foid and fragmental felsic metavolcanic rocks from other panels whose compositions include basalt, basaltic andesite and dacite.

The closely associated alkaline volcanic rocks and ﬂuvial sediments of panel 3 are characteristic of a Timiskaming-type sequence (Thurston and Chivers 1990; Carter 1993) and dictate revision to the panels of the Stull Lake greenstone belt. Although further mapping is needed to establish the extent and boundary relations (fault or unconformity) of the alkaline volcanic rocks, it is suggested that panel 3 be enlarged at the expense of panel 2 to group together the alkaline volcanic rocks and coarse clastic metasediments as shown in Figure 21.3.

The alkaline volcanic rocks and ﬂuvialite metasediments of the revised panel 3 have characteristics typical of Timiskaming-type sequences, including a high stratigraphic position in the greenstone sequence and close spatial association with regional-scale faults. The Timiskaming-type sequence indicates that the Stull Lake greenstone belt had late, fault-controlled volcanism and sedimentation, which is comparable to the ﬁnal tectonomagmatic stages in the evolution of southern Superior Province greenstone belts (Carter 1993).

The fault-bounded Timiskaming-type sequence at Stull Lake is of economic importance due to the occurrence of gold in sheared volcanic rocks at the margins of panel 3 at Twin Lakes Manitoba (Figure 21.3; see also Richardson et al. 1996). The gold mineralization is associated with boundary faults of the Timiskaming-type sequence comparable to the style of mineralization in major gold camps such as Kirkland Lake (Jackson and Fyon 1991, p. 440). By analogy, the Stull Lake area has signiﬁcant gold potential.

SUMMARY AND REGIONAL IMPLICATIONS

Mapping in the Ellard Lake, Pasqua River and Stull Lake areas provides new insight on regional geology, economic potential and tectonic evolution of the northern Superior Province. Through comparison with concurrent work in the Wabigoon Subprovince (Stone et al., this volume) several geologic similarities and contrasts are evident between the north and south Superior Province.

Greenstone belts of the northern Superior Province show a broad range of volcanic and sedimentary sequences that are not unlike those found in belts to the south. Stone and Haﬂé (1997) noted that the Ponask–Sachigo Lakes greenstone belt (see Figure 21.1) is unique in that it contains tonalite-cobble
conglomerates, marble and komatiites characteristic of the platform sequences of Thurston and Chivers (1990) that predate 2.8 Ga. This may be an indication that older crustal material, perhaps representative of the North Caribou Terrane, occurs in the southern part of the study area, whereas greenstone sequences of the Stull Lake, Ellard Lake and Yelling Lake greenstone belts are younger additions to the crust. In this context, both the northern and southern parts of the Superior Province appear to contain Mesourchean and Neoarchean greenstone sequences.

Figure 21.3 Geology of the Stull Lake area, Ontario and Manitoba.
Five major and one minor suite of felsic plutonic rocks are identified in the northern Superior Province and these can be readily correlated in terms of their mineralogy, composition and textural characteristics with plutonic suites in the south. Further, the relative ages of suites are the same in both areas. The biotite tonalite and tonalite gneiss suites are succeeded by the hornblende suite followed by biotite granite and sanukitoid suites. Although the absolute ages and relative abundance of plutonic suites may vary, the common sequence of magmas implies a similar process for magma generation and evolution in the north and south.

Perhaps the most significant difference between north and south Superior Province geology that can be made on the basis of available information concerns the form and distribution of greenstone belts and plutonic suites. Although greenstone belts of the Wabigoon Subprovince follow an overall east-west trend, they tend to be highly curved, bifurcated and tapered, with narrow arms transitioning to gneisses (see Figure 20.1 in Stone et al., this volume). Greenstone belts of the Wabigoon Subprovince are interspersed with oval to lobate intrusions and intrusive complexes. In contrast, greenstone belts of the northern Superior Province are linear, maintaining fairly uniform width and west-northwesterly orientation and have sharp contacts. With the exception of late oval sanukitoid intrusions, most felsic plutonic bodies are also highly elongate west-northwesterly (see Figure 21.1).

It is unclear whether the distinct map patterns imply a fundamentally different process for assembly of the supracrustal and plutonic sequences or merely different levels of strain and erosion in the north and south Superior Province. Pervasive west-northwesterly striking foliations and 3 regional-scale faults with associated splays allude to broad, late-stage deformation in the north. Possibly the straight nature of greenstone belts in the north also arises from higher levels of strain that may be unique to the north margin of the Superior Province during, and subsequent to, assembly of the craton. Thermobarometric studies that are in progress will test for vertical displacements on the regional faults in the north and help to evaluate whether or not the diverse large-scale architecture of the granite-greenstone areas is related to different levels of uplift in the north and south.

Regional faults enhance the gold potential of the northern Superior Province. Small, high-grade gold deposits such as the Sachigo River Mine appear to be developed on shear zones parallel to the regional structures, and gold is associated with the faulted margins of Timiskaming-type assemblage at Stull Lake. In the latter case, a favourable environment for gold mineralization extends for more than 30 km through Manitoba and Ontario in the Stull Lake area.

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INTRODUCTION

This multi-year project is conducted in conjunction with the Western Superior NATMAP program. The project was initiated to study the regional metallogeny of the northwestern Superior Province with emphasis on volcanogenic massive sulphide (VMS) mineralization and gold. Field work was conducted in the Sachigo and Uchi subprovinces and focussed on 1) greenstone belts with known VMS mineralization; and 2) greenstone belts with potential to host VMS-productive metavolcanic assemblages. This work will identify greenstone belts that host metavolcanic assemblages with high mineral potential by

- Documenting the distribution and features of VMS mineralization as well as: related alteration; structural and lithologic controls; and geochemical and mineralogical characteristics of metavolcanic host rocks and mineralization.
- Comparing the metallogeny of early (pre-2.74 Ga) and late (2.74 to 2.7 Ga) Neoarchean metavolcanic assemblages.

The sites visited in the Sachigo Subprovince (Figure 22.1) were: Shrimp Lake (NTS 53C/7) in the North Spirit Lake greenstone belt; Setting Net Lake (NTS 53C/13) in the Favourable Lake greenstone belt; and Eyapamikama Lake in the North Caribou greenstone belt (NTS 53B/14, 53B/15). Sites visited in the Uchi Subprovince (see Figure 22.1) were: the Red Lake greenstone belt (NTS 52N/3, 52N/4, 52M/1, 52K/13, 52L/16); Maskooch Lake (NTS 52K/16) in the Birch–Uchi greenstone belt; and Discovery Lake (NTS 52P/8) in the Attwood Lake greenstone belt. The general characteristics of VMS mineralization and host rocks in these study areas are summarized in Table 22.1.

VMS MINERALIZATION IN THE SUPERIOR PROVINCE

Productive VMS deposits in the Superior Province occur within metavolcanic assemblages ranging in age from 2.74 to 2.7 billion years old (Fyon et al. 1992). These assemblages are interpreted to have formed in extensional tectonic environments such as, thickened, bimodal oceanic rifts and primitive, tholeiitic to calc-alkaline rifted island arcs (Fyon et al. 1992; Galley 1995). VMS deposits consist of zinc-copper ± lead ± silver sulphide mineralization (Fyon et al. 1992) associated with felsic metavolcanic rocks extruded from subvolcanic, high-level magma chambers along synvolcanic faults. High-level magma chambers are considered to be the heat source for convective mineralizing hydrothermal systems required for the development of a VMS deposit (Lesher et al. 1986; Galley 1995).

Felsic metavolcanic rocks have distinctive trace element geochemical characteristics that 1) assist in distinguishing barren and VMS-productive metavolcanic assemblages; and 2) assist in defining tectonic environments where large VMS deposits are located (Galley 1995). F1 felsic metavolcanic rocks are considered to be barren of significant VMS deposits (Lesher et al. 1986) and are derived from deep-seated magma chambers in compressional subduction settings (Lesher et al. 1986; Lentz 1997). These magma chambers were too deep to function as effective heat sources for convective mineralizing hydrothermal systems (Lesher et al. 1986). F1I and F1II felsic metavolcanic rocks are VMS-productive and were derived from shallow magma chambers in extensional tectonic environments (Lesher et al. 1986; Lentz 1997). FII felsic metavolcanic rocks occur in VMS-productive successions at the Mattabi and Lyon Lake deposits at Sturgeon Lake and the Geco deposit at Manitouwadge (Lesher et al. 1986; Zaleski et al. 1995). F1II felsic metavolcanic rocks occur in VMS-productive successions at the Kidd Creek and Kamiskotia deposits in the Abitibi greenstone belt and at the South Bay deposit in the Birch–Uchi greenstone belt (Fyon et al. 1992).

Early Neoarchean metavolcanic assemblages older than 2.74 Ga, in the Sachigo and Uchi subprovinces, are not known to host significant VMS deposits. Felsic metavolcanic rocks with F1I and F1II geochemical signatures have not been recognized in the northern greenstone belts (Parker and Ayer 1997). Most VMS mineralization contains significant amounts of lead and silver, which is in contrast to lead- and silver-poor sulphide mineralization in younger (2.74 to 2.7 Ga) VMS deposits (Fyon et al. 1992). Several questions arise from these observations: 1) Were large VMS deposits formed in early Neoarchean tectonic environments in the Superior Province?; 2) Is the apparent absence of significant VMS deposits and F1I/F1II felsic metavolcanic rocks in early (pre-2.74 Ga) metavolcanic assemblages due to an incomplete database; poor bedrock exposure; and/or a lack of recent base metal exploration activity?; 3) Do the characteristics of VMS mineralization and host rocks in early (pre-2.74 Ga) metavolcanic assemblages differ from those in late (2.74 to 2.7 Ga) assemblages?
Figure 22.1. Location of study areas in the Sachigo and Uchi subprovinces, northwestern Ontario.
The above questions will be addressed during the course of this project.

**VMS MINERALIZATION IN EARLY NEOARCHEAN (< 2.74 Ga) META VOLCANIC ASSEMBLAGES**

**Setting Net Lake—Favourable Lake Greenstone Belt**

Volcanogenic sulphide mineralization at Setting Net Lake (see Figure 22.1) occurs within the Setting Net assemblage, a 2.925 billion-year-old unit (Corfu and Ayres 1991), that may represent oceanic volcanism (Thurston et al. 1991). The assemblage consists of deep-water, subaqueous, mafic and ultramafic metavolcanic flows and interflow metasedimentary rocks. The metasedimentary rocks include thin discontinuous units of argillite, silstone and wacke; felsic tuff and tuff breccia; quartz-magnetite and sulphidic iron formation; and chert (Ayres 1970, 1972). The oceanic assemblage is succeeded by a sequence of intermediate to felsic pyroclastic rocks, flows and metasedimentary rocks interpreted to be part of a collapsed volcanic caldera in an arc environment (Buck 1986).


Sulphide mineralization consists of fine-grained pyrite; pyrrhotite; minor chalcopyrite on fractures; and rare disseminated sphalerite dominantly hosted by black graphitic argillite, chert and felsic tuff. Sulphide minerals are disseminated and/or concentrated in thin layers, nodules or discontinuous massive lenses. Minor disseminated sulphide minerals are hosted by mafic metavolcanic flows. Sulphide mineral-bearing metasediments contain anomalous metal contents ranging between 63 and 5600 ppm Zn; 65 to 7000 ppm Cu; 1.6 to 3.9 ppm Ag; and 2 to 51 ppb Au (Atkinson et al. 1991). Significant sulphide mineralization is present at the Springer, McCluskey, Senet and Oliver occurrences (Ayres 1970, 1972; Atkinson et al. 1991).

Proximal, localized, synvolcanic alteration occurs stratigraphically below mineralization at a few occurrences, such as the McCluskey occurrence where disseminated pyrrhotite, sphalerite and chalcopyrite veinlets are hosted by a 30 m thick chert horizon (Ayres 1971, 1972). The chert is underlain by mafic metavolcanic flows altered to a calc-silicate assemblage of quartz, epidote, diopside, garnet and calcite. Alteration varies in thickness from a few metres to several tens of metres and increases in intensity toward the chert horizon. Grab samples from this

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**Table 22.1. General characteristics of VMS mineralization and associated assemblages in study areas.**

<table>
<thead>
<tr>
<th>STUDY AREA</th>
<th>U/Pb AGE (Ga)</th>
<th>ASSEMBLAGE</th>
<th>ENVIRONMENT</th>
<th>MAJOR HOST LITHOLOGIES</th>
<th>MINERALS METALS</th>
<th>ALTERATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Setting Net Lake; Berens River mine, Favourable belt</td>
<td>2.95</td>
<td>Setting Net</td>
<td>oceanic sequence; volcanic caldera, arc volcanism</td>
<td>mafic mv, interflow metasediments; inter-felsic caldera-fill sequence</td>
<td>po, py, cp, sp Zn, Ag, Cu, Pb</td>
<td>amp, gtr; ser, py, grt-amp</td>
</tr>
<tr>
<td>2. Arsoke Lake (Eyapamikama Lake), North Caribou belt</td>
<td>2.932 to 2.87</td>
<td>North Rim metavolcanics; Eyapamikama metasediments</td>
<td>Oceanic volcanism, medial to distal fan turbidites</td>
<td>mafic mv, elastic and chemical metasediments</td>
<td>po, py, sp, gn, cp Zn, Pb, Ag, Cu</td>
<td>fracture-controlled amp, gtr</td>
</tr>
<tr>
<td>3. Mulcahy Tp, Red Lake belt</td>
<td>2.9 to 2.96</td>
<td>Balmer</td>
<td>oceanic to shallow submarine</td>
<td>mafic mv, elastic and chemical metasediments</td>
<td>po, py, sp, gn, cp pent, Zn Ag, Pb, Cu, Ni</td>
<td>ath</td>
</tr>
<tr>
<td>4. Shrimp Lake, North Spirit belt</td>
<td>2.743 to 2.731</td>
<td>Hewitt</td>
<td>arc volcanism</td>
<td>syngenic porphyry, lahars</td>
<td>po, py, cp Cu</td>
<td>ath, gtr, crd, ser, amp, bio</td>
</tr>
<tr>
<td>5. Discovery Lake, Attwood belt</td>
<td>2.723 to 2.716</td>
<td>St. Joseph</td>
<td>arc volcanism</td>
<td>felsic pyroclastics and metasediments</td>
<td>po, py, sp, gn, cp Zn, Cu, Ag</td>
<td>fracture-controlled amp, gtr st, ath, bi</td>
</tr>
<tr>
<td>6. Maskooch Lake; Birch–Uchi belt</td>
<td>2.8 to 2.73</td>
<td>Confederation</td>
<td>arc volcanism</td>
<td>felsic flows, pyroclastics, metasediments</td>
<td>po, py, cp Zn, Ag, Au</td>
<td>fracture-controlled amp, grt, mag, ser</td>
</tr>
</tbody>
</table>

**Abbreviations:** Ag = silver, amp = amphibole, ath = anthophyllite, Au = gold, bi = biotite, cp = chalcopyrite, crd = cordierite, Cu = copper, gn = galena, gtr = garnet, int = intermediate, mag = magnetite, mv = metavolcanic rocks, Ni = nickel, Pb = lead, pent = pentlandite, po = pyrrhotite, py = pyrite, ser = sericite, sp = sphalerite, st = staurolite, Zn = zinc.
occurrence analyzed up to 0.32% Cu, 2.2% Zn and 0.72 ounce silver per ton (Ayres 1971).

The Berens River, Golsil or Zahavy mine is located within the caldera sequence in the Setting Net assemblage (2.925 Ga). The mine produced 898 tons of zinc; 3052 tons of lead; 5.676 million ounces silver; and 157,341 ounces gold from 560,707 tons of ore between 1939 and 1948 (Adams 1976). Semi-massive to disseminated pyrite, pyrrhotite, galena, sphalerite, tetrahedrite and minor chalcopyrite are hosted by extensional quartz-actinolite breccia veins in several east-northeast-striking faults and shear zones. The mineralized quartz veins are interpreted (Adams 1976) to have formed early in the metamorphic-deformational history of the area. This interpretation is consistent with the recrystallized nature of the quartz veins.

The intermediate to felsic metavolcanic rocks of the caldera sequence have been altered by widespread, pervasive, sericite-pyrite-pyrrhotite and more local garnet-amphibole alteration concentrated in networks of intersecting fractures or interconnected, diffuse patches. The garnet-amphibole alteration also replaces the matrix and clasts of pyroclastic and metasedimentary rocks. Some of this alteration, such as the sericite-pyrite-pyrrhotite, may be related to the porphyry-style, molybdenum-mineralized Setting Net Lake stock emplaced into the Setting Net assemblage at 2.708 Ga (Corfu and Ayres 1991).

**Eyapamikama Lake—North Caribou Greenstone Belt**

The Arseno Lake prospect (Figure 22.2) is 2 km north of the west end of Eyapamikama Lake (see Figure 22.1) and situated within the North Rim metavolcanics, dated at 2.932 Ga, and the Eyapamikama metasediments deposited between 2.87 and 2.932 Ga (Breaks and Bartlett 1991). The metavolcanic rocks are interpreted (Breaks and Bartlett 1991) to have been deposited in a relatively shallow, subaqueous, oceanic depositional environment with the Eyapamikama metasediments representing a submarine fan turbidite unit (Thurston et al. 1991). Metamorphic grade is lower to middle amphibolite with disseminated garnet, staurolite and cordierite porphyroblasts in metasedimentary rocks.

The North Rim metavolcanics at Arseno Lake include narrow, east-striking sequences of weakly amygdaloidal, pillows flows; medium- to coarse-grained massive flows; and minor ultramafic flows. The metavolcanic rocks are tectonically interlayered with the overlying Eyapamikama metasediments (see Figure 22.2) that range from conglomerate to mature, quartz-rich arenites, wackes and mudstone. Ferruginous chert, chert-grunerite and quartz-magnetite iron formation occur along contacts between the metavolcanic and metasedimentary rocks (Bartlett et al. 1985; Breaks and Bartlett 1991). A thick unit of poorly sorted, matrix-supported, volcaniclastic conglomerate with a dark green amphibole-rich matrix is situated south of the North Rim metavolcanics (see Figure 22.2). Narrow, discontinuous units of aphanitic to fine-grained felsic tuff, quartz ± feldspar crystal tuff and felsic lapilli tuff are interlayered with mafic metavolcanic flows and metasedimentary rocks. The felsic metavolcanic rocks are strongly foliated and variably sericitized with disseminated green mica and pyrite. All of the rocks are intruded by medium- to coarse-grained ultramafic dikes or sills.

The east-striking contact between metasediments and metavolcanic rocks is coincident with a 300 m wide deformation zone which hosts the Arseno Lake prospect.

In the late-1980s, mineral exploration at Arseno Lake by Northern Dynasty Explorations Limited outlined a diamond-drill indicated resource of 1 million tons averaging 8.7% combined zinc and lead and 1.5 ounces silver (The Canadian Mines Handbook, 1989–90, p.346). Sulphide mineralization occurs in narrow, discontinuous units of recrystallized, granular or sugary ferruginous white and grey chert and quartz-magnetite iron formation. Mineralization consists of disseminated to semi-massive, fine- to coarse-grained pyrrhotite; fine- to medium-grained, granular sphalerite; disseminated galena; euclidean to subhedral porphyroblasts of pyrite up to 4 mm in size; and chalcopyrite on fractures. Semi-massive sulphide mineralization has a distinctive baulk-texture where surrounded to angular fragments of chert are embedded in a sulphide matrix. This texture is indicative of deformation and sulphide mineral remobilization.

The sulphide mineral-bearing chert and/or iron formation is interlayered with feldspathic arenite; siliceous pelitic units; garnetiferous, actinolite- and chlorite-rich clastic metasedimentary rocks; felsic, sericitic, quartz porphyritic rock; felsic lapilli tuff; and sericite-garnet-biotite ± green mica schist. Garnet porphyroblasts range in abundance from 2 to 60% and are 2 mm to 3 cm. Garnet content in metasediments increases toward contacts with chert and/or iron formation. Small elliptical biotite clots and variable amounts of disseminated green mica are common in altered metasediments and felsic rocks.

Web-like networks of abundant, intersecting veins of dark green amphibole (hornblende, actinolite) ± garnet occur throughout the metasedimentary rocks adjacent to the chert and/or iron formation over a discontinuous strike length of at least 20 km. Amphibole veins are isoclinically folded by D1 deformation and interpreted (Breaks and Bartlett 1991) to have developed early in the metamorphic-deformational history of the area. The amphibole veins may represent syngenic, iron-rich alteration related to the disposition of the sulphide mineral-bearing chert and/or iron formation. This alteration is overprinted by late quartz-tourmaline and quartz-iron carbonate veins associated with arsenopyrite and gold mineralization.

**Mulcahy Township Prospects—Red Lake Greenstone Belt**

Significant zinc-copper and nickel-copper sulphide mineralization occurs in the Balmer assemblage in Mulcahy Township at the west end of the Red Lake greenstone belt (see Figure 22.1). The Balmer assemblage, dated at 2.99 to 2.96 Ga, consists of mafic and ultramafic metavolcanic flows and sills; a variety of shallow water clastic and
chemical metasedimentary rocks; and minor intermediate to felsic metavolcanic rocks that may represent an oceanic to shallow submarine depositional environment (Stott and Corfu 1991).

The Zinc Pit and Trout Bay nickel zone prospects (Figure 22.3) are situated within an east-to-northeast-younging, northwest-striking sequence of metasedimentary rocks that range from conglomerate to argillite. Metasedimentary rocks are interlayered with minor units of intermediate to felsic tuff, lapilli tuff and tuff breccia and narrow discontinuous units of quartz-magnetite iron formation. The metasedimentary rocks were indicated by Riley (1968) to be intruded by conformable, composite, gabbro sills (see Figure 22.3). However, recently improved bedrock exposures resulting from logging and mineral exploration activities indicate that the "gabbro sills" may actually represent metavolcanic flows. There is abundant evidence of pillowed flows, pillow breccia and hyaloclastite grading into more massive, medium-grained flows that are mafic and ultramafic in composition.

Metamorphic grade is upper greenschist to lower amphibolite. Metasedimentary rocks contain biotite, sericite, anthophyllite, actinolite-tremolite and porphyroblasts of garnet, staurolite, cordierite and andalusite. The mafic and ultramafic metavolcanic rocks commonly contain garnet, actinolite-tremolite and chlorite.

The metavolcanic-metasedimentary sequence is tightly folded with numerous S- and Z-drag folds plunging to the west and mineral lineations plunging moderately to the west-northwest. The rocks at Zinc Pit 1 and 2 are folded and dislocated along sinistral and dextral faults that exhibit considerable variation in attitude. A strong northwest-striking foliation is co-planar with layering and overprinted by a second, northeast-striking spaced cleavage. Staurolite porphyroblasts in argillite are aligned along the northeast-striking cleavage.

The Zinc Pit prospect consists of 2 separate zones of zinc-copper sulphide mineralization approximately 200 m apart (see Figure 22.3). Exploration has delineated 124,760 tons grading 7.86% Zn, 1.5% Cu, 0.24% Pb and

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**Eyapamikama Metasediments**
- Chert and/or iron formation
- Garnet-bearing metasediments ± staurolite ± amphibole veins; felsic tuff; felsic porphyries
- Volcaniclastic conglomerate (Lahar ?)

**North Rim Metavolcanics**
- Pillowed and massive mafic metavolcanic flows ± metasediments
- Mineral occurrence

*Figure 22.2.* General geology of the Arseno Lake prospect at Eyapamikama Lake, North Caribou greenstone belt (modified after Bartlett et al. 1985; Piroshco et al. 1989).
Figure 22.3. General geology of the Zinc Pit and Trout Bay Nickel Zone, Mulcahy Township, Red Lake greenstone belt (modified after Riley 1968).
1.7 ounces silver per ton in a 9 m wide and 140 m long mineralized zone at Zinc Pit 1 (see Figure 22.3). The west zone of Zinc Pit 2 (see Figure 22.3) is 6 m wide and 30 m long containing 13,776 tons grading 4.75% Zn, 0.68% Cu and 0.94 ounce silver per ton (Shklanka 1969).

Sulphide mineralization at Zinc Pit 2 is hosted within a brecciated, recrystallized, ferruginous chert unit at its contact with a massive, fine- to medium-grained mafic flow. The chert consists of dismembered chert layers and tabular, folded and rounded chert fragments in a fine-grained, dark green, garnet-amphibole matrix. The flow dominantly consists of actinolite and plagioclase with widely scattered feldspar phenocrysts and irregular clusters of tiny, pink garnet porphyroblasts. The flow contains a narrow unit of interflow metasedimentary rocks and pillow breccia with an albite-rich matrix. The flow also contains abundant albite in large irregular patches and along numerous fractures and veins. The abundance of intense albite alteration and some associated epidote may represent spilitization.

The chert breccia is enclosed within thinly bedded argillite, siltstone and felsic tuff hosting staurolite porphyroblasts and fibrous amphibole. A thin unit of felsic tuff breccia consists of matrix-supported, angular and tabular felsic clasts. The clasts and matrix host green-brown, fibrous amphibole (anthophyllite?) that forms radiating bow-tie structures.

The mineralized zone is 3 m wide and 10 m long and consists of massive, fine- to medium-grained black spherulite intergrown with pyrrhotite and minor chalcopyrite. Pyrite occurs in nodules concentrated along thick, irregular layers. Spherulite-poor mineralization consists of pyrrhotite with 5 to 10% chalcopyrite. The sulphide mineralization is weakly layered and hosts angular chert fragments. Two grab samples taken by the author from the mineralization analyzed 10.7% Zn, 2270 ppm Cu, 80 ppm Ni, 8500 ppm Pb, 19 ppm Ag, 271 ppm Au; and 12% Zn, 5200 ppm Cu, 28 ppm Ni, 5000 ppm Pb and 2.36 ounces silver per ton.

The geology at Zinc Pit 1 is similar and consists of a 3 m wide and 43 m long mineralized zone at the contact between an albited mafic flow and argillite interlayered with felsic tuff, siltstone and quartz-magnetite iron formation. The argillite hosts minor garnet and abundant staurolite porphyroblasts ranging in size from 0.5 to 4 cm. The sulphide mineralization is hosted within a strongly gossan-stained, amphibole-rich rock composed of radiating bow-ties of brown-black amphibole (anthophyllite?). The amphibole-rich rock hosts rectangular or elliptical pyrrhotite clasts and is interlayered with recrystallized, pyritic and sericitic chert. The sulphide mineralization is similar in composition to mineralization at Zinc Pit 2 and contains interlayered chalcopyrite and sphaerlite.

The Trout Bay nickel prospect (see Figure 22.3) consists of nickel-copper sulphide mineralization at the base of a composite gabbro sill (Kurylyk 1963) near its contact with quartz-magnetite iron formation. The mineralized zone is 122 m long and 10 m wide with an approximate grade of 0.5% Ni and 0.25% Cu (Shklanka 1969).

The gabbro sill may actually represent an ultramafic to mafic flow with disseminated and semi-massive chalcopyrite, pyrrhotite, pyrite and pentlandite in a tremolite-antigorite-magnetite schist (Kurylyk 1963) at the base of the flow. Sulphide mineralization is strongly deformed and consists of 1) small elongate lenses and narrow layers parallel to foliation, and 2) extension fractures that transect foliation. Chalcopyrite halos enclose pentlandite-rich lenses in altered host rocks (Shklanka 1969). The Trout Bay nickel prospect may represent a komatiite-hosted copper-nickel deposit.

Small occurrences of base metal sulphide mineralization are widespread throughout Mulcahy Township (Riley 1968; see Figure 22.3). The majority of copper, zinc, nickel and silver mineralization occurs at contacts between mafic to ultramafic rocks, clastic metasedimentary rocks and iron formation.

VMS MINERALIZATION IN LATE NEOARCHEAN (2.74 TO 2.7 Ga)
METAVOLCANIC ASSEMBLAGES

Shrimp Lake—North Spirit Lake Greenstone Belt

Shrimp Lake (Figure 22.4) is located in the southeast corner of the North Spirit Lake greenstone belt (see Figure 22.1) in the 2.743 to 2.731 Ga (Corfu and Wood 1986) Hewitt assemblage interpreted as an arc sequence (Thurston et al. 1991). Metamorphic grade is upper greenschist to middle amphibolite. Metasedimentary rocks contain porphyroblasts of garnet, andalusite and cordierite.

Metavolcanic rocks at Shrimp Lake consist of pillow, massive and coarse-grained mafic metavolcanic flows over lain by intermediate to felsic, lapilli tuff, tuff breccia and feldspar crystal tuff interlayered with fine-grained arkosic wackes, arenites and mudstones (see Figure 22.4). The most common rock type is intermediate, heterolithic, lapilli tuff to tuff breccia consisting of poorly sorted, subrounded to subangular clasts of feldspar porphyry and biotite-amphibole mafic clasts in a feldspar crystal-rich matrix. The matrix and clasts contain anomalously abundant fibrous green amphibole.

Supracrustal rocks are intruded by a synvolcanic feldspar porphyritic stock (see Figure 22.4). The porphyry is epidotized and hosts disseminated pyrrhotite, pyrite and chalcopyrite. The porphyry contains inclusions of black carbonate and ultramafic rock at its southwestern contact (Wood 1988).

The porphyry is enclosed in a clastic apron of sulphide mineral-bearing, poorly sorted, disorganized, clast- to matrix-supported conglomerate (see Figure 22.4). Wood (1988) mapped the conglomerate for at least 7 km along strike. The conglomerate is composed of subrounded to subangular, intermediate to felsic metavolcanic clasts; abundant angular clasts of recrystallized glassy quartz;
mafic metavolcanic clasts; and pyrrhotite clasts in an arenaceous, amphibole-rich matrix. Abundant pyrrhotite, pyrite and minor chalcopyrite are disseminated throughout the matrix and clasts with the exception of the quartz clasts. The composition of the conglomerate is highly variable and changes laterally along strike. Some sections consist only of recrystallized, glassy quartz breccia with quartz clasts of variable sizes in a green, amphibole-sulphide mineral matrix. Wood (1988) interpreted the conglomerate as a debris flow or lahar that may have slumped off the side of a volcano in a deep-water, subaqueous environment.

Alteration in pyroclastic rocks consists of variable amounts of disseminated biotite and green amphibole in matrix and clasts. Anthophyllite-cordierite-garnet alteration was identified in intermediate metavolcanic rocks north of Shrimp Lake (see Figure 22.4). Major oxide geochemistry of the altered rock indicates considerable enrichment in iron and magnesium relative to unaltered equivalent rocks (Wood 1988). Magnesium and iron enrichment alteration was also recognized in mafic metavolcanic flows south of Shrimp Lake (Wood 1988; see Figure 22.4).

Significant zinc-copper sulphide mineralization has not been identified at Shrimp Lake, however, the geological environment has VMS potential. Wood (1988) also noted widespread sericitization and silification affecting all supracrustal rocks in the Tahoe Lake area, northwest of Shrimp Lake. The only mineral exploration activity reported at Shrimp Lake was ground geophysical surveys conducted by Noranda Exploration Company Limited in 1983.

**Discovery Lake—Attwood Lake Greenstone Belt**

Discovery or “Disco” Lake is located 100 km north of Armstrong and 25 km west of Fort Hope (see Figure 22.1) at the northeast end of the Attwood Lake greenstone belt in the 2.723 to 2.716 Ga St. Joseph assemblage (Ontario Geological Survey 1992). The assemblage is interpreted to

![Figure 22.4. General geology of the Shrimp Lake area, North Spirit greenstone belt (modified after Wood 1988).](image)
represent proximal and vent facies arc volcanism (Stott and Corfu 1991).

The geology at Discovery Lake (Figure 22.5) consists of a north- to northwest-facing, northeast-striking, 0.5 to 1 km thick sequence of intermediate to felsic tuff, lapilli tuff and tuff breccia interlayered with fine-grained, intermediate to felsic, volcaniclastic metasedimentary rocks enclosed within thick sequences of mafic metavolcanic flows. The metavolcanic rocks are intruded by massive gabbro sills and felsic feldspar porphyry dikes. Metamorphic grade is lower to middle amphibolite.

The felsic metavolcanic rocks are generally fine-grained tuffs, quartz crystal tuffs and heterolithic to monolithic, poorly sorted, matrix-supported, lapilli tuff to tuff breccia. Felsic clasts are siliceous and subrounded to subangular whereas biotitic mafic clasts are dark green-black, elliptical to subrounded and contain garnet and/or staurolite porphyroblasts.

The felsic metavolcanic rocks have been affected by widespread alteration over an area that is at least 3.5 km long and 500 m wide (see Figure 22.5). Alteration consists of a persistent network of anastomosing and intersecting amphibole-garnet ± magnetite ± biotite ± staurolite veins that may represent hydro-fracturing of the metavolcanic rocks (see Figure 22.5). The veins are less than 1 to 4 cm wide; vary in intensity and abundance; and impart a "jigsaw-fit" or "brickwork" appearance to outcrops. Wide and/or closely spaced veins give the rock a "pseudo"-frag-

![Figure 22.5](image)

**Figure 22.5.** General geology of the Discovery or "Disco" Lake area, Attwood Lake greenstone belt *(modified after Felix 1995, 1996).*
mental appearance making it difficult to determine if the host rock was originally fragmental. The alteration also occurs as diffuse, interconnected patches of garnet-amphibole. Amphibole-garnet veins also occur in mafic metavolcanic flows, but were not observed in metasedimentary rocks such as argillite and siltstone. Edwards (1991) reported extensive calcium and sodium depletion and moderate to strong iron enrichment associated with the “brickwork” alteration. Strongly altered felsic metavolcanic rocks also contain abundant sericite, biotite and variable amounts of garnet, staurolite and amphibole porphyroblasts. Felsic metavolcanic rocks may also host irregular white patches of albite containing dark green, acicular amphibole needles and garnet and staurolite porphyroblasts.

A siliceous, intensely altered anthophyllite-garnet-staurolite-biotite ± cordierite rock was intersected in diamond-drill holes at the Goldfields and South Main occurrences (see Figure 22.5). The rock contains abundant, feathery anthophyllite bow-ties up to 2 cm long and narrow garnet veins. Felix (1996) reported that the Ishikawa alteration index for this rock (in diamond-drill hole PO-96-12) was 82 which is indicative of intense hydrothermal alteration.

Mafic metavolcanic flows are pillowved or massive, medium-grained and gabbroic and interlayered with thin units of interflow metasedimentary rocks. The mafic flows contain albite ± epidote veins and garnet-amphibole concentrated in pillow selvages. Abundant garnet veins and thick aggregates of large (2 to 4 cm) red garnets were observed in pillowved mafic flows at the Ryley–Cormac

occurrence. In diamond drill core, thick seams of biotite occur throughout the mafic flows as well as thick layers of biotite ± amphibole ± garnet at contacts between mafic and felsic metavolcanic rocks.

Abundant, late, retrograde chlorite alteration overprints prograde mineral assemblages. Garnet and amphibole are commonly altered or completely replaced by thick, fine-grained, green chlorite that also occurs along foliation and late fractures. Late epidote alteration is also common and consists of thick irregular patches in mafic metavolcanic rocks. Epidote also occurs along late hairline fractures that transect and alter garnet porphyroblasts.

Thin units of quartz-magnetite and sulphidic iron formation are intercalated with argillite, chert, felsic lapilli tuff and quartz crystal tuff at the top of the felsic metavolcanic sequence (see Figure 22.5). Abundant, pyrite- pyrrhotite layers and pyrite nodules occur throughout black argillite. Brecciated ferruginous chert and sulphidic iron formation also host semi-massive pyrite-pyrrhotite. This sulphide mineralization contains anomalous metal values such as 854 ppm Cu and 1170 ppm Zn (Felix 1995). Felsic metavolcanic rocks contain biotite and sericite and porphyroblasts of garnet and staurolite.

Several zinc-copper sulphide occurrences are situated along the south contact between the felsic and mafic metavolcanic sequences (see Figure 22.5). The major features of these occurrences are summarized in Table 22.2. The sulphide mineralization consists of disseminated to semi-massive pyrrhotite and pyrite, sphalerite, galena and chalcopyrite in interflow metasediments and felsic metavolcanic rocks or it is remobilized along fractures and

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Table 22.2: General characteristics of occurrences of sulphide mineralization in the Discovery Lake area. Location of occurrences are indicated on Figure 22.5.

<table>
<thead>
<tr>
<th>OCCURRENCE (COMMODITY)</th>
<th>HOST LITHOLOGY(IES)</th>
<th>STRUCTURE</th>
<th>MINERALIZATION</th>
<th>ALTERATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Nyla (Zn)</td>
<td>siltstone-chert with bio, grt, amp; mafic flows</td>
<td>east-northeast-striking foliation, shear zone</td>
<td>diss. py, po in metased; sp in veinlets</td>
<td>amp-grt veins in mafic flows; late ep, ab, chl</td>
</tr>
<tr>
<td>2. NJZ (Cu)</td>
<td>amphibolitized pillowed and massive mafic flows</td>
<td>strong east-northeast-striking foliation</td>
<td>late cp veins; net textured cp, po interstitial to amp; diffuse, diss cp, py</td>
<td>abundant diss grt; dark amp on fractures and in patches</td>
</tr>
<tr>
<td>3. Goldfields (Cu,Zn,Ag)</td>
<td>sericitic felsic tuff, with biotite clasts, quartz phenocrysts; gabbro</td>
<td>east-northeast-striking foliation</td>
<td>cp mineralization in fractures and veins</td>
<td>diss. ath, ser grt, st, bio in felsic rocks; grt-amp veins and patches</td>
</tr>
<tr>
<td>4. Ryley-Cormac (Zn)</td>
<td>pillowved mafic flows; felsic tuff; intermediate to felsic metasediments; gabbro</td>
<td>east-northeast-striking foliation, lineations plunge steeply west</td>
<td>vein and fracture hosted po, py, cp, sp, gn in felsic rocks</td>
<td>intense grt-amp in veins and patches; late ep, ab</td>
</tr>
<tr>
<td>5. South Main (Cu,Ag,Au)</td>
<td>intermediate to felsic tuff; mafic flows</td>
<td>east-northeast-striking foliation</td>
<td>remobilized po, py, cp, sp on fractures and in brecciated zones</td>
<td>diss. ath, ser grt, st, bio in felsic rocks; grt-amp veins</td>
</tr>
</tbody>
</table>

foliation. Some sulphide mineralization occurs in quartz veins and in late, retrograde chlorite and epidote veins. Felix (1996) reported anomalous enrichment of metals in altered metavolcanic rocks such as 0.1% Cu, 346 ppm Zn and 17.7 g/t silver across 23.3 m in diamond-drill hole PO–96–12. Numerous, narrow mineralized sections, generally less than 2 m, have also been reported (Edwards 1991) such as 2.42% Cu, 0.11% Zn and 24 g/t silver across 0.66 m; and 1.14% Cu, 4.42% Zn and 24.9 g/t silver across 0.36 m.

Maskooch Lake—Birch–Uchi Greenstone Belt

The Maskooch Lake occurrence is located about 5 km east of Slate Lake in the southeast corner of the Birch–Uchi greenstone belt (see Figure 22.1). The metavolcanic rocks at Maskooch Lake have not been dated, but are interpreted (Ontario Geological Survey 1992) to be part of the Confederation assemblage deposited within the age range 2.8 to 2.73 Ga. The assemblage may represent proximal and vent facies arc volcanism (Stott and Corfu 1991).

The occurrence is situated in a sequence of east-striking, intermediate to felsic flows and pyroclastic rocks interlayered with mafic metavolcanic flows and thin discontinuous units of quartz-magnetite and sulphidic iron formation (Bowen 1989). The supracrustal rocks are intruded by the Maskooch Lake stock, a late quartz porphyritic diorite to granodiorite with clotty aggregates of black amphibole.

Trenches have been sunk on a 5 to 6 m wide, generally east-striking unit of felsic tuff interlayered with siltstone and recrystallized, granular, brecciated and sericitized chert. The chert hosts disseminated pyrrhotite, pyrite and chalcopyrite; coarse-grained massive pyrrhotite and chalcopyrite; and pyrite nodules. Some sulphide mineralization is remobilized into late, glassy quartz veins. The chert and/or tuff horizon is Z-folded and plunges about 40° to the west-northwest. Widespread disseminated chalcopyrite, pyrite and pyrrhotite mineralization associated with extensive sericitization also occurs throughout the metavolcanic rocks at Maskooch Lake. Sampling by the author in 1991 demonstrated that these altered rocks consistently contain anomalous metal values ranging from 380 ppm to 1.18% Cu, 42 to 165 ppm Zn, 11 to 2940 ppb Au and trace to 0.37 ounce silver per ton (Atkinson et al. 1992).

Recent mechanical stripping south of Maskooch Lake has exposed synvolcanic amphibole-garnet-magnetite alteration. Bowen (1989, p.14) recognized the alteration and mapped it as autoclastic breccia over an area that is 5 km long and 400 m wide.

The incipient stages of the alteration consist of a chaotic network of intersecting fractures, less than 1 cm wide, in felsic metavolcanic rocks that have been brecciated or hydro-fractured into angular, shard-like fragments (Photo 22.1a). The incipient alteration imparts a distinct “jigsaw-fit” appearance to the rocks. Moderate to strong alteration consists of wide (greater than 5 cm) amphibole-filled fractures with subrounded to subangular brecciated metavolcanic “clasts” (Photo 22.1b) where angular corners on “clasts” have been corroded and rounded-off during alteration. The most intense alteration consists of widely scattered “clasts” in a dark green to black, amphibole-garnet-magnetite ± pyrrhotite ± chalcopyrite ± pyrite matrix. The alteration imparts a “pseudo”-fragmental appearance to the metavolcanic rocks making it difficult to identify original pyroclastic rocks. A large stripped exposure consists of giant subrounded “clasts” up to 3.8 m long and 1.6 m wide that appear to be brecciated and altered pillowed, mafic metavolcanic flows (?). The altered rock may also appear heterolithic where small “clasts” have been replaced by amphibole and have acquired a mafic appearance. Garnet occurs in veins or dense, fine-grained aggregates and commonly replaces small felsic “clasts”. Larger “clasts” have garnetiferous reaction rims and contain disseminated needles of amphibole and small, disseminated, garnet porphyroblasts. Silicified zones, exhibiting positive weathering, form irregular, interconnected patches within the amphibole-garnet-magnetite alteration. The alteration is overprinted by a strong foliation, shearing and folding.

Recent prospecting has identified abundant anthophyllite-garnet alteration in felsic metavolcanic rocks

Photo 22.1. Examples of synvolcanic alteration at Maskooch Lake: a) incipient amphibole alteration on fractures in a hydro-fractured felsic metavolcanic rock with a distinct “jigsaw-fit” texture; b) moderate to intense amphibole alteration with wider fractures and rounded edges on some “clasts”. Remnant “jigsaw-fit” textures are visible in the lower left corner of the photograph.
about 2 km west of Maskooch Lake as well as new occurrences of sulphide mineralization (J. Williamson, P. English, Prospectors, personal communication, 1998).

The Confederation assemblage hosts significant VMS mineralization at Confederation, Fly and Fredart lakes (see Figure 22.1) and includes the past-producing South Bay copper-zinc-silver massive sulphide deposit which produced 1.6 million tons of ore with an average grade of 1.8% Cu, 11.06% Zn and 2.12 ounces silver per ton (Atkinson et al. 1991). The South Bay area is dominantly underlain by felsic pyroclastic rocks and flows intruded by late and synvolcanic felsic intrusions. The metavolcanic sequence includes Type III felsic metavolcanic rocks in VMS-productive successions that are 2,738 billion years old (Fyon et al. 1992; Stott and Corfu 1991).

The South Bay area is characterized by extensive aluminous alteration consisting of andalusite ± sillimanite ± kyanite veins with more localized anthophyllite-cordierite-garnet ± biotite alteration associated with zinc-copper sulphide mineralization. Recent exploration by Noranda Exploration Company Limited has delineated several new mineralized zones consisting of high grade zinc mineralization with significant gold credits (A. Smith, Project Geologist, Noranda Exploration Company Limited, personal communication, 1998).

SUMMARY AND DISCUSSION

The majority of VMS mineralization in early (pre-2.74 Ga) metavolcanic assemblages, examined during this study, are hosted by chemical and clastic metasedimentary rocks interlayered with dominantly mafic metavolcanic flows and subordinate, intermediate to felsic pyroclastic rocks. The majority of sulphide mineralization is zinc, lead and silver-rich and has been deposited in distal, deep- to shallow-water, oceanic environments characterized by non-explosive effusive volcanism and deposition of clastic and chemical sediments. These depositional environments experienced periods of volcanic and tectonic inactivity that facilitated the accumulation of chemical sediments and sulphide mineralization originating from exhalative centres on the seafloor. Volcaniclastic, matrix-supported conglomerates and poorly sorted, matrix-supported intermediate to felsic pyroclastic rocks are spatially associated with sulphide mineralization. These rocks were derived from previously lithified or solidified volcanic rocks and deposited as subaqueous debris flows.

The majority of VMS mineralization in late (2.74 to 2.7 Ga) metavolcanic assemblages, examined during this study, are hosted by intermediate to felsic metavolcanic rocks in a tectonically active environment of deep-water, proximal, vent facies arc volcanism. These depositional environments are characterized by submarine eruptions resulting in massive felsic flows and angular breccias (Cas 1992), however, relatively few felsic flows were identified during this study except in the South Bay mine area. Fine to coarse, dominantly heterolithic, matrix-supported pyroclastic rocks are common and may represent subaqueous debris flows. Periods of volcanic quiescence allowed the deposition of fine felsic tuffs, sulphidic argillites and chert and quartz-magnetite and sulphidic iron formation. The majority of sulphide mineralization is zinc- and copper-rich although significant quantities of lead and silver are also present.

Alteration associated with sulphide mineralization was formed prior to D1 deformation in the early history of the metavolcanic assemblages. Widespread, semi-conformable alteration such as sericitization, disseminated sulphide mineralization, silicification or aluminous alteration; and more proximal iron- or magnesium-rich alteration has been recognized in both early and late Neoarchean assemblages. Extensive kilometre-scale, fracture-controlled iron and/or magnesium-rich alteration occurs over large areas in metavolcanic rocks of both early and late ages (e.g., Discovery Lake, Arseno Lake, Maskooch Lake).

RECOMMENDATIONS FOR EXPLORATION

Some important factors (based on empirical field observations) that might be useful in mineral exploration are:

1. Chemical metasedimentary rocks, such as chert, quartz-magnetite and sulphidic iron formation are important host rocks for sulphide mineralization in early and late Neoarchean metavolcanic assemblages.

2. Sulphide mineralization in early Neoarchean assemblages is commonly situated along contacts between mafic metavolcanic flows and clastic or chemical metasedimentary rocks.

3. Extensive hydro-fracturing that produces “brickwork” or “jigsaw-fitting” alteration consisting of amphibole ± garnet ± magnetite ± biotite is commonly associated with VMS mineralization in early and late Neoarchean assemblages.

4. Subaqueous debris flows or lahars derived from previously lithified volcanic rocks are spatially associated with sulphide mineralization in early and late Neoarchean assemblages. Lahars may be genetically important to massive sulphide deposits as source rocks for metals (Franklin 1976).

5. Deposits of sulphide mineralization in early Neoarchean metavolcanic assemblages may be relatively small due to the absence of significant felsic volcanism.

6. The Shrimp Lake, Discovery Lake, Maskooch Lake and Arseno Lake areas are relatively unexplored and have the potential to host VMS mineralization based on their metavolcanic-metasedimentary lithologies; presence of significant alteration; presence of sulphide mineralization; and in the case of Shrimp Lake, the presence of a synvolcanic intrusion.

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Riley, R.A. 1968. Precambrian geology, Muhlay Township; Ontario Department of Mines, Map 2295, scale 1:12 000.


INTRODUCTION

The Sioux Lookout orogenic belt is a relatively small segment of the vast network of Late Archean (2.7 Ga) supracrustal rocks and associated intrusive rocks that comprise the “western Wabigoon region” of the Wabigoon Subprovince (Blackburn et al. 1991). This report is a follow-up to preliminary reports from recent reconnaissance-scale bedrock mapping of the Sioux Lookout orogenic belt (Devaney and Borowik 1994; Devaney, King and Babin 1995, Devaney et al. 1995; Devaney and Babin 1996). The results and interpretations presented herein are an attempt to place the mineral occurrences of the Sioux Lookout orogenic belt in a tectono-stratigraphic context. This revised and updated knowledge of the Sioux Lookout orogenic belt should help to guide mineral exploration in the area and thus reduce exploration risk.

Figure 23.1 presents a series of cross-sectional cartoons outlining the interpreted sequential development of the Sioux Lookout orogenic belt; different episodes of potentially economic mineral deposition correspond with the distinct stages of volcanism, sedimentation and tectonism. Figure 23.2 and the accompanying text briefly describe and interpret the key features of the Sioux Lookout orogenic belt in map view. Figure 23.3 suggests that there may be an orderly regional structural pattern of gold occurrences. Many of the ideas listed below need to be tested via detailed structural mapping and geochronology.

EVOLUTIONARY STAGES OF THE SIOUX LOOKOUT OROGENIC BELT

Figure 23.1 attempts to outline the main stages of development of the Sioux Lookout orogenic belt in cross-section. The earlier the stage, the more speculative its interpretation is. The series of stages are time intervals spaced about 10 to 15 Ma apart. Note that the cartoons are not to scale and represent different scales. For simplicity, basinal water levels are not shown (deposits during Stages 1, 2 and 4 were subaqueous, Stage 3 deposits were partly subaerial, and Stage 5 and 6 deposits were mostly subaerial fluvial). Qualifying statements and explanatory comments regarding each of the cartoon views of the stages are listed below. Because of both the flat, penneplained type of Shield exposure and the stratigraphic reconnaissance nature of this project, three-dimensional fault kinematic indicators are not available; interpretations such as Stage 6 sinistral faulting are based on large-scale map patterns and, to a far lesser degree, small-scale structures. (Use of the term “orogenic belt” rather than “greenstone belt” is intended to draw attention to the similarities the tectonized Sioux Lookout orogenic belt has with many post-Archean orogenic belts.)

Stages 0, 1

The hypothesis that the Winnipeg River Subprovince (Beakhouse 1991) and central Wabigoon region (Blackburn et al. 1991) were formerly joined as a (micro-)cratonic block and were subsequently rifted apart (see Figure 23.1; see also Blackburn 1980) is admittedly highly speculative; presently there is no match of specific geological features between the two areas, and there are no known rift sediments or faults. (Could any riftiing have been subaqueous, with no subaerial erosion and no alluvial fan-fluvial sedimentation?) The vertical wall between the cratonic areas and the basinal basaltic pile is a diagrammatic simplification.

Massive to pillowed tholeiitic basalt flows and associated facies of the Northern Volcanic Belt (or Botham Bay Group: Trowell and Johns 1986) and Southern Volcanic Belt (Berger 1989) are predominantly deep water deposits (mafic plain-type assemblage of Williams et al. 1992, Table 25.2; lower mafic sequences of Blackburn et al. 1991). Trace element patterns indicate transitional mid ocean ridge basalt to volcanic arc basalt (MORB–VAB) compositions (see “Volcanic geochemistry highlights,” below). It has been suggested by Ayer (1991) that the basalts may have floored a back-arc basin, but there is no known arc succession of appropriate age for the basalts to have been behind (in “back” of), so the term “pre-arc” basin basalts (Pearce et al. 1984) may be more accurate.

Exhalite deposition likely produced the interlayered magnetite and pyrite iron formation units, which include the site of the North Pines pyrite mine of the 1920s (Johnston 1972).

Stage 2

There are distinct geochemical differences between the Central Volcanic Belt (or “Neepawa Group”) enriched-mid ocean ridge basalts (E-MORB basalts) and the Northern Volcanic Belt MORB basalts; e.g., TiO₂ content
STAGES IN THE EVOLUTION OF THE SIOUX LOOKOUT OROGENIC BELT

late orogenic stage:

6 compression/transpression, verticalization, NVB overturned, later granitic plutonism (2895 Ma), sinusoidal strike-slip faulting and offsets, WRS as Laramide-type uplift, CWR as block or thick-skin thrust

early orogenic stage:

5 fold-and-thrust belt (compression to NW or N) < 2710 Ma, imbricate thrust stack, strata dip and young to SE

late arc stage:

4 intermediate to felsic volcaniclastic stage (2719-2712 Ma)

middle arc extensional stage:

3 calc-alkaline intermediate volcanism (subvolcanic dioritic pluton: 2732 Ma)

early arc stage:

2 tholeiitic basalt (felsic tuff: 2733 Ma)

rift to ocean stage:

1 tholeiitic basalt, narrow ocean basin or pre-arc basin

precursor stage:

0 "Ancient Gneiss Continent" (Blackburn, 1980)

Acronyms: BIF=banded iron formation; BLB=Basket Lake Batholith; CVB=Central Volcanic Belt; CWR=Central Wabigoon Region; NVB=Northern Volcanic Belt; SVB=Southern Volcanic Belt; WRS=Winnipeg River Subprovince; all diagrams schematic, not to scale; U-Pb dates (+/- few Ma) from Davis et al (1988) and Davis (1990)

Figure 23.1. (a) Cross-sectional cartoon views of the interpreted evolutionary stages of the Sioux Lookout orogenic ("greenstone") belt. (b) Corresponding stages of mineralization.

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STAGES OF MINERALIZATION

6, late orogenic stage: late quartz or quartz-carbonate veins with erratic gold values; most Au occurrences are within the Neepawa (CVB) and Minnitaki groups and are distributed along two strike-parallel ENE-trending lineaments (this fact previously unrecognized); to the SW, the former Goldlund gold mine and the current Teck-Corona Thunder Lake gold prospect (in Zealand Twp.) are located along these two lineaments.

5, early orogenic stage: quartz and/or carbonate veins

4, late arc stage: iron formation units within the sedimentary Minnitaki Group and the Big Vermilion-Daredevil volcanic unit (minor chemical sediments in volcaniclastic sediments): magnetite, pyrite and chert exhalites, and related synvolcanic hydrothermal alteration (sub-exhalite) in the underlying upper Minnitaki Group (including volcanic dome facies) and horizons of the Big Vermilion-Daredevil unit.

3, middle arc extensional stage: within the northeast CVB, merely incipient porphyry copper-type mineralization (rare Cu-Mo in and marginal to an intermediate subvolcanic intrusion) and no evidence of epithermal mineralization in a genetically linked andesitic stratovolcano succession

2, early arc stage: minor exhalite layers

1, rift to ocean stage: iron formation units within the basaltic NVB: magnetite, pyrite and chert exhalites from synvolcanic hydrothermal flow

Figure 23.1. continued.
Figure 23.2. Semi-schematic map of the Sioux lookout orogenic belt. Discussions of sites 1 to 27 are given in the text. WRS = Winnipeg River Subprovince, NVB = Northern Volcanic belt, Big V-Dd = Big Vermilion–Daredevil unit, CVB = Central Volcanic belt, MG = Minnitaki Group, SVB = Southern Volcanic belt, BLB = Basket Lake batholith.
(see "Volcanic geochemistry highlights," below). In rocks to the southwest, Berger (1989) noted similar differences between the basalts of the Southern Volcanic Belt and Central Volcanic Belt.

The 2734 to 2732 Ma age reported for the Central Volcanic belt (Davis et al. 1988, Blackburn et al. 1991, Figure 9.40) is from a felsic tuff interlayered with basalt; unless the dated tuff is a thrust basal layer of the Central Volcanic belt, some of the underlying Central Volcanic belt basalts must be older than this (and thus also older than the 2734 to 2730 Ma age of Stage 3).

Stage 3

A partly subaerial basaltic andesite stratovolcano succession about 3 km thick appears to lie on a Stage 2-type basaltic platform (Devaney and Babin 1996; see Figure 23.1 and Figure 23.2, sites 15, 16). The comagmatic dioritic subvolcanic pluton (see Figure 23.2, site 16) has a preliminary age of 2734 to 2730 Ma (D. Davis, written communication, 1998).

The aerially restricted (non-regional) distribution of andesitic rocks (see Figure 23.2) offers admittedly weak support for the interpretation of a fault-bounded local volcanic centre (e.g., a graben in Figure 23.1), but the presumed presence of a concentration of synvolcanic faults prone to later (Stage 5, 6) tectonic reactivation would help account for the abnormal orientation of a large (7 km) ovoid part of the andesites (see comments on site 15 in Figure 23.2, below).

Relationship of Sioux Lookout Orogenic Belt Stage 3 Deposits to Correlative Volcanogenic Massive Sulphide Deposits at Sturgeon Lake

Given that the Sioux Lookout orogenic belt andesitic stratovolcano succession and its feeder pluton (2730 to 2734 Ma) can now be correlated with the caldera-hosted VMS deposits and subvolcanic intrusion (ca. 2734 Ma) 60 km to the east at Sturgeon Lake (same ages, similar subvolcanic pluton geochemistry, but significantly different volcanic facies: Morton et al. 1991, Blackburn et al. 1991), it is suggested that extensional volcanic regimes developed at both sites (see Figure 23.1). Indeed, a "rifted arc" scenario such as this is commonly thought to be favourable for the generation of VMS deposits (e.g., Syme et al. 1996).

The partially andesitic volcanic units or centres of similar age (about 2733 Ma) and character (lithology, paleoenvironment, geochemistry) that are sporadically distributed throughout western Wabigoon Subprovince (Edwards and Davis 1984; Berger 1990, 1991; Blackburn et al. 1991) may have a common origin, perhaps reflecting a regional extensional event that developed only local areas of extension, transtension or partial rifting. Within this group of supposedly tectono-stratigraphically related areas that might have elevated potential for VMS deposits, the south Sturgeon Lake belt is presently the only known area in western Wabigoon Subprovince where conditions

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**Figure 23.3.** The proposed lineament pattern joining gold occurrences (black dots) in the Sioux Lookout orogenic belt (and equivalent rocks to the southwest) resembles a large-scale conjugate Riedel shear set (theory shown in inset; major stratigraphic and fault surfaces trend at about 64°).
were suitable for the formation and preservation of economic VMS deposits (e.g., a subaqueous caldera as a restricted basin; the Biedelman Bay intrusion as an adequately large and deep subvolcanic heat source for the hydrothermal system). It is suggested that the ca. 2733 Ma volcanic-subvolcanic systems in the Sioux Lookout orogenic belt and south Sturgeon Lake belt formed in coeval rifted arc basins (Sioux Lookout orogenic belt Stage 3 in Figure 23.1).

Stage 4

Stage 4 basal deposits (see Figure 23.1) record a continuum of volcanic to sedimentary processes, with the latter being more common. In both the northern and southern areas (see Figures 23.1, 23.2): proximal lava flow or dome facies (Trowell et al. 1983, Page 1984) are rare; unworked pyroclastic tuffs (Devaney and Borowik 1994, Devaney King and Babin 1995) are extremely rare (felsic tuff dated at ca. 2719 to 2713 Ma: Davis et al. 1988, Table I); magnetite iron formation signifies volcanogenic exhalative chemical sedimentation during eruptive quiescence; thin bedded turbiditic wackes (Walker and Pettijohn 1971) are very common (they are ubiquitous to the south, in the Minnitaki Group, versus more variable degrees of sedimentary reworking in other units to the north); resedimented conglomerates are interbedded with the turbiditic wackes and iron formation. The conglomerates are composed of intrabasinally-sourced felsic volcanic and/or felsic porphyry clasts (quartz porphyry clast dated at 2714 to 2711 Ma: Davis et al. 1988).

The names used for these Stage 4 units have been a source of stratigraphic terminological confusion. To the north, the correlative “Big Vermilion–Daredevil unit” (new informal term) and Daredevil Formation are exposed in different fault slices (see Figure 23.2, sites 7, 10). (The Little Vermilion Formation of Turner and Walker (1973) is not a readily mappable unit and has been grouped herein with these northern units.) A rock containing zircons ranging from 2720 to 2702 Ma (“best analysis”) is 2712 Ma: Davis et al. 1988, p.183) was considered by Davis et al. (1988) to be from the “Pataru Formation” of Johnston (1972), but is here considered part of the eastern “Big Vermilion–Daredevil unit”. (The Pataru Formation is described herein as a unit of very different character, consistent with Pettijohn’s (1934) original definition.) A felsic pyroclastic rock from the western Minnitaki Group has been dated at 2707 to 2701 Ma (Davis 1990, Blackburn et al. 1991, Figure 9.40), based on a single zircon grain. However, the dated sample is from either the southern central volcanic belt or the northern Minnitaki Group and has lithological and geochemical similarities with Stage 4 rocks, which suggest that it is a Stage 4 deposit more than 2710 Ma old. The revised, simpler and more linear stratigraphy presented herein is supported by mapping (Devaney et al. 1995) and alternative applications of the geochronological data.

With regard to the interpreted volcanic arc context (Stages 2 to 4), any accretionary prism and subduction zone was likely to have been located beyond the south margin of the Wabigoon Subprovince.

Volcanic Geochemistry Highlights

Contents of SiO₂ versus TiO₂, P₂O₅, Zr and Zr/Y provide useful stratigraphic subdivisions of the metavolcanic rock samples. Based on plots of SiO₂ versus TiO₂, P₂O₅, La, Nb, Th, Zr and Zn, the basaltic subunits of the Sioux Lookout orogenic belt form two discrete groups; strongly similar compositions allow the correlation of the Northern Volcanic belt and the northern Central Volcanic belt basalts (as one group), versus the distinctly different (higher Ti, P, La, Nb, Th, Zr and Zn content), more southern (probably younger) Central Volcanic belt basalts.

Basalts of the Northern Volcanic belt and northern Central Volcanic belt have flat REE plot slopes and REE abundances at about 10x chondrite level. More southern Central Volcanic belt basalts have low REE plot slopes (chondrite-normalized La/Yb ratios of about 6) and abundances of about 40 to 70x chondrite level; these basalts are more evolved (E-MORB) and fractionated than the primitive (transitional MORB-VAB) basalts of the Northern Volcanic belt and north Central Volcanic belt. Most of the intermediate to felsic volcanic rocks display high REE plot slopes with chondrite-normalized abundances of about 100x La to 10x Yb.

Use of tectonic discrimination diagrams (Wood 1980, Pearce 1996) suggests the following progression: Stage 1 tholeiitic MORB-VAB seafloor basalt (pre-arc spreading); Stage 2 tholeiitic to more variable (E-MORB) basalt, transitional oceanic to juvenile island arc; Stage 3 calc-alkaline (transitional to tholeiitic) basaltic andesite, variable within-plate to island arc character (for rifted arc interpretation, see above); and Stage 4 calc-alkaline dacite-rhyolite of mature island arc character.

On extended trace element plots normalized to N-MORB (normal-MORB) composition, Stage 1 and 2 basalts display minor negative Nb anomalies, versus larger negative Nb anomalies (arc signature) in Stage 4 dacitic-rhyolitic rocks.

Thus the geochemical signatures corroborate the inferred tectono-stratigraphic scenarios in Figure 23.1 and the thrust repetition of the basaltic lower part of the basin (see Figure 23.2).

Stage 5

The predominantly layered architecture of the Stage 4 basin was compressed and segmented into a north- to northwest-vergent fold-and-thrust belt (see Figure 23.1), with the Winnipeg River Subprovince as a microcontinental foreland and the Central Wabigoon Region of the Wabigoon Subprovince as a hinterland block or thick-skin thrust. This interpretation is based on geochronology (in parallel units with the same prevalent younging direction, two cases of older units juxtaposed over younger ones), map patterns (see Figure 23.2), and field geology (e.g., faults in the Sioux Lookout orogenic belt are narrow, highly deformed zones in a belt typified by supracrustal rocks with well preserved primary features). Beakhouse
(1988) and Davis et al. (1988) previously offered generalized interpretations of the thrusted nature of the Sioux Lookout orogenic belt.

The Stage 5 cartoon (see Figure 23.1) is schematic; it is not a balanced cross-section. The nature of the sub-basaltic basement is unknown. Presumed mountainous topography is not shown, for ease of comparison with the Stage 6 cartoon.

The conglomeratic Ament Bay Formation (Petitjohn 1934, Turner and Walker 1973, Devaney and Borowik 1994) contains a granitic clast dated at 2702 to 2694 Ma (Davis et al. 1988), establishing a maximum age for deposition of at least some of the formation. The erosion of a presumably high-relief thrust belt exposed numerous layers of different rock types, resulting in the high polymictic composition of clasts in the Ament Bay Formation. Braided rivers transported sand and gravel to an intermontane basin, probably a piggyback basin atop a thrust slice (see Figure 23.1; e.g., Steidtmann and Schmitt 1988). The molassoid conglomerates were then overridden during continued thrusting (confirmed by geochronology; see Figure 23.2, sites 10, 11).

The north to northwest limits of sedimentation in the Stage 4 and 5 cartoons (see Figure 23.1) have been drawn conservatively, according to what has been preserved; extensions onto the Winnipeg River Subprovince (e.g., Cruden et al. 1997) are speculative. Detrital zircons in metawacke indicate sedimentation at a site in the Winnipeg River Subprovince during the ca. 2717 to 2700 Ma period (Cruden et al. 1997). It is uncertain whether this age signifies the northwest continuation of the Sioux Lookout orogenic belt Stage 4 volcaniclastic basin or a Stage 5 foreland basin, or some other type of basin coeval but unrelated to the Sioux Lookout orogenic belt.

Stage 6

Interpretation of late orogenic, Stage 6 sinistral strike-slip faulting is based on geometric forms of sub-regional map patterns (mostly "lazy S"-shaped sigmoidal forms; see Figure 23.2, sites 2, 3, 6, 15) which appear to postdate the Stage 5 thrust belt structures. Northward movement of the Wabigou Indenter (Stott and Corfu 1991), a large crustal block now exposed as the Central Wabigou Region of the Wabigou Subprovince, resulted in sinistral faulting along the indenter's western flank, where the Miniss River Fault and the Sioux Lookout orogenic belt are located (i.e., the Miniss River Fault forms a diffuse sinistral fault zone throughout the northern half of the Sioux Lookout orogenic belt).

Note: in the Stage 6 cartoon (see Figure 23.1), the hook-shaped forms with north-younging black arrows show mapped younging reversals interpreted to have been produced by hangingwall anticlimes (Stage 5 features?), and the sill-like form of granitoid intrusions is wholly conjectural. The Basket Lake batholith is poorly known; Szewczyk and West (1976) proposed that it is a 8 km deep intrusion.

In order to account for the widespread overturning of the Northern Volcanic belt, two hypotheses are suggested. The south margin of the Winnipeg River Subprovince (not studied in this project) may be a south-verging, Laramide-type thrust uplift (probably an oblique-slip fault, given the regional sinistral faulting), which could have rotated the north marginal strata of the Sioux Lookout orogenic belt away from the foreland, similar to the rotation and overturning seen adjacent to Laramide-type uplifts (e.g., Erskie 1986). Similar overturning in correlative boundary units far to the southwest was described by Ciceri et al. (1996) and Cruden et al. (1997). Along the boundary to the northeast, the "northwestern hanging-wall of the north-east-trending, sinistral, oblique-slip Miniss River Fault... brings to the surface granulate facies supracrustal gneisses" (Williams et al. 1992, p.1289; see also Beakhouse 1991, p.297). South- to southeast-directed reverse faulting of the north margin of the Sioux Lookout orogenic belt, between the above two areas, seems reasonable given this regional context.

Alternatively, a second hypothesis suggests that steeply oriented strata near the belt boundary may have been deformed into large flower structures by transpression (Syvester 1988) in which slices of the basaltic Northern Volcanic belt were thrust upward and outward (south-verging slices would be overturned). Regardless of the tectonic mechanism, uplifted basaltic areas supplied basaltic clasts and a few iron formation clasts (a nearly oligomictic composition) to the adjacent Patara strike-slip basin (see Figure 23.2, site 6). The relatively shallowly dipping, predominately north-dipping and north-younging Patara succession contrasts sharply with the other sedimentary units in the Sioux Lookout orogenic belt.

The orogenic scenario of Stage 5 thrusting followed by Stage 6 sinistral wrench faulting (see Figure 23.1) largely agrees with the results of the detailed structural study of Chorlton (1990, 1991) in the area adjacent to the southwest of the present Sioux Lookout orogenic belt study area. In this southwest area there is a kink-like bend in the regional layering of the supracrustal belt, so the possibility that 2699 to 2693 Ma and 2697 to 2693 Ma granitic plutons (Crosseau, Sandybeach: Davis 1990) may have formed in large releasing bends or a northwest-trending kink-band mega-structure should be investigated.

Postorogenic granitic and pegmatitic intrusions in areas near the Sioux Lookout orogenic belt are as young as ca. 2650 Ma (Larbi et al. 1996), and represent a potential seventh stage in the region's development.

HIGHLIGHTS OF THE SIOUX LOOKOUT BELT IN MAP VIEW

Figure 23.2 is a simplified sketch map of most of the Sioux Lookout orogenic belt study area (approximately 1000 km²; Devaney et al. 1995b). It is a semi-schematic cartoon view with approximate scale (some north-south expansion). As outlined in Figure 23.1 and below, the map pattern is interpreted to represent a pre-arc to arc succession compressed into a predominantly south-younging, steeply south-dipping fold-and-thrust belt which was
subsequently significantly modified by strike-slip faulting.

The most geologically important features of the area are labelled as sites 1 to 27 in Figure 23.2 and are discussed in sequence in the correspondingly numbered paragraphs below.

1. The boundary between the microcontinental Winnipeg River Subprovince and the north margin of the Northern Volcanic belt is drawn as a thrust fault to suggest the possible presence of a Laramide-type reverse fault; alternatively, large flower structures may be present (see "Stage 6," above).

2. In combination with regional sinistral faults (Stott and Corfu 1991) and nearby features suggestive of sinistral faulting (see 3, 6 below), the sigmoidal form of the iron formation units is thought to have been produced by sinistral strike-slip faulting within the Northern Volcanic belt. Also, structurally weak iron formation units are likely to have been decollement surfaces during earlier thrust stacking of the basaltic Northern Volcanic belt. (Notably, this presumed intra-Northern Volcanic belt thrust stacking (see Figure 23.1) occurs where the Northern Volcanic belt is widest and/or thickest)

3. Using concepts outlined by Hutton (1988), Glazner (1991) and D'Lemos et al. (1992), the teardrop shapes (interpreted as incompletely exposed sigmoidal shapes) of late granitic plutons mapped by Johnston (1972) suggest the interpretation that these plutons formed in the releasing bend (dilational jog) areas of sinistral strike-slip faults and are thus loosely analogous to large-scale tension gash infillings.

4.5. Large-scale structural features in the Big Vermilion Lake area (Johnston 1972, Page 1984) are ambiguous; basaltic strata (5) may be an anticlinal fold core or the base of a north-verging thrust, and a poorly exposed stratigraphic contact may have been faulted (4). (Presently there are no kinematic data with which to constrain these suggestions.)

6. The Patara Formation (Pettijohn 1934) is a 20 km long, sigmoidally shaped stratigraphic unit comprised of basalt-clast fluvial conglomerate and related finer metasediments. The best cross-sectional transect shows a north-dipping, north younging, coarsening-upward sequence (good top indications present) from aquabasinal (lacustrine?) mudstones and graded beds to fluvial sandstone to boulder conglomerate. Pillowed metabasaltic rocks of the Northern Volcanic belt immediately adjacent to the north are south-younging and highly overturned (north-dipping). These features allow the interpretation that the Patara Formation is the infilling of a sinistral strike-slip (to oblique-slip) pull-apart basin, with clasts supplied from the Northern Volcanic belt to the north. If this interpretation is correct, the pull-apart basin is a tectonically late feature superimposed on a previously tectonized part of the Sioux Lookout orogenic belt. (Note that this contrasts sharply with previous views that the Patara metasedimentary rocks were the oldest sedimentary unit in the Sioux Lookout orogenic belt and were a south-younging unit lying unconformably on top of the Northern Volcanic belt.)

7. In contrast with previous mapping (Johnston 1972), the "Big Vermilion-Daredevil" unit ("Big V-Dd" on Figure 23.2; likely equal to the "Redhat volcanics" of Page 1984) has been mapped as a much more laterally continuous, tectonized, formation-scale unit of intermediate-felsic, predominantly fragmental metavolcanic rocks (Devaney et al. 1995). Small iron formation units (at the two "7"s in Figure 23.2) signify at least partly subaqueous deposition for the unit. Dates from zircon grains (Davis et al. 1988) demonstrate correlation with the type area of the Daredevil Formation (Pettijohn 1934, Turner and Walker 1973) to the south (see 10, below).

8. Coarse sandstone and highly polymict conglomerate of the Ament Bay Formation (Pettijohn 1934, Turner and Walker 1973, Devaney and Borowik 1994) are braided river deposits interpreted to have formed in a linear, strike-parallel, intra-orogenic, syn-thrust piggyback "molasse" basin (see "Stage 5," above). Structural features of the Ament Bay Formation such as south-younging beds, steep dips, and schistosity (oblate clasts) are typical of most of the supracrustal units in the Sioux Lookout orogenic belt and contrast sharply with the Patara Formation (see 6, above).

9. The origin of a separate minor unit of deformed polymict conglomerate located to the northeast is less certain (thrust-related, like 8, above, or a small, local strike-slip basin?)

10. Geochronological data by Davis et al. (1988), although discordant, suggest that south-younging beds of the Daredevil Formation (type area) are older than the south-younging Ament Bay Formation adjacent to the north (separated by a sharp linear contact). If this age interpretation is valid it demonstrate out-of-sequence stratigraphy (explained via thrusting) and invalidates the previous stratigraphic schemes of Pettijohn (1934), Turner and Walker (1973) and Blackburn et al. (1991, Table 9.6). (The "Abram Group" was thought to consist of, from oldest to youngest, the Ament Bay, Daredevil and Little Vermilion formations. However, because the Ament Bay Formation is now known to be the youngest of the three formations, the term "Abram Group" can no longer be considered valid.)

11. Geochronology by Davis et al. (1988; Blackburn et al. 1991, Figure 9.40) demonstrate out-of-sequence stratigraphy and north-verging thrusting of the Central Volcanic belt over the younger Daredevil and Ament Bay formations (see 10 and 8, above). Metagabbroic intrusions preferentially located toward the north margin of the Central Volcanic belt suggest that such gabbro bodies become more abundant downward in the stratigraphic section of the Central Volcanic belt. It is suggested that the basal thrust of the Central Volcanic belt formed above a stratigraphic level welded together (i.e., made structurally competent) by gabbro intru-
sions and below a stratigraphic level of thinly layered basaltic flows.

12. Small felsic (ryholitic) volcanic bodies in the Central Volcanic belt (volcanic centres?) are undated.

13. The display of faults in this area (Pickerel Arm: Johnston 1969, 1972) on Figure 23.2 is highly conjectural (interpretation of basalt-based north-verging thrusts). Felsic porphyry bodies are near notable quartz vein-hosted gold occurrences (Johnston 1969, Devaney King and Babin 1995).

14. Western limit of andesitic pyroclastic rocks, mapped as anticlines and synclines by Page and Clifford (1977) and Page (unpublished data).

15. Within an ovoid area about 7 km in diameter, a basaltic andesite, partly subaerial, west-younging stratovolcanic succession about 3 km thick lies on a basaltic platform (Page and Clifford 1977, Devaney and Babin 1996, Devaney 1998). Within the predominantly south-younging structures of the Sioux Lookout orogenic belt, the west-younging orientation of this succession is anomalous. It is interpreted that during sinistral strike-slip faulting of the north half of the Sioux Lookout orogenic belt (Stage 6) this 7 km diameter area acted as a distinct structural domain; based on the structural examples of Gallo et al. (1980) and Sengor et al. (1985, Figure 14), it is interpreted that the previously tectonized 7 km area was sinistrally rotated counter-clockwise (presumably above a detachment surface) as a “Riedel flake.”

16. The very similar major and trace element composition of the dioritic to tonalitic, poorly mineralized (copper and molybdenum) pluton and the adjacent andesitic strata (see 15, above) show that they are comagmatic and the pluton is thus a subvolcanic intrusion (Page and Clifford 1977, Devaney and Babin 1996; Northeast Bay pluton of Trowell et al. 1983). As discussed above (under “Stage 3”), this pluton is the same age as, and of very similar composition to, the Biedelman Bay intrusion 60 km to the east at Sturgeon Lake (Poulsen and Franklin 1981, Morton et al. 1991, Blackburn et al. 1991).

17. Mapping by Page and Moller (1979) shows andesitic strata to the northeast (east of margin of Figure 23.2), suggesting a 7 km sinistral offset of a lithological marker unit.

18. The presence of faults (possible basalt-based thrusts?) may explain the north-south transect of alternating basaltic and wall units in this part of the Minnitaki Group (MG; western Minnitaki Lake area: Johnston 1969, Devaney King and Babin 1995, Devaney et al. 1995; see Stage 5 and 6 cartoons in Figure 23.1). If the view of thrust-repeated basalt-wall couples is correct, this calls into question the concept of “upper mafic sequences” (Blackburn et al. 1991) in the western Wabigoon region.

19. Laterally continuous (1 to 2 km long), oppositely facing zones with consistent top directions of graded beds suggest the presence of upright folds (syncline-anticline pairs).

20. Both clast size and bed thickness decrease westward along strike in the deep-water turbidite conglomerates of the Minnitaki Group, demonstrating a proximal to the east versus distal to the west polarity.

21. In contrast with a unit of volcanic rocks present along strike to the east (at east margin of Figure 23.2), rare outcrops of mafic metavolcanic rocks (see Johnston 1969) may delineate a fault within the Minnitaki Group (i.e., volcanic rocks structurally truncated laterally to the west, along a fault). Considerable structural complexity in an area of limited exposure (sites 18 to 22 on Figure 23.2) is suggested.

22. Coarse intermediate to felsic pyroclastic rocks (tuff-breccia and rare autobreccia; probable dome facies) pass laterally westward, across a 3 km covered interval but with an interpreted facies change, to more distal subaqueous sedimentary facies: conglomerate, coarse and fine sandstone (including graded beds), and magnetite iron formation. The iron formation might have been sourced from exhalative vents in the volcanic dome to the east. A north-trending geophysical lineament beneath the covered interval may be a fault (Trowell et al. 1983), but if so it appears that the fault has not seriously disturbed the orderly east-trending stratigraphy. The possibility that the dome facies are located adjacent to a synvolcanic fault (a fault prone to later orogenic reactivation) merits investigation.

23. Top directions in moderately south-dipping pillow basalts of the Southern Volcanic belt are fairly consistently to the south (Trowell et al. 1983), toward the belt-marginal batholith. This does not support older views (e.g., Trowell and Johns 1986, p.51) that lower mafic volcanic sequences and/or units invariably face away from the large belt-marginal granitic batholiths in the western Wabigoon Subprovince.

24. North-younging top directions in pillow flows near the north margin of the Southern Volcanic belt (Pettijohn 1937, Figure 4; Johnston 1969) may be the result of hanging-wall anticlines (e.g., north-younging north limb, south-younging south limb) that formed via drag at the base of a thrust in a generally north-verging thrust belt.

25. Berger (1989) provided evidence of intrusion of the Basket Lake batholith into the Southern Volcanic belt, but the composite Basket Lake batholith is poorly known to the east and elsewhere (see Szewczyk and West 1976, Blackburn et al. 1991). Gneissic rocks of the Basket Lake batholith to the east may be older and part of a Central Wabigoon Region cratonic block (Blackburn et al. 1991) or Wabigoon Indenter (Stott and Corfu 1991). The presence of such a thick-skinned cratonic block pushing north to northwestward would account for the thin-skinned north-verging thrust belt style that appears to be preserved (despite subsequent sinistral strike-slip faulting) in the Sioux Lookout orogenic belt.
26. Farther to the southwest along strike, the Sandybeach Lake granitic pluton crops out (Berger 1989; Chorlton 1990, 1991; 2697 to 2693 Ma old: Davis 1990). Note that the supracrustal successions in the Sioux Lookout area have a paucity of granitic plutons, versus the more intruded areas adjacent to the east and west (see Blackburn et al. 1991).

27. Farther to the southwest along strike, the deformed, granitic Lateral Lake stock crops out (Page 1984; 2711 to 2705 Ma old: Davis 1990). The Lateral Lake stock appears to form the intrusive core of a local anticlinal structure.

Testing of Interpretations and Hypotheses

Further testing of the interpretive ideas proposed herein will require much detailed mapping of the faults, deformation zones, and pluton margins in the Sioux Lookout orogenic belt, including their form and orientation at depth. Small-scale kinematic indicators and other structures that formed during Stage 5 thrusting may be difficult or impossible to distinguish from those that formed during Stage 6 transpressional and/or transtensional faulting (e.g., early orogenic, syn-thrust horizontal lineations later rotated to vertical, or vertical lineations formed by late orogenic transpressional dip-slip movements?).

Suggested geochronological priorities for the Sioux Lookout orogenic belt are dating of the: 1) Northern Volcanic belt (basalt or any astatic interbeds; compare with other ages in the western Wabigoon Subprovince); 2) rhyolitic bodies in the Central Volcanic belt (possible ca. 2735 Ma bimodal basalt-rhyolite volcanism); 3) small, roughly sigmoidal granitic plutons within the Northern Volcanic belt (this may date the Stage 6 strike-slip faulting); and 4) Patara Formation sandstone beds (how young is the sand in this pull-apart basin?).

A REGIONAL RIEDELSHEAR LINEAMENT PATTERN OF GOLD OCCURRENCES?

Figure 23.3 shows gold occurrences in the Sioux Lookout orogenic belt and equivalent rock units to the southwest, plus a proposed lineament pattern connecting the occurrences. Major stratigraphic and fault surfaces trend at approximately 64 (Johnston 1969, 1972; Devaney et al. 1995b). Given the importance of late (Stage 6) sinistral strike-slip faulting outlined above, structural theory (see Figure 23.3 inset) allows that a conjugate Riedel shear fracture set (R and R' surfaces) could have formed. Based on this pattern, it is hypothesized that a regional-scale network of such R and R' fractures (Figure 23.3) formed a permeability system for the flow of gold-bearing fluids (i.e., present auriferous quartz veins; for Teck-Corona gold prospect and Goldlund Mine, see Page 1984, Chorlton 1991, Blackburn et al. 1998).

Although other lineament patterns are also possible (e.g., transpressional P shears), it is suggested that an orderly sinistral set of R and R' surfaces (faults, fractures, shears) might be expected as late transtensional tectonic features in such a region of sinistral strike-slip faulting. The lineaments of the proposed R and R' system are not mapped continuous faults, but may be preferred sites and orientations for discontinuously developed fracture, fault or shear systems. Note that planar features need not be in their ideal orientations; late faulting and progressive deformation may have rotated R, R' and other surfaces from their ideal theoretical orientations (e.g., Hampton and Neher 1986), and local anisotropies are likely to have influenced or deflected stresses.

This R-R' hypothesis (see Figure 23.3) is consistent with much of the detailed structural analysis of Chorlton (1990, 1991), whose study of the area immediately southwest of the present study area noted the importance of northeast- and east-northeast-trending veins and shear zones (some auriferous) and sinistral movements, amidst various structural complexities, in the southwest area's late stage tectonism and related gold mineralization.

If the concept of a regional-scale Riedel shear/fracture set pattern (see Figure 23.3) is correct, the pattern would have to be a very late, cross-cutting overprint on the previously complexly tectonized Sioux Lookout orogenic belt. This would conform with the common view that gold was typically deposited very late in the tectonic history of Superior Province greenstone belts (Colvini et al. 1988). Similarly late granitic plutons intruded into transtensional sites (e.g., releasing bends; see above) could have provided heat for hydrothermal vein systems.

Given that regional-scale Riedel and P-shear and/or P-fracture patterns in other orogenic belts have been interpreted as controlling factors in pluton emplacement (Petford and Atherton 1992, Tikoff and Teyssier 1992) and Archean gold mineralization (Mueller et al. 1988, Peterson 1997), the similar interpretation in Figure 23.3 might be mechanically reasonable. Although requiring much future testing, this new hypothesis could serve as an exploration model for the Sioux Lookout orogenic belt and other similar areas.

REFERENCES


24. Project Unit 98–001. Rare-metal Mineralization Associated with the Berens River–Sachigo Subprovincial Boundary, Northwestern Ontario: Discovery of a New Zone of Complex-Type, Petalite Subtype Pegmatite and Implications for Future Exploration

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INTRODUCTION

The distribution pattern of rare-metal mineralization in northwestern Ontario is generally characterized by linear belts that represent subprovince boundary zones (Breaks 1989; Breaks and Osmani 1989; Fyon et al. 1992, p.1118-1120). In particular, rare-metal pegmatites are especially abundant where one of the bounding subprovinces is characterized by a low to high grade metamorphic transition and contains abundant clastic metasedimentary and derived S-type, peraluminous granite masses. A good example is the Bird River–Separation Lake belt which is situated between the English River and Winnipeg River subprovinces of Ontario and Manitoba (Cerny et al. 1981; Breaks and Tindle 1997).

Other linear zones of peraluminous granite and associated rare-metal mineralization potentially exist in northwestern Ontario but have been examined only on a reconnaissance basis, for example, in the Sachigo subprovince (Ayres 1972a,b; Ayres et al. 1973; Stone et al. 1998). Therefore, to better evaluate the exploration potential for rare-metal mineralization in the Sachigo subprovince, this initial study was undertaken on one zone, which lies between the Sachigo and Berens River subprovinces (Figure 24.1). This zone is considered one of the most significant within northwestern Ontario, in terms of exploration potential, as five rare-metal mineral occurrences are known (Figure 24.2):

- Bearhead Lake holmquistite occurrence,
- Mattless Lake Zn-Be-Bi-Mo occurrence,
- Pennock Lake holmquistite occurrence,
- Pennock Lake spodumene occurrence, and,
- Pakeagama Li-Be-U-tourmaline occurrence.

Furthermore, the zone contains nine peraluminous granite masses, emplaced at 2697 ± 2 Ma (Corfu and Stone in press), which were delineated by Stone (1998: Map P.3382) over a strike length of 140 km, between Favourable Lake and McDowell Lake. Peraluminous granite and pegmatitic granite represent common parent magmas of rare-metal mineralization elsewhere in the Superior Province (e.g., Breaks and Moore 1992).

PREVIOUS GEOLOGICAL WORK

Rare-metal mineralization in the Sachigo subprovince was first documented by a 1:15 000 detailed mapping program in the Setting Net Lake region (Ayres 1970; 1972a; 1972b). Three rare-metal mineral occurrences were detected along the Bearhead Lake fault zone over a 16 km strike length, however, these are only shown on the 1:253 440 Bearhead–Favourable Lake geological compilation map (Ayres et al. 1973). Two of the occurrences contain holmquistite, whilst the third is represented by spodumene in white aplite near Pennock Lake with a bulk sample containing 0.52% Li (Ayres 1972b, p.12; Vos et al. 1987, p.219). Elsewhere, Stone et al. (1993a; 1998, p. 81) encountered anomalous Li (1063 to 5450 ppm), Cs (493 to 1410 ppm), Ta (48 to 242 ppm) and Be (47 to 164 ppm) values in three samples of tourmaline-rich rocks selected near Pakeagama Lake and these results are shown by a mineral occurrence symbol on the 1:50 000 Whetloon Lake sheet. However, the true significance of the mineralization (i.e., classification of the pegmatite type, its mineralogical and chemical nature and dimensions) remained unknown until the present study. It is now clear that the Pakeagama Lake occurrence represents the second largest complex-type, petalite subtype pegmatite in Ontario, being only surpassed by the Big Whopper pegmatite in the Separation Lake area (Breaks and Tindle 1997). This pegmatite system also contains the pollucite, the only ore mineral for cesium. This occurrence represents only the fourth occurrence in Ontario, and indicates a high exploration potential for zones rich in cesium.

PRESENT STUDY

Five weeks of field study were allotted to a reconnaissance examination of the northwest and southeast parts of the known rare-metal-mineralized zone along the Berens River–Sachigo subprovincial boundary zone, at Favourable Lake and Pakeagama Lake.
FAVOURABLE LAKE AREA

Several peraluminous granite masses were delineated in this area by Stone et al. (1993b; 1998; Map P.3382). A reconnaissance examination of pegmatitic exposures was undertaken coupled with mineral chemistry sampling that focused upon potassium-feldspar as a means of discerning potential fertile granite masses and pegmatite swarms (Cerny 1989, p. 288-292). Analytical work is currently in progress and will be reported at a later date.

Two relative ages of peraluminous granitic pegmatite were noted. The oldest is a swarm of garnet-biotite granitic pegmatites which are common along the southwestern shoreline of the middle arm of Favourable Lake across from the point of discharge into the Severn River. Here, pegmatite

Figure 24.1. Location map of study areas.
masses occur as deformed, commonly sheared boudins up to 10 m in thickness which are parallel to the 120° strike of the host-rock foliation. The most common varietal minerals are biotite and garnet; muscovite, tourmaline, molybdenite and black oxide minerals are rare.

A younger pegmatite dike was noted near the northwestern end of the middle arm of Favourable Lake (GPS: 52°53'47.6"N/93°57'42.7"W), where a 10 m thick, undeformed dike of biotite-garnet-muscovite pegmatitic granite was emplaced at a high angle to strike of the older pegmatite swarm. Muscovite is abundant in contrast to the older pegmatites and suggests a higher degree of pegmatite evolution.

Severn River Pegmatitic Granite Mass

This 0.5 to 2 by 10 km elongate mass of tourmaline-biotite-muscovite was mapped by Stone et al. (1993b; 1998, Map P. 3382) and several exposures can be examined along the Severn River north of Favourable Lake (Figure 24.2). Widespread black tourmaline in this pluton is a favourable indicator of possible fertile granite status, as boron is an important volatile in the complexing and transport of rare-metals and in the lowering of liquidus temperatures in pegmatitic melts (London 1986).

PAKEAGAMA LAKE AREA

Three weeks of detailed examination were directed towards the Pakeagama Lake pluton and its rare-metal pegmatite occurrence.

Geological Setting

The Pakeagama Lake rare-metal pegmatite is situated within the northwestern end of the Pakeagama Lake pluton (Figures 24.2 and 24.3), an elongate, 2 to 3 by 15 km mass of peraluminous granite first delineated by Stone et al. (1993a; 1998: [Map]. General distribution of peraluminous granite plutons and rare-metal mineralization along the Sachigo-Berens River subprovincial boundary zone. Rare-metal occurrences: 1. Bearhead Lake holmquistite; 2. Mattless Lake Zn-Be-Bi-Mo; 3. Pennock Lake spodumene; 4. Pennock Lake holmquistite; 5. Pakeagama Lake Li-Rb-Cs-Ta pegmatite. 1= Pakeagamo lake area; 2= Favourable Lake area.

Figure 24.2. General distribution of peraluminous granite plutons and rare-metal mineralization along the Sachigo-Berens River subprovincial boundary zone. Rare-metal occurrences: 1. Bearhead Lake holmquistite; 2. Mattless Lake Zn-Be-Bi-Mo; 3. Pennock Lake spodumene; 4. Pennock Lake holmquistite; 5. Pakeagama Lake Li-Rb-Cs-Ta pegmatite. 1= Pakeagamo lake area; 2= Favourable Lake area.
Map P.3382). Rare-metal pegmatite mineralization is mostly contained within this pluton, however, at its northwest end the main pegmatite mass is in contact with banded iron formation and metapelite that comprise part of an extensive metasedimentary unit which extends southeast into the North Spirit Lake greenstone belt. Furthermore, an exocontact zone of muscovite-tourmaline granitic pegmatite dikes is emplaced within mafic metavolcanic and metasedimentary rocks that lie adjacent to the northwestern terminus of the pluton and have been noted at least 1 km from the main rare-metal pegmatite mass.

**Pakeagama Lake Pluton**

This pluton is the likely parent granite for the enclosed rare-metal pegmatite. However, the pluton is not composed of pegmatitic granite units more typical of fertile granite masses that spawned rare-metal mineralization elsewhere in the Superior Province (e.g., Cerny et al. 1981; Breaks and Moore 1992; Breaks and Tindle 1997). The granitic rocks are generally weakly to modestly foliated and/or lineated, and medium to coarse-grained. Varietal minerals comprise biotite, muscovite and garnet, which collectively account for less that 10% of the mode. Muscovite is mostly present in only the northwestern part of the pluton; southeast of Pakeagama Lake, biotite is the main varietal mineral within the pluton. Textures are prevailingly granoblastic, resulting from shearing and recrystallization induced by deformation along the Bearhead Lake fault system, delineated by Stone et al. (1993a).

**Pakeagama Lake Rare-Metal Pegmatite**

A large, undeformed, steeply-dipping, strongly zoned pegmatite mass with a minimum strike length of 250 m was delineated by 1:250 detailed mapping (Figure 24.4). The pegmatite is open to the southeast and northwest. The width is variable from 10 to 30 m and it appears to widen to the northwest where there is a notable bifurcation marked by a smaller apophysis extending to the east.

![Figure 24.3. Detailed geology of Pakeagama Lake pluton and location of Pakeagama Lake rare-metal pegmatite.](image-url)
Figure 24.4. Detailed geology and sample locations of Pakegama Lake rare-metal pegmatite and immediate host-rocks.
Internal Units

Five internal units were delineated (Figure 24.4). All geochemical sample sites discussed below are also indicated on this map.

QUARTZ-RICH WALL ZONE

This 2 to 10 m-thick unit is restricted to the southeast part of the pegmatite and is characterized by 70 to 90% grey to white, locally vitreous or recrystallized quartz. The unit is in gradational contact with the potassium-feldspar-rich pegmatite zone described below. Subordinate phases include white to light brown stained, euhedral to subhedral montebrasite (a lithium aluminum phosphate), typically 2 to 3 cm in diameter or as irregular masses (Photo 24.1), blocky potassium-feldspar up to 27 cm diameter, recessively-weathered, deep blue fluorapatite up to 2 by 3 cm, deep green tourmaline (elbaite species) and sporadic, clear white to light brown beryl up to 2 by 4 cm. Local patches up to 1 m in length rich in mauve mica (possible lepidolite or lithium muscovite) occur sporadically in the wall zone. These zones are well foliated and contain blobs of quartz up to 8 by 14 cm and sparse, inconspicuous, black and brown oxide minerals, that in part are represented by cassiterite.

Potassium-Feldspar-Rich Pegmatite

This unit is dominated by grey-blue, generally blocky potassium-feldspar that typically exceeds 80% and is up to 1 m in diameter. The unit has been traced over a 90 m strike length in the southeast part of the pegmatite body and varies in width from 3 to 10 m. Euhedral and subhedral megacrysts of a light brown weathering, fine-grained aggregate of spodumene + quartz up to 0.7 by 1.1 m occur sporadically, and these are interpreted as pseudomorphs after petalite. Such megacrysts increase in abundance towards the spodumene-rich core zone and texturally envelop blocky potassium-feldspar, which indicates a later crystallization for the petalite. Subordinate phases in the potassium-feldspar-rich pegmatite include quartz and green, lithium-rich tourmaline that occur interstitial to the coarse blocky potassium-feldspar crystals.

SPODUMENE-QUARTZ-RICH CORE ZONE

This unit only occurs in the southeast part of the pegmatite and occupies a minimum area of about 10 by 15 m and is completely open to the southeast along strike. The rock consists of a generally fine, randomly-oriented intergrowth of 60% white to light pink spodumene and 40% lamellar to blob-like grey quartz.

Several large, blocky to triangular aggregates of spodumene-quartz are evident and up to 55 by 70 cm. Quartz also occurs as irregular, much coarser segregations up to 2 by 15 cm accompanied by minor montebrasite, green tourmaline and mauve mica. Blocky potassium-feldspar is rare in the core zone and up to 10 by 12 cm. Montebrasite (Photo 24.2) is disseminated throughout the core zone and generally evident as recessive and chalky-weathered euhedral crystals up to 6 by 8 cm, either enclosed by the spodumene-quartz intergrowth or as smaller crystals completely isolated in the irregular quartz-rich masses. The spodumene-quartz unit is overprinted by wispy veins, 1 to 5 mm in width, composed of fine-grained albite and quartz. Small patches of fine-grained lepidolite alteration has replaced parts of spodumene crystals. Black, fine-grained possible oxide phases are only sparsely evident.

PETALITE-POTASSIUM-FELDSPAR PEGMATITE

This unit may represent a variant of the potassium-feldspar-rich pegmatite unit as both grain size and colour of the potassium-feldspar megacrysts are similar. However, a marked increase in the amount of light brown, fine-grained, elongate, generally subhedral, spodumene-quartz aggregates after petalite is notable. These comprise 30 to 50% of the mode and exhibit maximum dimensions on outcrop surface of 1.2 by 5.2 m and envelop blocky potassium-feldspar crystals up to 34 by 61 cm.

LAYERED SPODUMENE PEGMATITE-APLITE

This unit dominates in the northwestern part of the pegmatite where it is associated with subordinate, much coarser
petalite-potassium-feldspar pegmatite. Typically there is intercalation of recessively-weathered, faint pink, spodumene-rich layers, 5 to 50 cm thick, with similar thicknesses of creamy white aplite. Larger pods rich in spodumene also occur in the zone that reveal the coarsest size for the mineral in the entire pegmatite: 2 by 5 by 10 cm.

**External Units**

Two types of pegmatite and aplite dike, generally less than 2 m thick, are common in the Pakeagama Lake pluton immediately southwest of the rare-metal pegmatite mass (Figure 24.4). Here such dikes strike parallel to the pegmatite body and most commonly comprise aplite with dark green lithian tourmaline up to 1 cm diameter, muscovite and rare dark brown liothite (LiMnPO₄). Pods with abundant, randomly oriented white spodumene occur locally in some of the aplite dikes. The aplite dikes are characterized by a core rich in black tourmaline enveloped by bleached margins rich in white feldspar (probably albite) typically less than 5 cm thick, which can locally be traced into larger quartz-rich masses.

The most important exocontact unit is sparse dikes of potassic pegmatite that were emplaced at high angle (070° strike) to the 125° strike of the aplite dikes. Especially noteworthy is the presence of pollucite [(Cs,Na)₂ Al₂O₅·H₂O], as several recessively-weathered crystals up to 2 by 4 cm (Photo 24.3) in part of a 30 cm thick dike of potassium-feldspar-green tourmaline-quartz-albite which is situated 120 m south of the main rare-metal pegmatite mass (Figure 24.4). This mineral was verified by X-Ray diffraction and represents only the fourth occurrence in Ontario.

Pegmatite dikes also occur in the supracrustal rocks that enclose the northwestern end of the pluton. In mafic metavolcanic rocks situated 500 m from the pluton, folded and sheared dikes of garnet-quartz-potassium-feldspar-tourmaline-muscovite-albite form boudins up to 1 by 3 m. Anomalously high levels of Rb and Cs in blocky potassium-feldspar (Table 24.1 and Figure 24.5) support a genetic linkage with the nearby Pakeagama Lake rare-metal pegmatite.

**Host-Rock Metasomatism**

Granitic rocks of the Pakeagama Lake pluton have undergone conspicuous alteration proximal to the rare-metal pegmatite and three types of metasomatic transformation have been recognized:

1. Li-rich tourmaline metasomatism
2. black tourmaline veins with bleached margins rich in white feldspar
3. holmquistite porphyroblasts.

Type 1 metasomatism occurs within one meter of the pegmatite, and involves the granitic host-rocks, either in contact with the quartz-rich wall zone or the potassium-feldspar-rock pegmatite zone. Type 1 is a conspicuous development of 5 to 10% coarse, dark green tourmaline porphyroblasts in the host-rocks (Photo 24.4) within diffuse, bleached splotches and veins rich in white feldspar and devoid of muscovite and biotite that otherwise comprise 10% of unaltered granite. Colour of the tourmaline porphyroblasts is dark green within 30 cm of contact with the pegmatite, however, the tourmaline changes to black at greater distance. One bulk analysis reveals 0.61% Li₂O and 2653 ppm Rb in the green tourmaline-rich granite. Electron microprobe work

**Photo 24.2.** Euhedral crystal of montebasite in spodumene-quartz core zone. Coin diameter = 2.5 cm.

**Photo 24.3.** Several, recessively-weathered crystals of pollucite (arrows) in an exocontact pegmatite dike situated 120 meters south of the Pakeagama Lake Pegmatite.
indicates a compositional range of elbaite to schoen with minor liddicoatite compositions (these are respectively Li-, Fe- and Ca-Li-rich end-member tourmaline compositions).

Type 2 metasomatism consists of veins rendered conspicuous by the presence of a thin core rich in black tourmaline that is enveloped by thicker margin zones rich in white feldspar. These veins are most common within 20 m of the pegmatite mass but have been observed up to 500 m to the southwest along the shoreline of Pakeagama Lake. Thickness of the veins is typically 3 to 5 cm but may reach up to 10 cm.

Type 3 metasomatism is characterized by subtle, dark purple, elongate fibrous aggregates of holmquistite (a lithium amphibole), generally 5 mm in length, which occurs southwest of the Pakeagama Lake pegmatite (Figure 24.4). Occurrences of the mineral were traced for a minimum distance of 120 m from the pegmatite and coincides with a significant bulk rock lithium anomaly, marked by Li values between 295 and 843 ppm. The widespread distribution of holmquistite is quite unusual as the mineral normally favours mafic host-rocks within 20 m of a rare-metal pegmatite body (Cerny et al. 1981, p. 101). Holmquistite is rarely developed in granitic host-rocks such as the Pakeagama Lake pluton and its extensive occurrence in these granites, over at least a 120 m distance normal to the rare-metal pegmatite contact has, to the authors' knowledge, never been previously described. We therefore have interpreted that a large, lithium-rich, blind pegmatite mass occurs at shallow depth within the anomalous Li area.

Subsolidus Alteration in the Pakeagama Lake Pegmatite

Minor subsolidus alteration has affected several minerals in the Pakeagama Lake pegmatite. Narrow white veins, normally less that 5 mm thickness and rich in fine-grained, sugary albite and local green tourmaline, blue apatite and quartz subtly cross-cut blocky potassium-feldspar and the light brown spodumene-quartz intergrowths in the petalite-potassium-feldspar zone. Such veins locally contain zoned pods composed of a white fluorapatite core and fibrous montebrasite margins. Albite-rich veins also pervade the spodumene-quartz core zone and are normally devoid of accompanying phases except for sporadic black specks of possible oxide minerals. Fine-grained lepidolite also occurs in this zone and replaces small areas of spodumene.

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**Figure 24.5.** K/Rb versus cesium in potassium feldspar from Pakeagama Lake rare-metal pegmatite and adjacent exocontact pegmatite dikes. Bernic Lake pegmatite group after Cerny et al. (1981); Separation Rapids pegmatite group and Separation Rapids pluton from Breaks and Tindle (1997).
MINERAL CHEMISTRY

Potassium-feldspar chemistry has been established as a useful geochemical tool in the economic evaluation of pegmatites (Cerny 1989, p.288-291 and also references therein). In particular, the Rb and Cs contents and the K/Rb ratio in blocky potassium-feldspar are important in the discrimination between rare-metal-enriched pegmatites and those that are barren of such metals. Such data are typically examined in the K/Rb versus Cs plot devised by Cerny et al. (1981) whose extensive work in the Winnipeg River pegmatite district established boundaries between various geochemical pegmatite types (i.e., Be, Li-Be and Li, Cs, Be, Ta types). Furthermore, their work provided fields of K/Rb versus Cs variation for fertile parent granites that spawned various pegmatite groups in this district and hence facilitates geochemical recognition of potential fertile granites elsewhere in the Superior Province (e.g., Separation Rapids pluton: Breaks and Tindle 1997, p.17).

Other useful minerals that can be utilized in the recognition and economic evaluation of rare-metal pegmatites include muscovite and beryl (Cerny 1989).

Potassium-Feldspar

Twenty-one blocky potassium-feldspar samples were analyzed from the Pakegama Lake pegmatite system and data received to date are presented in Table 24.1. Concentrations of rubidium (mean=1.11 %; range = 0.97 to 1.2 %) and cesium (range from 400 to 1349 ppm) are highly anomalous and indicative of a complex type pegmatite system (Cerny 1989, p. 292). This assessment is supported by the K/Rb versus Cs diagram (Figure 24.5), a plot useful in assessing economic potential (Cerny et al. 1981), where most analyses from the Pakegama Lake area (84 %) plot within the Bernice Lake pegmatite group of Cerny et al. (1981), which includes the Tanco Pegmatite. About one-third of the Cs values in potassium-feldspar exceed the 900 ppm upper limit for XRF analysis in the Geoscience Laboratory (currently being reanalyzed) and such levels of Cs generally suggest presence of polliclite, which was verified in part of the pegmatite system.

Two potassium-feldspar samples (Table 24.1: 98KF-26 and 98KF-27) from exocontact dikes, situated to the northwest of the Pakegama Lake pegmatite, also indicate an advanced evolution by virtue of anomalous cesium (155 to 247 ppm) and Rb concentrations (6400 to 9100 ppm). Such values compare with fertile granite units elsewhere in the Superior Province as exemplified by the Separation Rapids pluton (Breaks and Tindle 1997) and the Ghost Lake batholith (Breaks and Moore 1992).

Petalite

Textural and chemical evidence indicates that most of the spodumene + quartz intergrowths are likely pseudomorphs after petalite. No primary petalite was observed within any of the pegmatite zones which contrasts with the Big Whopper Pegmatite where unaltered petalite is pervasive (Breaks and Tindle 1997). Analyses of intergrowths taken from several sites of the Pakegama Lake pegmatite reveal a similarity of its Li2O content (Table 24.2: mean=4.42 %; range = 3.77 to 4.74 %) to petalite analyses in the literature (Table 24.3; London and Burt 1982, p.104; Cerny et al. 1981, p.106). In contrast, the Li2O contents of spodumene are considerably higher (6.60 to 7.87 %; London and Burt 1982, p.105). Furthermore, the successive channel samples taken across one-half of the width of the spodumene-quartz core unit, have revealed similar Li2O levels (Table 24.2: mean= 4.52%; range = 4.33 to 4.79%), which also implies that most of this unit was initially composed of petalite prior to subsolidus breakdown to the present intergrowth (London 1984).

Oxide Minerals

Several economically important, tantalum-rich minerals (e.g., Photo 24.5) have been documented in 70 electron microprobe analyses undertaken to date at the Department of Earth Sciences of the Open University. These include manganotantalite, ferrotantalite, manganocolumbite, micro-lite (including antimony- and uranium-rich species) and ferrotalpiolite. Dark brown cassiterite contains up to 5% Ta2O5. Representative analyses of the various oxide phases are presented in Table 24.4. The oxide minerals are typically fine grained and sparsely evident in the zones observed to date.

Photo 24.4. Dark green, lithium tourmaline porphyroblasts in granite host-rocks adjacent to potassium-feldspar-rich zone. Coin diameter = 2.5 cm.
Table 24.1. Ba, Cs, Li, Rb and Sr content (ppm) in blocky potassium-feldspar from the Pakeagama Lake rare-metal pegmatite and adjacent excontact dikes.

<table>
<thead>
<tr>
<th>Sample #</th>
<th>Rb</th>
<th>Cs</th>
<th>Ba</th>
<th>Sr</th>
<th>Li</th>
<th>K/Rb</th>
<th>K/Cs</th>
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<td>130</td>
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</tbody>
</table>

*ND = not detected*

Important, however, from an exploration perspective, is the trend of compositional variation in the columbite-tantalite quadrilateral (Figure 24.6). A well-defined trend of chemical evolution, expressing a change from manganocolumbite into manganotantalite is obvious. This trend is similar to columbite-tantalite oxide populations, for example, from the Peerless Pegmatite, part of the Tanco Pegmatite (Ericit and Cerny 1985) and the eastern pegmatite subgroup of Separation Rapids group (Tindle and Breaks 1998, p. 613). Such evolutionary trends indicate a high potential for tantalum-rich zones in the pegmatite system.

A subordinate trend of manganocolumbite to manganotantalite at lower Mn/Fe ratios is also notable and expresses a very similar range of Ta/Nb ratios to that of the main trend. A narrow aplite vein situated 400 m southeast from the Pakeagama Lake pegmatite contains sparse ferrotiaplite and ferrotantalite grains associated with tourmaline and blue apatite. These oxide minerals have not been found to date in the main pegmatite mass. Nevertheless, the occurrence of this aplite unit with such evolved tantalum-rich oxide minerals, suggests a continuation of the rare-metal pegmatite system at least 400 m to the southeast beyond the outcrop containing the spodumene-quartz core unit. The aplite and tourmaline-metasomatized granite host-rocks are also anomalous in lithium, respectively at 120 ppm and 1068 ppm, which supports this contention.

An absence of relatively primitive ferrocolumbite compositions in the Pakeagama Lake pegmatite is also notable and supports the high exploration potential for tantalum. Ferrocolumbite was only documented in narrow, albite-rich dikes that occur in mafic metavolcanic rocks, 2 km from the pegmatite body.

**SUMMARY**

This section summarizes salient features in regards to the geology, mineralogy and geochemistry of the Pakeagama Lake rare-metal pegmatite:

1. Surface dimensions: minimum 250 m strike length; width 10 to 30 m.
2. Five internal zones: a) Quartz-rich wall zone (quartz >> montebrasite-potassium-feldspar-muscovite-green tourmaline); b) potassium-feldspar-rich pegmatite zone (potassium-feldspar>>petalite-quartz-green tourmaline); c) Spodumene-quartz-rich core zone (spodumene >> quartz-montebrasite-green tourmaline); d) Petalite-potassium-feldspar pegmatite (potassium-feldspar-petalite-quartz); Layered spodumene pegmatite-aplite zone (spodumene-albite-quartz).
3. External units: a) Green tourmaline-muscovite-quartz-albite aplite dike; b) potassic pegmatite (potassium-feldspar-green tourmaline-quartz-albite with local pollucite).
4. Metasomatic zones in granite host-rocks: a) lithium-rich green tourmaline metasomatism; b) veins rich in white feldspar with black tourmaline cores; c) holmquistite porphyroblastosis.
<table>
<thead>
<tr>
<th></th>
<th>Megacrysts of Fine-Grained Spodumene-Quartz Intergrowth</th>
<th>Coarse Spodumene-Quartz Assemblage from Core Zone</th>
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<tbody>
<tr>
<td></td>
<td>98FWB-29A</td>
<td>98FWB-29B</td>
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<td>78.43</td>
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<tr>
<td>Al₂O₃</td>
<td>17.22</td>
<td>16.72</td>
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<tr>
<td>MnO</td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>MgO</td>
<td>0.06</td>
<td>0.05</td>
</tr>
<tr>
<td>CaO</td>
<td>0.05</td>
<td>0.03</td>
</tr>
<tr>
<td>Na₂O</td>
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<td>0.61</td>
</tr>
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<td>0.15</td>
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<td>Li₂O</td>
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<td>4.65</td>
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<td>P₂O₅</td>
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<td>0.01</td>
</tr>
<tr>
<td>TiO₂</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>FeO</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>0.01</td>
<td>0.03</td>
</tr>
<tr>
<td>CO₂</td>
<td>0.13</td>
<td>0.10</td>
</tr>
<tr>
<td>H₂O+</td>
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</tr>
<tr>
<td>H₂O⁻</td>
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<td>0.08</td>
</tr>
<tr>
<td>TOTAL</td>
<td>101.4</td>
<td>100.92</td>
</tr>
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</table>

Samples from following zones: potassium-feldspar-pegmatite zone: 98FWB-29A, 98FWB-29B; potassium-feldspar-rich zone: 98KF-5 and 98KF-17; Spodumene-quartz core zone as channel samples across part of zone: 98FWB-39 (0-1 m), 98FWB-40 (1-2 m), 98FWB-41 (2-3 m), 98FWB-42 (3-4 m) and 98FWB-43 (4-4.9 m).
Table 24.3. Major element analyses of petalite and spodumene-quartz pseudomorphs after petalite from the literature and unpublished analyses of the authors.

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<td>76.79</td>
<td>77.30</td>
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<td>16.60</td>
<td>16.85</td>
<td>17.68</td>
<td>18.1</td>
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<td>0.03</td>
<td>ND</td>
<td>ND</td>
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<td>0.03</td>
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<td>ND</td>
<td>ND</td>
<td></td>
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<tr>
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<td>0.08</td>
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<tr>
<td>Na₂O</td>
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<td>4.61</td>
<td>4.22</td>
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<tr>
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<td>ND</td>
<td>0.01</td>
<td>0.04</td>
<td></td>
<td></td>
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<tr>
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<td>0.16</td>
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<td>0.09</td>
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<td>0.66</td>
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</tr>
<tr>
<td>H₂O⁻</td>
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<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>TOTAL</td>
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<td>99.87</td>
<td>99.08</td>
<td>100.30</td>
<td>100.4</td>
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</tbody>
</table>

ND = not detected

1. Big Whopper Pegmatite, Separation Rapids pegmatite group. Unpublished analysis of light pink petalite of the authors.
2. Tanco Pegmatite, grey petalite from Tanco zone 5 (Cerny and Ferguson 1972)
5. MNW Pegmatite, Georgia Lake pegmatite field, near Nipigon, Ontario. Unpublished analysis of the authors.

Potassium-feldspar chemistry: a comparison of the mean values and ranges of Cs, Rb and K/Rb in the Pakeagama Lake pegmatite with data from the Bernic Lake pegmatite group in Manitoba and the Big Whopper Pegmatite in the Separation Rapids group is given in Table 24.5. It can be seen that the Pakeagama Lake Pegmatite has a comparable K/Rb ratio but has higher Cs values than the Big Whopper pegmatite and overlaps the lower ranges for Rb, Cs and K/Rb with the Bernic Lake pegmatite group.

COMMODITY TYPES MINED FROM RARE-METAL PEGMATITES

Rare-metal pegmatites of the complex-type are unique sources of several important commodities. These are listed in Table 24.6 along with current prices where available.

FUTURE MINERAL EXPLORATION

Rare-metal mineralization along Berens River–Sachigo subprovincial boundary is now known over a 65 km strike length, between Bearhead and Pakeagama lakes. Peraluminous granites, the plausible progenitor of the rare-metal mineralization, occur over at least a 140 km strike length, (from Favourable to MacDowell lakes), and hence there is, accordingly, a high potential for further such discoveries of complex-type, petalite subtype pegmatites. Petalite-bearing pegmatites represent the most important exploration target for ceramic/glass grade petalite and/or spodumene, tantalum, rubidium and cesium. The discovery at Pakeagama Lake, is the direct result of previous OGS work (Stone et al 1993a; Stone 1998) in the region, with follow-up represented by this project. This discovery may well be significant and warrants follow-up work by the private sector. Recommended priorities for exploration area:
Table 24.4. Representative electron microprobe analyses of various oxide minerals from the Pakeagama Lake area. Operating conditions for the electron microprobe are given in Tindle et al. (1998).

<table>
<thead>
<tr>
<th></th>
<th>98SS-11</th>
<th>98-34B-i</th>
<th>98AT-2</th>
<th>98AT-3i</th>
<th>98AT-3ii</th>
<th>98FWB-36i</th>
<th>98-34B-ii</th>
<th>98-34iii</th>
<th>98FWB-36ii</th>
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<td>0.97</td>
<td>11.59</td>
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<td>0.36</td>
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<td>MnO</td>
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<td>13.37</td>
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<tr>
<td>TiO₂</td>
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<td>0.70</td>
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<td>0.18</td>
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<tr>
<td>Nb₂O₅</td>
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<td>4.71</td>
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<tr>
<td>Ta₂O₅</td>
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<tr>
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<tr>
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<td>99.17</td>
<td>98.52</td>
<td>99.74</td>
<td>100.86</td>
<td>100.18</td>
</tr>
</tbody>
</table>

98SS-11: Ferrocolumbite from albite-rich dyke, 500 m southwest of Pakeagama Lake pegmatite.
98-34B-i: Ferrotantalite from tourmaline-apatite aplite dyke, 400 m southeast of Pakeagama Lake pegmatite.
98AT-2: Manganocolumbite from albite-rich veinlet that cross-cuts blocky potassium-feldspar. From potassium-feldspar sample site 98KF-4 (Figure 24.3) in Pakeagama Lake Pegmatite.
98AT-3i: Manganocolumbite from patchy zoned grain (Photo 24.x) in muscovite-green elbaite-bearing aplite dike situated 3 m from southwest contact of Pakeagama Lake Pegmatite.
98AT-3ii: Manganotantalite from same oxide grain as analyzed in 98AT-3i.
98FWB-36i: Manganotantalite from 3 m-thick layered aplite dyke situated 15 meters southwest of Pakeagama Lake Pegmatite.
98-34B-ii: Ferrocolumbite from same location as 98-34i.
98-34Biii: Microlite from same location as 98-34i.
98FWB-36ii: Cassiterite from same location as 98FW
Table 24.5. Comparison of mean values and ranges for Cs, Rb and K/Rb in K-feldspar for Pakeagama Lake pegmatite with Bernic Lake pegmatite group and Big Whopper Pegmatite of Separation Rapids pegmatite group.

<table>
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<th>Rb (ppm)</th>
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<th>K/Rb</th>
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</tr>
</thead>
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<tr>
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<td>Range</td>
<td>Mean</td>
<td>Range</td>
<td>Mean</td>
<td>Range</td>
</tr>
<tr>
<td>Pakeagama Lake</td>
<td>827</td>
<td>415 to 1349</td>
<td>11 100</td>
<td>9700 to 11 100</td>
<td>9.5</td>
<td>9.3 to 11.0</td>
</tr>
<tr>
<td>pegmatite</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bernic Lake</td>
<td>1262</td>
<td>165 to 2866</td>
<td>14 800</td>
<td>3880 to 26 600</td>
<td>9.5</td>
<td>4.0 to 24.3</td>
</tr>
<tr>
<td>pegmatite group</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Big Whopper</td>
<td>249</td>
<td>54 to 506</td>
<td>12 592</td>
<td>9654 to 15 000</td>
<td>8.5</td>
<td>7 to 11</td>
</tr>
<tr>
<td>pegmatite</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
| N = number of analyses

1. Surface stripping to better reveal various pegmatite zones and to help extend southeasterly and northwestern limits of pegmatite system,

2. Trenching and continuous chip sampling across the various internal pegmatite zones to establish representative grades for Li₂O, Cs₂O, Rb₂O and Ta₂O₅.

3. Lithochemical sampling of granite host-rocks to establish limits of the large lithium anomaly. Follow-up with Cs and Rb bulk analysis to help establish drill targets for search for blind pegmatites, and,

4. Prospecting and regional mineral (potassium-feldspar) and lithochemical sampling in the supracrustal units to the northwest of the Pakeagama Lake pegmatite in search of other rare-metal pegmatite systems in the area.

Our results reinforce the mineral exploration advice given by Breaks and Osmani (1989): “Favourable targets for rare-metal pegmatites in northwestern Ontario include the Bearhead, Pounak Lake and North Kenyon fault systems, and some subprovince boundary zones, such as the area along the English River–Uchi boundary between Attwood and Pakwash lakes”.

Ayres (1972) indicated that the Pennock Lake spodumene occurrence (0.52 % lithium) represents an aplite zone within a white pegmatite dike that is common in a 500 m wide zone on the north side of the Bearhead Lake fault system. Furthermore, Ayres states that “since none of the localities were identified in the field, the distribution of lithium-bearing minerals may be much more extensive than that documented by the three samples” [i.e., lab samples selected by Ayres from the Bearhead Lake holmquistite, Pennock Lake spodumene, and Pennock Lake holmquistite occurrences].

In the Pakeagama Lake pegmatite, the presence of tantalum-rich minerals such as manganotantalite, ferrotantalite, microlite and ferrotiolite coupled with pollucite and rubidium concentrations that exceed 1% in potassium-feldspar, indicate a high potential for zones rich in tantalum, cesium and rubidium-rich potassium-feldspar. Primary petalite has not been observed to date, and always is altered to SQUI (acronym for spodumene+quartz intergrowth). Montebraisite, which occurs abundantly in some zones of the Pakeagama Lake pegmatite, is an additional commodity to consider in exploration of the pegmatite. There is currently a limited market for this high lithium mineral (9.30 to 10.22% Li₂O; London and Burt, 1982, p.114), which has application to production of high phosphate frits and ceramic glazes (Cerny et al. 1996, p.50).

The area immediately northwest of the Pakeagama Lake pluton and that extending to the Bearhead Lake area is recommended as one in which additional discoveries of rare-metal mineralization may occur, as evolved tourmaline pegmatites, described above, have been documented by this survey. Such pegmatites are potentially likely to be hosted in mafic metavolcanic and metasedimentary units adjacent to terminations of peraluminous granite plutons, as, for example, adjacent to the northwestern end of the Pakeagama Lake pluton.
Table 24.6. Current prices for various commodities from rare-metal pegmatite deposits

<table>
<thead>
<tr>
<th>Commodity</th>
<th>% rare metal</th>
<th>Price (US $)</th>
<th>Price/kg or lb of rare metal</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spodumene concentrate</td>
<td>7.5% Li₂O</td>
<td>$1763-2204/tonne</td>
<td>Current market price</td>
<td>1</td>
</tr>
<tr>
<td>Glass grade spodumene</td>
<td>4.8% Li₂O</td>
<td>$363-385/tonne</td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>Petalite</td>
<td>4.3% Li₂O</td>
<td>$195-260/tonne</td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>Montebrasite</td>
<td>7% Li₂O, 8% P₂O₅</td>
<td>Price unknown</td>
<td></td>
<td>2</td>
</tr>
<tr>
<td>Tantalum concentrate and</td>
<td>35-38% Ta₂O₅;</td>
<td>Price in recent past:</td>
<td></td>
<td>3</td>
</tr>
<tr>
<td>refined metal product</td>
<td>14-18% SmO₂;</td>
<td>$40/pound for concentrate</td>
<td>per pound of Ta₂O₅ powder: US</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>5 to 8% Nb₂O₅</td>
<td>with 40% Ta₂O₅</td>
<td>$40-50</td>
<td></td>
</tr>
<tr>
<td>Niobium concentrate and</td>
<td>Concentrate with</td>
<td>$2.80 to 3.20/pound in 1996;</td>
<td>$8.17 per pound of</td>
<td>5</td>
</tr>
<tr>
<td>refined metal product</td>
<td>65% Nb₂O₅</td>
<td>alloy: $18-50/tonne of Nb in 1996</td>
<td>Nb₂O₅ oxide powder;</td>
<td></td>
</tr>
<tr>
<td>Pollucite (crushed ore)</td>
<td>24% CaO, 1.5% Rb₂O, 0.75% Li₂O</td>
<td>Up to $1000/tonne in past</td>
<td>high purity ferrocolumbium</td>
<td>2, 4</td>
</tr>
<tr>
<td>Rb-rich K-feldspar</td>
<td>Price unknown</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lepidolite</td>
<td>3% Rb₂O</td>
<td>Price unknown</td>
<td></td>
<td>2</td>
</tr>
</tbody>
</table>


REFERENCES


Stone D., Crawford, J. and Halstead, J. 1993c Precambrian geology, Kember Lake area; Ontario Geological Survey, Preliminary Map P3220, scale: 1:50 000.


25. Project Unit 98–001. Time-scales of Formation of Rare-metal Class Pegmatites and Associated Peraluminous Granite in the Superior Province of Ontario

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<sup>2</sup>Ontario Geological Survey, Precambrian Geoscience Section

INTRODUCTION

As part of a project agreement between the Ontario Geological Survey and The Open University, rare-metal class pegmatites and associated granites and pegmatitic granites were sampled in order to constrain pegmatite genetic processes with age data. Samples were obtained from most subprovinces in the Superior Province (Figure 25.1). There is currently a dearth of meaningful geochronological data on such rocks (e.g., Larbi et al. in press), hence, the initial goals of the research are:

1. date the time of emplacement of rare-metal pegmatites and associated peraluminous parent granites (where exposed) as a function of subprovince setting across the Superior Province; and,

2. to define the thermal history within well-defined zoned pegmatite swarms.

The characteristic internal zoning of some pegmatites has been the subject of much debate (e.g., Cerny 1992). As a specific example, the recently discovered Big Whopper Pegmatite (Breaks and Tindle 1997), situated within the Bird River–Separation Lake metavolcanic belt, exhibits thin albite-rich border zones that pass into a pervasive petalite-potassium-feldspar-rich core zone. Due to the large size of some rare-metal pegmatite bodies, as exemplified by the Big Whopper Pegmatite which is up to 60 m wide, it is plausible that such masses remained as a liquid melt for millions of years, especially within the insulated core zones. Pegmatites are suspected to form a continuum of magmatic, supercrustal and hydrothermal regimes (Cerny 1992). By dating the early-crystalized wall zones and the later core units, it may be possible to determine the time interval that the pegmatite dikes remained active with respect to these regimes.

Dating of rare-metal pegmatites from different subprovinces will allow a comparison of the age of their formation across the Superior Province. To date, pegmatites have been sampled from the Abitibi, Quetico, Wabigoon, English River subprovinces and also from the following subprovincial boundary zones: Berens River–Sachigo, English River–Winnipeg River (Bird River–Separation Lake metavolcanic belt), and Winnipeg River–Wa-
bigoon. Details of the localities of samples are given in Table 25.1.

Cerny (1990) stated that fertile granites and pegmatites are, in most cases, late- to post-tectonic and conclude the granitic magmatic activity for most Archean shield areas. Thus it will be of interest to determine whether the southward-younging evolution of the Superior Province as proposed by Scott et al. (1989) is mirrored by the intrusion ages of rare-metal class pegmatites and their associated parent granites. Geochronological studies undertaken in the Baltic Shield (Smeds 1994; Romer and Smeds 1997) have obtained ages for LCT (lithium-cesium-tantalum-) type rare-metal pegmatites (LCT = lithium-cesium-tantalum: Cerny 1992). Such pegmatites were argued by these authors to have originated from crustal melts derived from orogenically-thickened continental crust and therefore yield a minimum estimate for the age of orogenic thickening. Thus it should be possible to show how differences in the ages of pegmatites related to differences in the timing of crustal growth in the various subprovinces of the Superior Province. Reliable ages of rare-metal pegmatite emplacement thus far have revealed a late Archean development which, in the Separation Rapids pegmatite group, is 2.45 Ma with a late thermal event at 2.35 (Larbi et al. in press) which significantly post-dates the 2.7 Ma Kenoran orogenic event.

SAMPLE SELECTION PROCEDURE

The ⁴⁰Ar/³⁹Ar isotope laser dating technique will be used to obtain absolute ages for the pegmatites. This dating technique is similar to the K-Ar method and is dependent upon the radioactive decay of ⁴⁰K to ⁴⁰Ar. Thus it is necessary to collect samples that contain minerals with a high potassium content and muscovite is the obvious choice as it contains high levels of potassium coupled with a widespread distribution in rare-metal pegmatites and associated peraluminous granites. In all cases, primary muscovite samples were stringently collected which also had to be undeformed and exhibited well-defined crystal boundaries with adjacent phases. This protocol was adopted so as to avoid analysis of micas that may have been influenced by post-emplacement events or by weathering forces.
Figure 25.1. Location map of study areas.
Table 25.1. Location and classification of pegmatite types/subtypes sampled in Superior Province of Ontario.

<table>
<thead>
<tr>
<th>PEGMATITE NAME or AREA and MAP REFERENCE #</th>
<th>SUBPROVINCE</th>
<th>PERALUMINOUS PARENT GRANITE</th>
<th>RARE-METAL PEGMATITE TYPE</th>
<th>SUBTYPE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Case Pegmatite, Steel Township</td>
<td>Abitibi</td>
<td>Unknown</td>
<td>Albite-spodumene type possibly transitional into complex-type</td>
<td>Albite-spodumene type</td>
</tr>
<tr>
<td>2. Lowther Township Pegmatite</td>
<td>Quetico</td>
<td>Unknown</td>
<td>Albite-spodumene type</td>
<td></td>
</tr>
<tr>
<td>3. Raleigh Lake pegmatite group, Ignace area</td>
<td>Wabigoon</td>
<td>Unknown</td>
<td>Albite-spodumene type</td>
<td></td>
</tr>
<tr>
<td>4. Mavis Lake and Gullwing-Tot lakes pegmatite groups, Dryden area.</td>
<td>Wabigoon Subprovince near boundary with Winnipeg River Subprovince</td>
<td>Ghost Lake batholith</td>
<td>a. Albite-spodumene type (Mavis Lake group)</td>
<td>b. Spodumene subtype present at Tot Lake complex type pegmatite only.</td>
</tr>
<tr>
<td>5. Tustin Township</td>
<td>Wabigoon-Winnipeg River subprovincial boundary zone</td>
<td>Medicine Lake pegmatitic granite stock</td>
<td>Beryl type pegmatite pods</td>
<td></td>
</tr>
<tr>
<td>6. Graphic Lake area, Work Township</td>
<td>Wabigoon</td>
<td>Unknown</td>
<td>Beryl type</td>
<td></td>
</tr>
<tr>
<td>7. Separation Lake area</td>
<td>Bird River-Separation Lake greenstone belt in English River-Winnipeg River subprovince boundary zone</td>
<td>Separation Rapids pluton</td>
<td>Beryl type and complex type</td>
<td>Petalite subtype of complex type pegmatites are widespread.</td>
</tr>
<tr>
<td>8. Sandy Creek Pegmatite, north of Ear Falls</td>
<td>English River Subprovince near boundary with Uchi Subprovince</td>
<td>Unknown</td>
<td>Beryl type</td>
<td>Phosphate subtype</td>
</tr>
<tr>
<td>9. Pakeagama Lake Pegmatite</td>
<td>Sachigo-Berens River subprovincial boundary zone</td>
<td>Pakeagama Lake pluton</td>
<td>Complex type</td>
<td>Petalite subtype</td>
</tr>
<tr>
<td>10. Favourable Lake area</td>
<td>Sachigo Subprovince</td>
<td>Severn River pluton potential parent to rare-metal pegmatites</td>
<td>Holmquistite and spodumene occur in Bearhead Lake area but as yet examined in detail.</td>
<td></td>
</tr>
</tbody>
</table>

The main reason for undertaking Ar-Ar laser dating of the pegmatites is that precise U/Pb zircon dating is difficult to obtain as peraluminous pegmatites are notoriously lacking in zircon. Whole rock Rb/Sr dating is affected by the very high Rb/Sr ratios in rare-metal pegmatites and can be reset by tectonic disturbances. The Ar-Ar laser technique has a unique ability to analyse areas less than 1 mm across which enables one to ascertain the thermal history of a single mineral grain.

**ANALYTICAL METHODS**

All pegmatite samples will be examined on the Cameca SX100 electron microprobe facility at The Open University. The data from these probe sections will be used to determine if metamorphism has affected the pegmatites, to identify possible alteration and provide a chemical explanation for any variations in age within a single sample.

The \(^{40}\text{Ar}/^{39}\text{Ar}\) method relies upon the reaction during neutron irradiation that produces \(^{39}\text{Ar}\) from \(^{39}\text{K}\). Dates obtained from measured \(^{40}\text{Ar}/^{39}\text{Ar}\) ratios have been shown to be similar to conventional K-Ar dates, but have the advantage of being based upon Ar isotopic ratios only and without the need for a further measurement of potassium concentration (Faure 1986). Ideally, step-heating of mineral samples that have undergone a partial loss of radiogenic argon will produce a series of dates. These dates with be a \(^{40}\text{Ar}/^{39}\text{Ar}\) plateau at high temperatures and will be equal to the time of intrusion and initial cooling. The second peak of \(^{40}\text{Ar}/^{39}\text{Ar}\) at lower temperatures may correspond to a younger metamorphic event.

For the initial work on the muscovite separates, the most pristine, transparent and undeformed grains have been selected and priority has been given to the following pegmatites and parent granites: Big Whopper Pegmatite, Big Mack Pegmatite (Emerald Fields Resources), Marko’s Pegmatite, Separation Rapids pluton and Pakeagama Lake Pegmatite. Such selection criteria will help ensure that the first analyses are free from inclusions that could affect the \(^{40}\text{Ar}/^{39}\text{Ar}\) ratio and hence allow a correction procedure to be set-up that will compensate for atmospheric argon in the
samples and other interfering reactions produced during the irradiation.

REFERENCES


B. Berdusco
Ontario Geological Survey, Precambrian Geoscience Section

INTRODUCTION

The Ontario Geological Survey has initiated several projects in an effort to streamline its digital mapping methodology and internal cartographic process. Included in these projects are:

1. The OGS Guide to Digital Mapping
2. The Point of Observation Data Collection Process
3. The Geochemical Analyses Request Application
4. Metadata and Geologic Map Standards

THE OGS GUIDE TO DIGITAL MAPPING

The OGS Guide to Digital Mapping is under development. This guide will provide standards and procedures for creating digital maps in conjunction with AutoCAD and Fieldlog largely for OGS mapping projects. We expect, though, that this guide may be useful to others; therefore, it may be released during the winter of 1998–1999. AutoCAD is a desktop drafting application distributed by Autodesk, Inc. Fieldlog is an AutoCAD add-on package that assists users in linking attribute data in databases to vector AutoCAD drawings. Fieldlog originated in the Ontario Geological Survey. The Geological Survey of Canada has continued development of Fieldlog. The latest version, version 3.0, runs on AutoCAD versions 12, 13 and 14.

Standard symbol libraries continue to be developed at the Ontario Geological Survey. The OGS Library of Bedrock Mapping Symbols was published in 1995 (Jackson, Muir and Romkey 1993 and Muir 1995). This comprehensive 1660 symbol library contains symbology for bedding, cleavage, fold axial planes, faults, foliations, fractures, igneous contacts, igneous layering, lineations, palaeocurrents and veinings. Recently, the OGS completed a final review of a line type library consisting of over 400 line types and line annotations for geologic folds, faults and contacts. This standard will also be released as an Open File Report.

The OGS is continuing development on mineral deposit symbol standards and symbol standards reflecting zones of alteration and deformation (e.g., stipple patterns). Other symbol libraries under consideration include metallic classification symbology, Quaternary symbology, commodity symbology and mining symbology. The development process for each symbol library requires considerable resources. The OGS has reviewed symbol libraries from other provincial, state and federal surveys in an effort to decrease development time.

The Guide to Digital Mapping provides instructions for using the various annotation standards developed by the OGS. In addition, guidance is provided regarding digitizing techniques specifically for our internal cartographic process and the development of Fieldlog databases. Data transfer instructions help the user to import and export data to and from their AutoCAD digital mapping projects.

THE POINT OF OBSERVATION DATA COLLECTION PROCESS

Recently, the OGS has moved towards data collection on the outcrop that utilizes hand held pen based computer systems. Currently, geoscientists enter data into Apple Newtons. The FieldWorker PRO software package from FieldWorker Products Ltd. replicates on the Newton, the Fieldlog database structure used in association with AutoCAD. Daily traverse data is dumped to field laptop computers and coordinates are entered from air photos. Though FieldWorker PRO is capable of handling satellite GPS data directly, this process is rarely used since most mapping continues on air photos, a necessity in remote mapping projects. Typically, point of observation data includes rock type, alteration, structural readings, bedding orientations, mineralization, photograph data and sample analysis data. The use of pen based systems has significantly diminished data entry time in the evenings. However, geoscientists do find that they tend to spend up to 20% more time on the outcrop entering data. At the end of each field season, geoscientists provide debriefings that result in technical modifications of the data capture process. This iterative design process is continuously ongoing as new technology is constantly integrated into the field mapping projects.

THE GEOCHEMICAL ANALYSES REQUEST APPLICATION

The Geochemical Analyses Request (GAR) application was designed and created, in concert, by the Data Services Section, Geoscience Laboratories and the Precambrian Geoscience Section. GAR has several purposes. It provides complete sample documentation in digital form for requesting analytical work. It provides sample documentation to be used in geoscience reports. GAR is also used to populate the Lithogeochemical Database (LGC). It stores
descriptive information including sample number, requested analysis type, coordinate locations and sample descriptions. The basic process of which GAR is a part can be summed up as follows. Geoscientists sample rocks in the field. Sample information is input into GAR through Fieldlog or through data entry directly in the application. GAR contains a listing of all analytical packages and methods in use by the Geoscience Laboratories. The geoscientist uses GAR to specify custom analyses or request analyses from analytical packages for their samples. Typically the types of analyses are dependent on the type of study that the geoscientist is doing, (e.g., assay, rock identification, rare earth element study, alteration study, etc.). Once the analytical request has been defined, the request is made in electronic fashion to the laboratory. The Geoscience Laboratories perform the request and return the sample information to the geoscientist electronically. The GAR application imports the analyses and merges this with the descriptive data. The geoscientist can extract sample descriptions and analyses for a final report. The geoscientist verifies the analyses and removes any confidentiality flags, if permissible under any standing agreements, prior to uploading the data through GAR to the Lithogeochemical Database (LGC). This process is required to keep the LGC current. The data in the LGC is made available to clients through the Earth Resources and Land Information System (ERLIS) and on various media including CD-ROM. In the future, analyses in the LGC will be made available through the Earth Resources Mineral Exploration Site (ERMES) on the World Wide Web.

METADATA AND GEOLOGIC MAP STANDARDS

The Ontario Geological Survey is involved in the joint Geological Survey of Canada (GSC)–United States Geological Survey (USGS) National Geologic Map Database (NGMD) Project. (At the time of writing, a draft copy of the Digital Geologic Map Data Model was available on the World Wide Web at http://geology.usgs.gov/dm.) Wherever possible, the OGS participates with other organizations in developing metadata and geologic map standards. The importance of metadata is growing, as is the volume of accessible geoscience data. Metadata provides a description of the types of data making up a digital product such as a map, a contact name, the date that the data was acquired and guidelines for using the data. Additional information relating to map projections and transformations provides immeasurable assistance especially when map data is incorporated into geoscientific compilations in later years. Currently, the OGS participates in standards projects with the Geological Association of Canada (GAC), the Geological Survey of Canada and the United States Geological Survey. In addition, the OGS is informed of provincial standards through partnerships with other ministries. The information acquired through participation is used to develop map products with a common look and feel and a consistent means of use. The OGS strives to remain current with the ongoing development of provincial, federal and international geospatial data infrastructures. The goal of all geospatial data infrastructures is to facilitate the sharing of data.

REFERENCES


27. Project Unit 95–027. Refinement of Hafnium (Hf) and Zirconium (Zr) ICP-MS Analysis by Improvement in the Sample Digestion Procedure

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2Geoscience Laboratories, Ontario GEOservices Centre

INTRODUCTION

Accurate and precise analysis of Zr and Hf by inductively coupled plasma mass spectrometry (ICP-MS) is becoming increasingly important as these elements are petrogenetically significant. Zirconium and Hf, along with the rare earth elements (REE) and other high field strength elements (HFSE), are used by researchers and industry for interpreting the petrogenesis and tectonic environment in which packages of volcanic rocks formed (e.g., Xie, Kerrich and Fan 1993). Determining the petrogenesis and tectonic environment of volcanic rocks can be a key factor in deciding whether the rocks are likely to host ore deposits (e.g., Kerrich and Wyman 1996). The Geoscience Laboratories have been routinely analyzing geological samples for Zr by X-ray fluorescence (XRF) and Hf (and the REE) by ICP-MS. Samples to be analyzed by ICP-MS are prepared using acid digestion to achieve, as close as possible, total dissolution. Zirconium is also measured by ICP-MS but is not reported, rather it is compared with the Zr XRF results and used as a check on the efficiency of the ICP-MS acid digestion procedure. Certain mineral phases (e.g., zircon, spinel, titanite and garnet) are resistant to dissolution and low Zr values by ICP-MS compared with XRF have been found to result from incomplete acid digestion. This incomplete digestion also produces low values of Hf and may produce low values of heavy rare earth elements (HREE). For this reason, during 1997, an attempt was made by the authors to improve the acid digestion procedure so that both Zr and Hf could be analyzed accurately and precisely by ICP-MS. This report describes a variant of existing techniques (e.g., Jenner et al. 1990).

THE PROBLEM

The routine sample preparation method for ICP-MS analysis in the Geoscience Laboratories has involved mixed acid digestion (HF, HCl, HClO4—the T4A acid) on a hot plate in open beakers. Hydrofluoric acid (HF) is the important acid in attacking the phases that host Zr and Hf, and during routine preparation the samples are in contact with the HF for 2 days. If incomplete digestion is suspected (e.g., if a residue is present), then a second addition of the T4A acid (containing HF) can be carried out to double the time in which the samples are in contact with HF. As the samples are in open beakers on a hot plate, the T4A acid evaporates and is gradually lost as the digestion proceeds, with most of the acid being lost in the first 24 hours. Table 27.1 and Figure 27.1 show the results of open-beaker double digestion on 3 samples. In Figure 27.1, the samples are plotted on a conventional spider diagram, the profile of which is useful in determining the petrogenesis and possible tectonic environment of a particular rock. In a conventional spider diagram, elements should generally plot with a smooth profile unless some process has removed specific elements. Such processes may be petrological or related to alteration or incomplete sample digestion. The closed circles in each spider diagram represent the open-beaker digestion of each sample. Samples 96-KYT-13 and 96-KYT-66 are altered mafic volcanic rocks from the Archean Unbmy Lake greenstone belt (western Superior Province). The pronounced negative Zr and Hf anomalies on the first two spider diagrams are the result of incomplete digestion of the samples prepared in open beakers. This is suggested because the XRF values for Zr (shown with squares on the first two spider diagrams) are higher and plot on a smooth line compared with the other elements in the spider diagrams. Similar low ICP-MS values of Zr and Hf have been found for several hundred samples of mafic and ultramafic volcanic rocks analyzed by the lab (K.Y. Tomlinson, Laurentian University, unpublished data, 1994–1997).

Negative Zr and Hf anomalies are rare in mafic rocks because such anomalies commonly result from the fractionation of zircon (a Zr and Hf rich mineral). Fractionation of zircon commonly occurs in felsic rocks but not mafic rocks. Xie, Kerrich and Fan (1993) have suggested that negative Zr and Hf anomalies may also result from the high-pressure (deep in the mantle) fractionation of majorite garnet and hence such anomalies may indicate a plume source for a particular package of rocks. It is therefore important in petrogenetic studies to recognize whether negative Zr and Hf anomalies are a result of crystal fractionation or a result of incomplete sample digestion. Once we had recognized that negative Zr and Hf anomalies were resulting from incomplete sample digestion, it was important to refine the digestion procedure.
THE SOLUTION

Samples 96-KYT-13, 96-KYT-66 and a standard reference material (SRM) BIR-1 were prepared in a preliminary test using a sample digestion procedure that involved refluxing the samples in the same T4A acid but in closed (screw-top) beakers in an oven. Using closed beakers has the advantage of increasing the pressure inside the beakers, and also the acid is not allowed to evaporate until the beaker lids are removed after several days of acid attack, thereby allowing more efficient and continuous exposure of the sample to the acid. In the test, one set of samples was refluxed for 2 days while the other set was refluxed for 8 days.

RESULTS

The results are shown in Table 27.1 and Figure 27.1. Sample 96-KYT-13 shows that after a 2 day reflux in a closed beaker the negative Zr and Hf anomaly is less than it was for the open-beaker digest, but it still exists. This indicates that digestion of the sample is incomplete after 2 days of refluxing. After 8 days of refluxing in a closed beaker, the negative Zr and Hf anomaly has almost disappeared and sample digestion appears to be almost complete. Sample 96-KYT-66 shows that after 2 days of refluxing in a closed beaker the negative Zr and Hf anomaly is almost gone and the result remains essentially the same after 8 days of refluxing. Digestion of this sample was therefore complete after 2 days of refluxing in a closed beaker. Refluxing the samples in closed beakers is obviously a vast improvement over the open-beaker digest for these two altered volcanic rock samples, although the duration of refluxing required for complete digestion appears to vary for each sample.

The SRM shows different results. BIR-1 is an unaltered basalt and whichever digestion procedure is used, the results for Zr and Hf are all similar and comparable to published proposed values (shown with squares on Figure 27.1; Govindaraju 1994). Although this sample shows a negative Zr anomaly, the anomaly appears to be real as the different preparation methods give results that are in agreement. The other SRMs commonly used by the Geoscience Laboratories also show good agreement in the results for Zr between XRF and ICP-MS (after open-beaker digestion). The SRMs used are commonly unaltered rocks, and it appears that open-beaker digestion may be satisfactory for unaltered rocks but unsatisfactory when more highly altered and metamorphosed Archean volcanic rocks are being analyzed. Further results from a number of SRMs (BHVO-1, UB-N, DR-N and BE-N) analyzed using the closed-beaker preparation method are published in Tomlinson et al. (in press).

In conclusion, when altered rocks are being analyzed, and high quality Zr and Hf data are required for detailed petrogenetic work, it is necessary to prepare the samples by a closed-beaker reflux method before ICP-MS analysis.

TANTALUM VARIATION

The variation in tantalum (Ta) values in Figure 27.1 and Table 27.1 is also significant and worthy of note. In
Table 27.1. Comparison of results for the REE and HFSE for different sample preparation techniques.

<table>
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<tr>
<th>Sample</th>
<th>Y</th>
<th>Zr</th>
<th>Nb</th>
<th>La</th>
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<th>Pr</th>
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<th>Sm</th>
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<th>Tb</th>
<th>Gd</th>
<th>Dy</th>
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<th>Er</th>
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<td><strong>Recommended/proposed values (Govindaraju 1994)</strong></td>
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particular, the values for BIR-1 are much higher than the proposed value. The Ta variations may result from a number of possible areas, including the sample preparation procedure or calibration of the ICP-MS. This is under further investigation in the ICP-MS lab.

**SUMMARY**

Zirconium and Hf are important elements in determining the petrogenesis and tectonic environment of suites of volcanic rocks. In altered and metamorphosed Archean volcanic rocks that have been analyzed for Zr and Hf by ICP-MS following open-beaker acid digestion, a negative Zr and Hf anomaly commonly occurs on spider diagrams (K.Y. Tomlinson, Laurentian University, unpublished data, 1994–1997). Zirconium values measured by ICP-MS also fall below those values measured by XRF in the same sample. These low Zr and Hf values are a result of incomplete acid digestion of the samples. The sample digestion procedure has been improved by using closed (screw-top) Savillex® beakers instead of open beakers and refluxing the samples in acid in an oven. The new sample preparation procedure being used for ICP-MS analysis when high quality Zr and Hf data are required is described in the appendix.

**ACKNOWLEDGMENTS**

We are grateful to James Schweyer for help with preparing samples.

**REFERENCES**


Tomlinson, K.Y., Hughes, D.J., Thurston, P.C. and Hall, R.P., in press. Plume magnetism and crustal growth at 2.9 to 3.0 Ga in the Steep Rock and Lumby Lake area, Western Superior Province; Lithos.


**Appendix: Closed-beaker reflux sample preparation procedure now in use by the Geoscience Laboratories prior to ICP-MS analysis**

**REAGENTS**

Hydrofluoric acid (HF) 48 to 51%, hydrochloric acid (HCl) 36 to 38%, nitric acid (HNO₃) 69 to 71%, perchloric acid (HClO₄) 62 to 70%.

One batch of T4A reagent comprises 800 mL HF, 80 mL HCl and 80 mL HClO₄.

One batch of T4B reagent comprises 760 mL deionized distilled water, 140 mL HCl and 30 mL HClO₄.

One litre of internal standard solution comprises 10 mL of 1000 µg/mL Ru, 10 mL of 1000 µg/mL Re and 10% HNO₃.

**SAMPLE DISSOLUTION**

1. Weigh 0.200 g of sample into a 60 mL Savillex® screw-top bomb. Add 15 mL of T4A reagent. Cap the bomb tightly and place into an oven at 120°C for about 7 days.

2. Remove from the oven and allow to cool.

3. Uncap and pour contents into 50 mL PTFE beaker. Rinse bomb with 10% HNO₃. Evaporate sample to dryness on a hot plate at 120°C.

4. Add 15 mL of T4A reagent, evaporate sample to dryness on a hot plate at 120°C.

5. Add 8 drops of concentrated HCl, allow to cool for 1 minute, add 1 mL of concentrated HNO₃, add 4 drops of concentrated HF and swirl the solution, add 15 mL of deionized distilled water and evaporate on a hot plate to reduce to 10 mL.

6. Remove sample from hot plate and allow to cool.

7. Pipette 1.0 mL of internal standard solution to a 100 mL volumetric flask. Transfer sample to flask and bring to volume with 10% HNO₃. Shake thoroughly.

8. Transfer an aliquot of the solution to a 12 mL snap-cap polystyrene test tube. The solution is ready for analysis by ICP-MS.
28. Project Unit 97–018. The Project and Results Management Process

L.L. Churchill
Ontario Geological Survey, Precambrian Geoscience Section

INTRODUCTION

Modern businesses, whether public or private sector, are faced with new challenges. These challenges range from modernizing comptrollership (linking costs to results), to requiring an enhanced understanding of accountability, to entering into partnerships/collaborative agreements to deliver programs effectively and efficiently with reduced resources.

The project and results management process is commonplace in both the public and private sectors. The current operating environment demands that public-sector programs

- be accountable for public funds and demonstrate value for public investment,
- have costs linked to results, and
- be effectively managed for optimal program delivery and client service.

THE PROJECT MANAGEMENT CYCLE

To meet these demands, the Ontario Geological Survey (OGS) has revitalized its project management and results analytical cycle. The project management cycle consists of the following stages:

1. Project Proposals: accept all geoscience proposals from internal and external clients and log and track proposals through the next steps in the cycle.
2. Project Evaluation: a joint OGS/client committee evaluates proposals using established selection criteria; those proposals not meeting these criteria may be returned to proponents for updating or filed for future reference.
3. Project Selection: resource availability reviewed by OGS; determination of whether proposal can be undertaken immediately by OGS alone, or collaboratively, or may be deferred until resources become available.
4. Project Definition: project specifications are drafted by OGS to include all project-related information, such as objectives, milestones, deliverables, resources required, benefits, internal/external dependencies and partnerships.
5. Project Implementation: project is implemented as per project specifications.

6. Results and Impact Monitoring: economic and scientific impacts are monitored prior to project implementation, throughout project execution and after project completion.

This project management cycle ensures that the OGS has essential program and project information to

- ensure clients receive quality and relevant geoscience products in a timely fashion;
- demonstrate that geoscience results meet external/internal client needs and internal business plan performance standards, such as stimulation of economic activity in Ontario;
- identify potential issues and potential successes (scientific and/or economic) prior to initiating a project;
- attribute private-sector investment to specific OGS projects, where applicable;
- assign, monitor and adjust staff resources and budgets for in-house OGS projects and projects jointly delivered with private sector, university and other government (federal or provincial) partners; and
- feed into the ministry business planning process, helping to ensure that the geoscience program is adequately funded.

BENEFITS

Clients and OGS program staff benefit through this project in a number of ways.

Clients

Benefits to clients include

- a geoscience program, planned and resourced, to deliver products in a timely fashion;
- products that are scientifically sound and designed to be results oriented;
- timely and accurate information about the OGS geoscience program, project-related information and project impacts available to the OGS Advisory Board, its technical subcommittees and other client groups; and
- more effective communication about the OGS geoscience program and its impacts.

OGS Program Staff

Benefits to OGS program staff include
• a clear and simple process to communicate project successes to senior management and political staff;
• recognition from clients and senior ministry and political staff for their geoscience project accomplishments and contributions to science and the economic well-being of Ontario;
• more effective work scheduling, particularly where work or milestone dependencies exist between staff and external partners;
• identified project resource shortfalls, thereby mitigating project disruption; and
• a clearer understanding of project goals, deliverables and related milestones.

METHODOLOGY

The approaches of other public and private sector organizations have been researched to identify the most appropriate project and results management approach for the OGS.

During the design stage, the project focused on the Precambrian Geoscience Section and the Resident Geologist Program. Emphasis was placed upon the
• collection of information to describe the project goals, objectives, specifications, deliverables and milestones;
• definition of project and program performance and successes;
• evaluation of project success using measures of productivity (e.g., area mapped), scientific quality and private-sector investment (verified by clients, some of which provided testimonials documenting OGS program relevance and accounts of private-sector jobs created) levered by the OGS program; and
• definition of a process to measure OGS program success and exploration investment that can be attributed to OGS programs.

Information collected from all phases of the project and results management process is stored in a standard database application. Linkages throughout the database allow standardized reports to be produced that can portray the information in a number of formats, dependent upon the audience.

Credibility and integrity of the data are paramount; therefore, data are continually validated and approved by those who provided the information.

CLIENT AND STAFF INVOLVEMENT

Input is required from our clients to monitor the use and impact of OGS products. If an OGS map or report were instrumental in stimulating exploration, then it is essential that we capture this information. To date, those companies that have quoted our products or services in a press release have been contacted. Information is requested on their exploration expenditures levered by the OGS program and these activities are monitored. These private-sector expenditures are then compared with OGS expenditures to calculate a ratio of private-sector to public-sector investment.

NEXT STEPS

The project and results management process is not fully implemented. We will modify, update and continue to populate the database to capture all OGS project information.

The OGS will finalize and implement the project proposal and project evaluation stages in the near future, completing the design of the project management cycle.

Rigorous project management provides many benefits to clients and OGS staff. The success of relating project impacts and benefits is highly dependent upon the information provided by clients and staff. We will continue to solicit project-related impact information.

By implementing a project and results management process, the impact and value of OGS projects can be documented. By capturing this information in an objective and consistent format, the OGS is better prepared to answer to internal and external requests for information. This facilitates effective and efficient program management.

R.M. Easton
Ontario Geological Survey, Precambrian Geoscience Section

INTRODUCTION

Geologic mapping in Street Township (Figure 29.1), which straddles the Grenville–Southern province boundary, was initially undertaken in 1996 in order to resolve discrepancies in Huronian Supergroup stratigraphy as presented on several published geologic maps of the area, to determine why the trend of the Grenville–Southern province boundary changes direction in the township, and to understand the setting of garnet deposits in the township (Easton et al. 1996; Easton 1996). This mapping led to the identification of several previously unidentified varieties of mafic rocks within the Grenville Province in Street Township. This report discusses the mineral potential of these mafic rocks in light of subsequent geochemical, geochronological and detailed mapping studies, and supplements results reported in Easton (1998). Although this report focuses on rocks located within the Grenville

Figure 29.1. Location of Street Township, distribution of Paleoproterozoic mafic intrusive rocks and major geological boundaries, and U/Pb zircon ages on intrusive rocks in the Sudbury area. Sources of U/Pb ages are given in Easton (1998). Abbreviations: EBLI – East Bull Lake Intrusion, EBLI suite – East Bull Lake Intrusive Suite, RVI – River Valley intrusion, SIC – Sudbury Igneous Complex.
Province in Street Township, the recommendations described herein are applicable also to mafic rocks within Crerar, Dana, Henry, Loughrin, and southernmost Davis and Janes townships.

**MAFIC ROCK SUITES WITHIN THE GRENVILLE PROVINCE**

Five main groups of mafic rocks can be identified within the Grenville Province in Street Township, using a combination of field and geochemical techniques (Easton and James 1997; Easton 1998). Where geochemical data are available for the mafic gneisses, rock names based on normative mineralogy following International Union of Geological Sciences (IUGS) nomenclature are given in italics for comparative purposes (e.g., melanoamphibolite = *gabbronite*). This approach is similar to that used by Peck et al. (1995) in studying the East Bull Lake intrusion. Briefly, from most to least abundant, these are:

1. Migmatitic amphibolite (*gabbronite* and *norite*). These can be divided into garnet-rich and garnet-poor subtypes, and can be further subdivided locally using migmatitic textures.

2. Non-migmatitic melanoamphibolite and amphibolite (*gabbronite* and *norite*). These rocks can be divided into 2 chemical groups, one with Mg-numbers greater than 60 and another with Mg-numbers between 35 and 60. Both chemical groups are similar in chemistry to rocks of the Nipissing diabase suite (ca. 2.22 Ga).

3a. Non-migmatitic gabbroic meta-anorthosite, anorthositic metabasalt, melange and melanosite (gabbroic anorthosite, anorthositic gabbro, leuconorite) are likely correlative with the River Valley gabbroic meta-anorthosite (ca. 2.475 Ga) based on similarity in rock types, relative age relationships, and on-surface proximity to mapped parts of the River Valley body. These rocks are located in a body present in southwestern Street Township, herein referred to as the Street anorthositic gabbro (new name), and are described in greater detail below.

3b. Orthopyroxene hornblendite (olivine orthopyroxenite and olivine websterite), which occurs as small (<1 km diameter) bodies throughout the area. Based on U/Pb geochronology, these bodies are likely related to the River Valley meta-anorthosite suite (Corfu and Easton 1998).

4. Amphibolite and garnet amphibolite (*tholeiitic basalt*) located between the Ess Creek fault and the Grenville Front Boundary Fault as defined by Lumber (1973), which are geochemically correlative with the Huronian Stobie Formation (Easton and James 1997; Easton 1998).

**ASSAY RESULTS FROM MAFIC ROCKS WITHIN THE GRENVILLE PROVINCE**

Table 29.1 lists assay results for chromium, copper, nickel, platinum, palladium and gold obtained from representative samples from all 5 mafic rock groups in Street Township, except for those of the anorthositic metagabbro association which are summarized in Table 29.2. None of the samples listed in Tables 29.1 or 29.2 contained significant amounts of visible sulphide, consequently, these samples serve mainly to establish background levels for chromium, copper, nickel, platinum and palladium within the various mafic rock units. Of the 5 mafic rock groups under consideration, only 3 show metal values above detection limits.

As has already been noted in Easton (1998), orthopyroxene hornblendite contains significant chromium (see Table 29.1), hosted mainly in Fe-chromite and Cr-spinel. Orthopyroxene hornblendite and hornblendeite associated with the Street anorthositic metagabbro body (see Table 29.2) also show elevated chromium contents, but lower than those found in bodies located marginal to anorthositic metagabbro bodies (see Table 29.1). In the orthopyroxene hornblendeites currently sampled, platinum and palladium are near or below detection limits (see Tables 29.1 and 29.2). Exceptions are sample 98RME-0050, a gabbroic anorthosite collected within 50 m of an orthopyroxene hornblendite pod within the Street anorthositic metagabbro body, and sample 98RME-0056, an anorthositic gabbro containing hornblendeite layers also from the body, both of which show enrichment in PGE. This suggests at least a spatial link between the hornblendeites and PGE mineralization.

Three of 9 non-migmatitic amphibolites samples show anomalous Pt and Pd values (see Table 29.1, samples 96RME-0446, –1056, –3150) ranging from 32 to 51 ppb combined Pt and Pd. Sample 96RME–3150 also contains anomalous copper, which may be a reflection of the higher sulphur content of this sample (i.e., the sample contains chalcopyrite). The anomalous platinum group elements (PGE) and copper content in the non-migmatitic amphibolite is consistent with the Nipissing diabase-like geochemical signature of these rocks. It is also consistent with the observation that PGE- and copper-nickel-bearing Nipissing sills within the Southern Province occur mainly in the Sudbury area (e.g., Fyon et al. 1992; Lightfoot and Naldrett 1996).

Two samples spatially associated with the Street anorthositic metagabbro body contained the highest platinum and palladium values obtained in the initial sampling program, namely samples 96RME–0200 and 96RME–0213, which yielded 49 and 112 ppb combined Pt and Pd, respectively (see Table 29.2). Sample 96RME–0200 was from a high Mg-number, non-migmatitic, melanoamphibolite which may either be a Nipissing diabase or a feeder to the anorthositic metagabbro body. Sample 96RME–0213 is from a gabbroic meta-anorthosite horizon. In order to further evaluate the significance of these 2 anomalous samples, additional sampling and
<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Easting</th>
<th>Northing</th>
<th>Field Rock Name</th>
<th>Cr</th>
<th>Cu</th>
<th>Ni</th>
<th>Pt</th>
<th>Pd</th>
<th>Pt/Pd</th>
<th>S</th>
<th>Au</th>
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<td></td>
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<td></td>
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<tr>
<td>96RME-0028</td>
<td>528675</td>
<td>5156195</td>
<td>grt amphibolite</td>
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<td>83</td>
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<td>N.D.</td>
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<td>N.D.</td>
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<td>amphibolite</td>
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<td>40</td>
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<td>5162009</td>
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<td><strong>Orthopyroxene hornblendite</strong></td>
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<td>528938</td>
<td>5157144</td>
<td>opx hornblendite</td>
<td>&gt;3000</td>
<td>160</td>
<td>870</td>
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<td>0.01</td>
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<td>524960</td>
<td>5153570</td>
<td>opx hornblendite</td>
<td>&gt;3000</td>
<td>16</td>
<td>850</td>
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<td>N.D.</td>
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<td>5153690</td>
<td>opx hornblendite</td>
<td>2500</td>
<td>N.D.</td>
<td>880</td>
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<td>5151800</td>
<td>opx hornblendite</td>
<td>~4000</td>
<td>9</td>
<td>750</td>
<td>N.A.</td>
<td>N.A.</td>
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<tr>
<td><strong>Non-migmatitic amphibolite (Mg#&gt;60, possible high-Mg Nipissing diabase?)</strong></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td>96RME-0446</td>
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<td>5162960</td>
<td>gabbro-norite</td>
<td>240</td>
<td>50</td>
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<td>amphibolite</td>
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<td>54</td>
<td>190</td>
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<td>N.D.</td>
<td>&lt;0.01</td>
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<td>5153600</td>
<td>amphibolite</td>
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<td>140</td>
<td>N.D.</td>
<td>N.D.</td>
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<td>N.A.</td>
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<td>170</td>
<td>30</td>
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<td>5162840</td>
<td>amphibolite</td>
<td>750</td>
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<td>140</td>
<td>11</td>
<td>N.D.</td>
<td>&lt;0.01</td>
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<td>310</td>
<td>50</td>
<td>15</td>
<td>17</td>
<td>0.9</td>
<td>0.03</td>
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<td>amphibolite</td>
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<td>72</td>
<td>50</td>
<td>N.D.</td>
<td>N.D.</td>
<td>&lt;0.01</td>
<td>N.A.</td>
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</tr>
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<td><strong>Huronian metavolcanic and dike rocks</strong></td>
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<td></td>
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<td>5162930</td>
<td>mafic metavolcanic</td>
<td>81</td>
<td>83</td>
<td>77</td>
<td>N.D.</td>
<td>N.D.</td>
<td>&lt;0.01</td>
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<td>5162892</td>
<td>felsic metavolcanic</td>
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<td>12</td>
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<td>0.12</td>
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<td>531900</td>
<td>5162120</td>
<td>grt amphibolite dike</td>
<td>38</td>
<td>180</td>
<td>65</td>
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<tr>
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<td>mafic metavolcanic</td>
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<td>5162800</td>
<td>mafic metavolcanic</td>
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<td>10</td>
<td>9</td>
<td>N.D.</td>
<td>N.D.</td>
<td>0.04</td>
<td>N.D.</td>
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</table>
| 96RME-3004 | 532009  | 5162502  | mafic metavolcanic | 63 | 290| 73 | N.D.| N.D.| 0.21 | 6
| 96RME-3030 | 531960  | 5162357  | schist           | 41 | 9  | 30 | N.D.| N.D.| 0.01 | N.D.|
| 96RME-3031 | 531900  | 5162260  | schist           | 140| 40 | 54 | N.D.| N.D.| 0.05 | N.D.|
| 96RME-3032 | 532200  | 5162530  | schist           | 61 | 9  | 20 | N.D.| N.D.| 0.01 | N.D.|

Abbreviations: grt = garnet, opx = orthopyroxene, N.D. = not detected, N.A. = not analyzed

* number in brackets indicates detection limit for the respective element
mapping was conducted in 1998, as reported in Table 29.2 and Figure 29.2a and 29.2b.

The 2 additional samples with the highest PGE contents, samples 98RME-0024 and 98RME-0064, containing 166 and 229 ppb combined Pt and Pd, respectively, come from the same area of the body as sample 96RME-0213 (see Figure 29.2b), suggesting that there is a discrete mineralized horizon is present in this part of the intrusion. There is no obvious relationship between PGE content and nickel, copper or sulphur contents in these 3 samples. Four additional samples showing atypical PGE values are samples 98RME-0023, -0054, -0056 and -0073, which contain 59, 37, 82 and 38 ppb combined Pt and Pd, respectively. These 4 samples are all from the northwestern margin of the body (see Figure 29.2b).

The 1998 sampling indicates that initial sampling was reliable, and that atypical PGE values occur in two areas of the intrusion, namely the central core and the northwestern margin. All samples with combined platinum and palladium greater than 37 ppb are from anorthositic gabro or gabbroic anorthosite units, as is the case in the East Bull Lake intrusion (Peck et al. 1995).

### Table 29.2. Assay results from the Street Township anorthositic metagabbro body.

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Easting</th>
<th>Northing</th>
<th>Field Rock Name</th>
<th>Cr (ppm)</th>
<th>Cu (ppm)</th>
<th>Ni (ppm)</th>
<th>Pt (ppb)</th>
<th>Pd (ppb)</th>
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* abbreviations: grt = garnet, opx = orthopyroxene, N.D. = not detected, N.A. = not analyzed
* number in brackets indicates detection limit for the respective element
Figure 29.2. a) Geological map of the Street anorthositic metagabbro body; b) location of assay samples from the Street anorthositic metagabbro body reported in Table 29.2. The prefixes 96RME– and 98RME– in the sample numbers have been omitted for clarity; c) inferred cross section of the Street anorthositic metagabbro body showing the effects of folding on the body.
This “border zone” may represent a feeder dike system to the body, as melanoamphibolite indistinguishable from that in the border zone occurs within the main mass of the body.

Figure 29.3 is a composite section through the metagabbro body, showing the main units present within the intrusion. Foliated anorthositic metagabbro and gabbroric meta-anorthosite comprise the lowermost unit (lower series) within the intrusion, with a maximum thickness of roughly 150 m. Variation in plagioclase content within this unit may reflect primary layering in the body, as this layering is cut by 50 to 75 cm wide mafic dikes, now garnet amphibolites, which were metamorphosed and deformed with the anorthositic rocks. Locally, near the top of this unit are found elongate inclusions of white-green weathering, fine-grained epidote-garnet-plagioclase rock and dark green weathering amphibolite which may represent recrystallized anorthosite fragments and fragments of country rock. Because of refolding, the

Figure 29.3. Generalized cross sections for the Street, East Bull Lake and Shakespeare-Dunlop intrusions of the EBLI suite. East Bull Lake section modified from Peck et al. (1995), Shakespeare-Dunlop section modified from Vogel (1996).
lower series is re-exposed in the core of the body (see Figure 29.2).

The lower series is overlain by the main series, a sequence of melanoamphibolite (gabbro-norite), thinly layered melanoamphibolite (gabbro-norite) and norite (leucogabbro norite), that has a maximum thickness of 250 m. The relationship of the norite to the melanoamphibolite is unclear. The norite appears as both metre-scale pods and layers in the melanoamphibolite. The norite may simply represent areas more resistant to metamorphic recrystallization, and thus has no stratigraphic significance. This is suggested by the fact that primary orthopyroxene in the norite is armoured by pale amphibole, biotite and garnet, and that a progression from almost completely preserved orthopyroxene (thinnily armoured) to almost completely recrystallized orthopyroxene (thicker armoured) is present. Alternatively, the norite could represent disrupted original layers within the main series that, because of their primary mineralogy, were more resistant to metamorphic recrystallization.

Discontinuous hornblende and orthopyroxene hornblende (olivine orthopyroxenite and olivine websterite) layers and pods on the order of 100 to 150 m long and up to 15 m wide occur sporadically at the contact between the lower and the main series (see Figure 29.3). These pods contain 30 to 40% poikiloblastic orthopyroxene. Orthopyroxene poikilolasts are smaller (<2 cm diameter) and sub-rounded in these pods with respect to the larger (>2cm), more euhedral poikilitolasts typical of the orthopyroxene hornblende elsewhere in Street Township. Megaboulders of orthopyroxene hornblende occur along the western contact of the body (see Figure 29.2a), and may represent part of a small hornblende body imbedded at the contact of the metagabbro and the host gneisses.

Until now, orthopyroxene hornblende within the Grenville Front Tectonic Zone had only been identified along the contact between anorthositic metagabbro and metagabbro and country rock. The presence of these hornblende layers within the anorthositic metagabbro confirms the previously observed spatial association between these rock units, and is consistent with U/Pb zircon geochronology (Corfu and Easton 1998) indicating that the hornblendites are the same age as the River Valley meta-anorthosite.

Figure 29.3 compares the stratigraphy of the Street anorthositic metagabbro body with that of the East Bull Lake and Shakespeare-Dunlop Gabbro-Anorthosite intrusions. The sequences are grossly similar, with the Street body being most similar to the Shakespeare-Dunlop sequence, particularly in terms of location of the inclusion-bearing zone and the fact that the gabbro-norite overlying the gabbroic anorthosite zone is poorly layered. The most significant difference is that the Street intrusion lacks part of the main series and all of the upper series present in both the East Bull Lake and Shakespeare-Dunlop intrusions. Thickness differences between the lower and main series between the 3 intrusions (see Figure 29.3) may, in part, reflect tectonic thinning in the Street body related to increased deformation.

SIMILARITIES BETWEEN THE STREET ANORTHOSITIC METAGABBRO BODY AND THE EAST BULL LAKE INTRUSIVE SUITE

The Street anorthositic metagabbro body shows a number of similarities with the East Bull Lake and Shakespeare-Dunlop (Agnew Lake) Anorthositic Gabbro intrusions located in the Southern Province west of Sudbury (see Figure 29.1), both of which contain PGE and copper-nickel mineralization. These similarities are:

1. Rocks structurally beneath the intrusions are extensively diked by metagabbroic rocks which may be feeder systems related to the intrusions.
2. Anorthositic rocks occur mainly in the basal part of the intrusions, with metamorphosed gabbroic anorthosite, leucogabbro and leucogabbro-norite being the main plagioclase-cumulate phases. All anorthositic zones contain inclusions of country rock.
3. The stratigraphy of the bodies is broadly similar (see Figure 29.3), although only the lowermost stratigraphic units are exposed in Street Township.
4. Ultramafic phases are present locally.
5. Iron- and titanium-oxide horizons are not significant phases within the intrusions.
6. The intrusions are chemically similar, characterized by low sulphur tenor, low titanium, moderate to high Al₂O₃ contents, moderate Mg-numbers, high background PGE, and show LREE enrichment with no or slightly positive europium anomalies.
7. Geochronology on the related orthopyroxene hornblende unit (Corfu and Easton 1998) suggests that the Street anorthositic metagabbro, the River Valley gabbroic meta-anorthosite, and the East Bull Lake and Shakespeare-Dunlop intrusions were all emplaced at ca. 2475 Ma. For convenience, these rocks will be referred to hereafter as East Bull Lake Intrusive Suite (EBLI suite), which includes the East Bull Lake Gabbro-Anorthosite Intrusion, Shakespeare-Dunlop Gabbro-Anorthosite Intrusion, River Valley Anorthosite, and Drury, Falconbridge, May, Street and Wisner township intrusions. The Red Deer Lake anorthositic metagabbro is also likely part of the EBLI suite.

RECOMMENDATIONS FOR EXPLORATION OF ANORTHOSITIC METAGABBROS WITHIN THE GRENVILLE FRONT TECTONIC ZONE

The presence of low sulphur tenor, high background PGE, and low titanium contents in rocks of the EBLI suite are an essential characteristic of the “second-stage magmas” from which most of the world’s major PGE deposits were formed (Hamlyn and Keays 1986). On this basis, Peck et
al. (1995) suggested the presence of the following mineralization environments within EBLI suite intrusions. All 3 types are likely present within the Grenville Front tectonic zone.

1. Large tonnage, PGE-enriched, disseminated copper-nickel sulphide deposits occurring near the floor of the intrusions. Within the Grenville Front Tectonic Zone, the floors can be identified by the presence of orthopyroxene hornblende bodies near the host rock-metagabbro contact and an increase in non-migmatic mafic dikes near the contact. Preliminary observations within the Grenville Front Tectonic Zone suggest that intrusive contacts are better preserved on the south side of the bodies, whereas contacts on the north side are generally tectonic.

2. Massive PGE and copper-nickel sulphides occurring in footwall or embayments along the floor of the intrusions. Within the Grenville Front Tectonic Zone, the presence of orthopyroxene hornblendites at the country rock-metagabbro contact may help to identify the location of such embayments.

3. Reef-type PGE and PGE-chromite deposits in the EBLI suite intrusions which contain both mafic and ultramafic cumulates. Within the Grenville Front Tectonic Zone, identifying hornblende and orthopyroxene hornblendite layers within the River Valley gabbroic meta-anorthosite, similar to those present within the Street anorthositic metagabbro, may be useful in targeting such mineralization.

Although study of the Street anorthositic metagabbro suggests that the body preserves a gross stratigraphy (see Figure 29.3) similar to that of the East Bull Lake and Shakespeare-Dunlop intrusions (and by analogy, the River Valley gabbroic meta-anorthosite), recognition of this stratigraphy is complicated by the upper amphibolite metamorphic conditions experienced by these rocks, as well as refolding of early folds and subsequent faulting. Any exploration program must take into account these geological complications.

**RECOMMENDATIONS FOR EXPLORATION OF NIPISSING DIABASE IN THE GRENVILLE FRONT TECTONIC ZONE**

The mineral potential of the Nipissing diabase sills located within the Grenville Front Tectonic Zone should be similar to that present within the immediately adjacent Southern Province. The presence of anomalous PGE and copper values reported in Table 29.1 suggests that this is indeed the case. Amphibolites derived from Nipissing diabase can be identified within the Grenville Front Tectonic Zone both by their non-migmatic character and the typical occurrence of both hornblende and clinopyroxene in hand sample. Sulphide-bearing zones near the base of the intrusions are the preferred targets. It may be possible to identify the base of the intrusions geochemically, as rocks near the base will be characterized by higher Mg-numbers.

**REFERENCES**


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²Geoscience Laboratories, Mines and Minerals Division

INTRODUCTION

Age determinations ("dating") with the uranium-lead (U/Pb) method, using chemically determined lead and uranium concentrations was introduced by Holmes as early as 1913. The major problem of this method lay in the inability to distinguish between common lead and lead produced by radioactive decay. With the development of mass spectrometric techniques allowing reliable determination of isotopes, chemical dating fell into disuse. Recently, interest in chemical dating has been revived, due to the ability to precisely determine uranium, thorium and lead using the electron microprobe (e.g., Suzuki et al. 1991, 1994, 1996; Suzuki and Adachi 1991, 1998; Rheed et al. 1996; Montel et al. 1994). These studies have focussed on monazite, mainly because the contents of uranium, thorium and lead are usually high enough to be measurable with reasonable precision by the electron microprobe. A variety of monazite types have been dated, including monazite in granites (Rheed et al. 1996; Suzuki and Adachi 1998), metamorphic rocks (Suzuki and Adachi, 1998), and detrital monazite in sedimentary rocks (Suzuki et al. 1991, 1994).

Due to the potential of the microprobe chemical dating method to provide reconnaissance level geochronology on the timing of metamorphism, a test of the method was conducted using samples from a study area within Street Township, east of Sudbury. This area was selected because of the presence of monazite-bearing schists and gneisses for which pressure-temperature (P-T) conditions of metamorphism were known (e.g., Buckley et al. 1997; Murphy et al. 1998) and were sufficiently high to eliminate any inherited component in the monazite (Williams 1996), and because of the presence of complementary U/Pb zircon, monazite, titanite and apatite ages on some of the same samples (Corfu et al. 1998).

MONAZITE ANALYSIS ON THE MICROPROBE

Monazite grains on polished thin sections were analyzed on a Cameca electron microprobe equipped with 3 wavelength-dispersive type spectrometers. The instrument operating conditions were: 20kV accelerating voltage, 200 nA probe current and 3-5 μm probe diameter. The ThO₂, UO₂ and PbO lines were measured with a PET crystal for several spots on individual monazite grains, using count times of 30, 100 and 130 seconds, respectively. The detection limit of PbO at a 2 sigma confidence level is 0.007 weight %, and the possible error in PbO determination is about 10% for 0.02 weight %.

CHIME AGE CALCULATION

Details of the chemical Th-U-total Pb isochron (CHIME) method are given in Suzuki and Adachi (1994) and Suzuki et al. (1994). Two ages can be calculated. A model age can be calculated for each analyses by solving the equation:

\[ \frac{\text{PbO}}{W_{\text{pb}}} = \frac{\text{ThO}_2}{W_{\text{th}}} \left( \exp(\lambda_3 t) - 1 \right) + \frac{UO_2}{W_{\text{u}}} \left( \exp(\lambda_5 t) + 137.88 \exp(\lambda_8 t/138.88) - 1 \right) \]

where\( W \) is the gram-molecular weight of each oxide,\( W_{\text{pb}} = 224, W_{\text{th}} = 264 \) and \( W_{\text{u}} = 270 \) and \( \lambda \) the decay constant for each isotope (\( \lambda_2 = 4.9475 \times 10^{-11} \text{y}, \lambda_5 = 9.8485 \times 10^{-10} \text{y} \) and \( \lambda_8 = 1.55125 \times 10^{-10} \text{y} \)).

In addition, the sum of measured ThO₂ and UO₂ can be combined to yield ThO₂* which yields the equation:

\[ \text{ThO}_2^* = \text{ThO}_2 + \frac{UO_2 W_{\text{th}}}{W_{\text{th}}} \left( \exp(\lambda_2 t) - 1 \right) \left( \exp(\lambda_5 t) + 137.88 \exp(\lambda_8 t/138.88) - 1 \right) \]

If individual parts of a single monazite grain or cognetic grains contain the same amounts of initial Pb, but different amounts of Th and U, data will plot on a straight line with slope \( m \) and intercept \( b \).

\[ \text{PbO} = m \lambda \text{ThO}_2^* + b \]

The best-fit regression line and then be used to calculate the age \( T \) from the slope \( m \) using the equation:

\[ T = \frac{1}{\lambda_2 \ln (1 + m W_{\text{th}}/W_{\text{pb}})} \]

The intercept \( b \) of the line represents the concentration of the initial PbO. Significant amounts of initial Pb or Pb-loss would deviate the regression line from the origin, or would not form an isochron. Additional refinement of the model ages obtained using equation 1 can be made by substituting the age approximation \( T \) from equation 4, and so on.

Table 30.1 illustrates the range in ThO₂, UO₂ and PbO contents of monazites measured by other workers. The
Street Township samples have UO₂ and PbO contents near the lower range of previously chemically dated monazites, and the lowest ThO₂ contents.

**STREET TOWNSHIP RESULTS**

Sample 96RME-0234 is an upper amphibolite facies kyanite-staurolite-garnet-biotite-muscovite schist located south of the Ess Creek fault and about 100 m north of the Grenville Front Boundary Fault as mapped by Lumber (1973). Pressure-temperature conditions for this sample, using a variety of mineral calibrations, as well as TWEQU and INVEQ, were 630 to 680°C and 7 to 8 kb (Murphy et al. 1998). All analyzed grains were located in the biotite-muscovite-quartz-feldspar matrix of the sample. Results of 28 analyses of 9 different grains are given in Table 30.1 and illustrated in Figure 30.1a, a plot of ThO₂⁺ versus PbO. The average model age of the 28 analyses is 1070 ± 50 Ma, and the data points define a straight line with minimal scatter passing through the origin, yielding an age of 1245 ± 50 Ma. Alternatively, the line can be regressed through an intercept of 0.03 PbO, to give an age of 1070 ± 50 Ma, similar to the average model age. Both regressions suggest that the assumption of negligible common lead in this sample is incorrect; however, the fact that the points define a straight line suggests that sample attained isotopic equilibrium. Regardless, the model ages and the regression ages are significantly older than the 989 ± 4 Ma age for monazite.

**Table 30.1** Range of ThO₂, UO₂ and PbO in monazites dated using the microprobe.

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<th>UO₂ (wt %)</th>
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<td>0.17-0.466</td>
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</tbody>
</table>

**Figure 30.1.** ThO₂⁺ versus PbO plots for monazite analyses from 3 samples from Street Township. Solid lines in all figures represent regression lines, dashed lines represent reference lines. Numbers in (b) refer to individual analyses listed in Table 30.2 for this sample.
from the same sample obtained by Corfu et al. (1998) using standard isotope dilution-mass spectrometric techniques.

Sample 96RME–0242 is a staurolite-kyanite-garnet-biotite-muscovite gneiss located about 150 m south of 96RME–0234, about 50 m south of the Grenville Front boundary fault. Metamorphic conditions are slightly higher, 685 to 725°C and 8 to 9 kb (Murphy et al. 1998), consistent with the development of partial melt within the gneiss. Results of 32 analyses of 12 different grains are given in Table 30.2 and illustrated in Figure 30.1b. Considerably more scatter is present in this sample than was seen in sample 96RME–0234 (see Figure 30.1a). The data falls along 2 reference lines (see Figure 30.1b, solid lines) passing through the origin (i.e., assuming minimal common lead). Roughly 50% of the data points lie on a 1310 Ma reference line, with most (30%) of the remaining data points lying on a 2145 Ma reference line. This is consistent with the model ages, which fall into two groups (Tables 30.2 and 30.3, having an average age of 1873 Ma and 1136 Ma, respectively. Alternatively, the data can be interpreted as falling on a series of parallel 990 Ma reference lines containing various amounts of common lead (see Figure 30.1b, dashed lines). In either case, it is clear that the negligible common lead assumption is not valid in this instance, and it is possible that individual monazite grains are in disequilibrium with the whole rock. There appears to be some mineralogical control in sample 96RME–0242. Analyses of monazite grain 6, which is included in garnet, lies on the 2145 Ma reference line, as do analyses from grain 3, which is in contact with apatite. Grain 12, however, which also falls on the 2145 Ma line, shows no obvious mineralogical control. Except for 2 analyses from grain 9, monazites included in kyanite fall along the 1310 Ma reference line. Corfu et al. (1998) obtained an age of 987 ± 3 Ma for monazite from this sample.

Sample 96RME–0345 is a garnet-biotite-muscovite gneiss located about 2.75 km south of 96RME–0234, about 2.5 km south of the Grenville Front Boundary Fault. Metamorphic conditions are similar to 96RME–0242, i.e., 730°C and 8.1 kb (R.M. Easton, OGS, unpublished data, 1998), also consistent with the development of partial melt within the gneiss. Results of 30 analyses of 10 different grains are given in Table 30.2 and illustrated in Figure 30.1c. The data fall mainly along a 2250 Ma reference line (see Figure 30.1c), can be regressed to yield an age of 1985 Ma (see Figure 30.1c) and an average model age of 1699 Ma. Again, much of the data also lie along a 990 Ma reference line (see Figure 30.1c, dashed lines) having a significant common lead component (0.175 weight %). Monazite from this sample has not been analyzed by conventional methods. However, as samples of metamorphic zircon and titanite a similar distance from the Grenville Front also have ages in the 980 to 990 Ma range (Corfu et al. 1998), there is no reason to suspect that monazite from this sample would yield an age older than 990 Ma. As in the case of 96RME–0242, there is some indication of mineralogical control. Again, grain 7, which is included in garnet, lies on the 2250 Ma reference line. Grain 8 shows a difference from core to rim, with a more radiogenic core and a less radiogenic rim, consistent with scanning electron microscope observations of zoning in some monazite grains. Like sample 96RME–0242, there is no consistent mineralogical control that explains all the variation in the sample.

DISCUSSION
The data from samples 96RME–0234 and 96RME–0345 give the best linear arrays, however, they give ages inconsistent with known geological events in the area and ages on monazites obtained using conventional U/Pb techniques. The most likely explanation is that the critical assumption of the CHIME method, namely minimal common lead, is invalid. That this assumption is erroneous is suggested by calculations that indicate that the presence of a common lead component of only 0.05 weight % in sample 96RME–0234 would yield an age consistent with the isotope dilution age of ca. 990 Ma. In this regard, it may be noteworthy that the monazites used in our study contain half the thorium and lead of monazites used in previous studies using the CHIME method (see Table 30.1). As a consequence, the presence of even a small common lead component would have a significant influence on our results; that is, lower thorium and uranium contents result in the production of less radiogenic lead relative to common lead, which would in turn enhance the effect of even a small common lead component.

An additional complication may be the fact that 2 samples, 96RME–0242 and 96RME–0345 underwent partial melting, most likely during the Grenville Orogeny. Partial melting would result in the redistribution of lead, thorium and uranium throughout the rock, which would affect the initial isotopic composition of the monazites. The fact that the degree of discordance in the data increases from the lowest grade sample (96RME–0234) through to the highest grade sample (96RME–0345) suggests isotope mobility with increasing metamorphic grade. Loss of uranium and thorium from the rock, with retention of lead during partial melting would also serve to increase the apparent age recorded in the monazites.

CONCLUSIONS
1. As shown by sample 96RME–0234, the method can produce ages roughly consistent with ages obtained by isotopic methods. The CHIME method tells us this is a Grenvillian versus an Archean monazite, even though the presence of a common lead component makes determination of an accurate age difficult in the absence of other information.
2. The method appears to be capable of resolving differences on the order of 50 million years, as shown by the differences observed between matrix and included grains in sample 96RME–0242.
3. Samples subjected to partial melting give indecipherable results, possibly due to the presence of common lead in the monazites or the removal of uranium and thorium from the rock into leucosome. Thus, the method may not be suitable for migmatitic rocks.
Table 30.2. Analytical data and model ages for monazites from the Grenville Front tectonic zone.

<table>
<thead>
<tr>
<th>Sample Location</th>
<th>ThO₂ (wt %)</th>
<th>UO₂ (wt %)</th>
<th>ThO₂ model age (Ma)</th>
<th>PbO (wt %)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample 96RME-0234, ky-st-grt-bt-msschist</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td>grain 1, matrix</td>
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<td>0.559</td>
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<td>grain 1, matrix</td>
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<td>0.531</td>
<td>3.234</td>
<td>0.183</td>
</tr>
<tr>
<td>grain 1, matrix</td>
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<td>3.405</td>
<td>0.189</td>
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<tr>
<td>grain 2, matrix</td>
<td>2.746</td>
<td>0.519</td>
<td>3.253</td>
<td>0.178</td>
</tr>
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<td>0.528</td>
<td>3.224</td>
<td>0.195</td>
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<td>0.505</td>
<td>3.116</td>
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<td>grain 3, matrix</td>
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<td>0.496</td>
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<td>0.178</td>
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<td>0.170</td>
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Sample 96RME-0242, st-ky-grt-bt-mgsneiss

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<th>Sample Location</th>
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<th>UO₂ (wt %)</th>
<th>ThO₂ model age (Ma)</th>
<th>PbO (wt %)</th>
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<td>0.776</td>
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Table 30.2. Continued.

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<th>Sample Location</th>
<th>ThO₂ (wt %)</th>
<th>UO₂ (wt %)</th>
<th>ThO₂* (wt %)</th>
<th>PbO (wt %)</th>
<th>ThO₂* model age (in Ma)</th>
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Sample 96RME-0345, grt–bt–msgneiss

<table>
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<th>Sample Location</th>
<th>ThO₂ (wt %)</th>
<th>UO₂ (wt %)</th>
<th>ThO₂* (wt %)</th>
<th>PbO (wt %)</th>
<th>ThO₂* model age (in Ma)</th>
</tr>
</thead>
<tbody>
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<td>3.547</td>
<td>0.254</td>
<td>1398</td>
</tr>
<tr>
<td>grain 8, matrix</td>
<td>5.230</td>
<td>0.379</td>
<td>5.601</td>
<td>0.376</td>
<td>1313</td>
</tr>
<tr>
<td>grain 8, matrix</td>
<td>4.809</td>
<td>0.381</td>
<td>5.182</td>
<td>0.387</td>
<td>1456</td>
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<tr>
<td>grain 9, matrix</td>
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<td>0.468</td>
<td>4.968</td>
<td>0.414</td>
<td>1618</td>
</tr>
<tr>
<td>grain 9, matrix</td>
<td>3.982</td>
<td>0.505</td>
<td>4.476</td>
<td>0.391</td>
<td>1693</td>
</tr>
<tr>
<td>grain 9, matrix</td>
<td>3.873</td>
<td>0.419</td>
<td>4.283</td>
<td>0.406</td>
<td>1831</td>
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<tr>
<td>grain 10, matrix</td>
<td>2.388</td>
<td>0.413</td>
<td>2.792</td>
<td>0.241</td>
<td>1674</td>
</tr>
<tr>
<td>grain 10, matrix</td>
<td>3.132</td>
<td>0.984</td>
<td>4.094</td>
<td>0.413</td>
<td>1943</td>
</tr>
<tr>
<td>grain 10, matrix</td>
<td>3.366</td>
<td>0.929</td>
<td>4.274</td>
<td>0.421</td>
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</tr>
<tr>
<td>average model age</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1699</td>
</tr>
</tbody>
</table>

Abbreviations: ap–apatite; bt–biotite; grt–garnet; ky–kyanite; ms–muscovite; pl–plagioclase; rt–rutile; st–staurolite
Table 30.3. Summary of results.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Avg. Model Age</th>
<th>Regression Age through origin</th>
<th>Slope, line through origin</th>
<th>Regression Age with intercept</th>
<th>slope and intercept</th>
<th>Conventional U/Pb age (Corfu et al. 1998)</th>
</tr>
</thead>
<tbody>
<tr>
<td>96RME-0234</td>
<td>1070 Ma</td>
<td>1245 Ma</td>
<td>0.0539</td>
<td>1070 Ma</td>
<td>0.046, 0.03</td>
<td>986 ± 2 Ma</td>
</tr>
<tr>
<td>96RME-0242</td>
<td>1873 Ma</td>
<td>2128 Ma</td>
<td>0.0942</td>
<td></td>
<td></td>
<td>987 ± 3 Ma</td>
</tr>
<tr>
<td>96RME-0242</td>
<td>1136 Ma</td>
<td>1310 Ma</td>
<td>0.0568</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>96RME-0345</td>
<td>1699 Ma</td>
<td>1985 Ma</td>
<td>0.0875</td>
<td>1791 Ma</td>
<td>0.077, 0.037</td>
<td></td>
</tr>
</tbody>
</table>

4. The method may not work well for monazites containing low thorium and uranium (<5.0 weight % ThO₂ and <4.0 weight % UO₂) due to common lead effects.

5. Interpretation of the results requires a thorough knowledge of the mineralogy of the sample, including what mineral species are in contact with the monazites being analyzed.

6. Further testing of isotopically dated samples is necessary before the method can be considered a reliable reconnaissance dating tool.

REFERENCES


Williams, I.S. 1996. The response of zircon, monazite and their U-Pb isotopic systems to low-P, high-T regional metamorphism leading to host rock partial melting; in Abstracts with Program, Geological Society of America, v.28, no.6, p.A-357.
31. Project Unit 98–05. Geology and Mineral Potential of the Puzzle Lake Area, Central Metasedimentary Belt, Grenville Province

R.M. Easton

Ontario Geological Survey, Precambrian Geoscience Section

INTRODUCTION

Approximately 3.5 weeks were spent during 1998 in 1:50 000 scale geological mapping in the Puzzle–Fifth Depot Lake area (NTS 31C/10SW) (Figure 31.1) in order to better evaluate the mineral potential of this poorly known area. The mineral potential of this area was not easily predictable, as the area lies near the confluence of 3 geological domains of the Central Metasedimentary Belt (see Figure 31.1), namely the Mazinaw, Sharbot Lake and Frontenac domains, each possessing different mineral commodities and potential. In fact, existing mapping was insufficient to properly assign rocks of the Puzzle Lake area to the various established domains during the preparation of the Geology of Ontario volume (Easton 1992).

In addition to evaluating the mineral potential and geologic domain structure of the Puzzle Lake area, additional questions to be addressed by this study included determining:

1. the location of the Robertson Lake mylonite zone, including any splays
2. the extent and character of the Sharbot Lake domain in the area
3. characterization of granitoid rocks in the area with respect to established plutonic suites (e.g., Easton 1992) present in the Central Metasedimentary Belt.

Much of the mapping during this study was concentrated in the western half of the area shown in Figure 31.2.

Figure 31.1. Domain subdivision of the Central Metasedimentary Belt (after Easton 1992) showing location of the Puzzle Lake area.
PREVIOUS WORK

The only previous description of the geology of Sheffield and Camden townships within the Puzzle Lake area is that on the 1:126,720 scale regional compilation map of Hewitt (1964). Hinchinbrooke township was mapped at 1:63,360 scale by Harding (1951). In addition, Wallach (1973) studied the Hinchinbrooke pluton as part of a thesis study. Detailed mapping (1:31,680 scale) of the area immediately to the north and west of the Puzzle Lake area was conducted by Wolff (1982) and Bright (1986), respectively. Davidson and Ketchum (1993) did some limited reconnaissance work in the southeastern part of the map area.

Known mineral occurrences within the area are shown in Figure 31.3, with molybdenum and zinc being the most common commodities reported. During the first 2 decades of this century, several molybdenite occurrences were discovered and exploited, the most significant of these in terms of production being the Chisholm mine (see Figure 31.3). Malczak et al. (1985) summarize most of the base metal and molybdenite occurrences within the map area, although not all previous workings have been reported, as noted by Brack (1993), who discovered additional molybdenite workings north of Cranberry Lake.

Zinc exploration has focussed on the past-producing Slave Lake mine (see summary in Malczak et al. 1985; also see Kingston 1967; Holmstead et al. 1992; Brack 1993).
and on an area south of Norway Lake (St. Joe Canada 1985; Brack 1993) (see Figure 31.3).

**GEOLOGY**

**Overview**

The area is underlain by rocks of 2 geological domains of the Central Metasedimentary Belt, the Mazinaw domain underly-
ing the westernmost part of the area, with the Sharbot Lake domain underlying the remainder, as shown in Figure 31.2. The boundary between the 2 domains is defined by the main branch of the Robertson Lake mylonite zone (RLMZ) (see Figure 31.2). Rocks between the two branches of the RLMZ shown in Figure 31.2 are assigned to the Mazinaw domain, however, as no rock types characteristic of either domain occur in this area, these rocks could also be assigned to Sharbot Lake domain. Metamorphic grade ranges from upper amphibolite facies within Mazinaw domain rocks west of the main branch of the RLMZ, through lower to middle amphibolite facies along the main branch and immediately east of the of the RLMZ, rising to granulite facies in the eastern third of the map area. The granulite facies isograd shown in Figure 31.2 is located mainly through observations along roads, supplemented by mineralogical observations from Wallach (1973) and, consequently, it is only an approximation.

Some workers have assigned the Hinchinbrooke pluton to Frontenac domain, based on the presence of granulite facies mineralogy in much of this body, however, no significant break was observed between rocks in the eastern

![Map](image_url)

**Figure 31.3.** Mineral occurrences, soil geochemical anomalies, and interpreted mineral potential of the Puzzle Lake area.
map area and the Hinchinbrooke pluton. In addition, rocks of the Hinchinbrooke pluton are similar in age, lithology and chemistry to the Elzevir suite (Easton 1992; Wallach 1973) of Mazinaw and Sharbot Lake domains, consequently, the Hinchinbrooke pluton is assigned herein to Sharbot Lake domain. It is speculated that the granulite facies conditions observed in the eastern part of the Puzzle Lake area represents a situation similar to that observed in the Wolf Grove area to the northeast, where Frontenac domain metamorphism of ca. 1168 Ma overprints rocks of the Sharbot Lake domain (cf. Buckley et al. 1997). If so, then by analogy with the Wolf Grove area, metamorphic pressures may be higher within the Hinchinbrooke area than in the Frontenac domain proper.

**Mazinaw Domain**

Rocks within this domain consist of bands of compositionally heterogeneous, thinly layered, mafic to felsic gneisses, likely representing metamorphosed metavolcanic and metasedimentary rocks. These gneisses are probably equivalent to the supracrustal rocks preserved at lower grade within the Clare River synform in the Mellon Lake area to the west based on similarity in rock composition, general stratigraphy and geochemistry, as noted by Bright (1986). Quartzose gneisses within this sequence may represent siliciclastic metasedimentary rocks of either the Grenville Supergroup or the Flinton Group. Calc-silicate and marble units are uncommon west of Puzzle Lake. Locally, these mafic to felsic paragneisses are "juicy" migmatites, containing greater than 20% leucosome. In the ease of many of the intermediate to felsic composition gneisses, the presence of a significant orthogneiss component cannot be excluded. The heterogeneous layered gneisses envelop 2 bodies of migmatic orthogneiss, namely a granitic body west of Puzzle Lake, and the granodioritic to tonalitic Puzzle Lake pluton (see Figure 31.2). The relationships between the granitic and granodioritic gneisses at Puzzle Lake is similar to that preserved at lower metamorphic grade between the Mitten granite dike (Bright 1986) and the Mellon Lake granodiorite-tonalite-migmatite complex in the adjacent Mellon Lake area. In both the Mellon Lake and Puzzle Lake areas, thin (100 to 200 m) zones of amphibolite are present typically between the plutonic bodies and the surrounding paragneisses. Complex folding patterns were observed in these gneisses south of Gull Lake, however, insufficient traverses were completed to fully unravel the structural style present in this domain.

Between the Puzzle Lake pluton and the RLMZ, is an east-dipping zone of amphibolite, marble and siliciclastic metasediments, which become more strongly mylonitized and altered toward the east, approaching the RLMZ. These rocks may be preserved at slightly lower metamorphic grade than rocks west of Puzzle Lake, based mainly on the reduced presence of leucosome within the mafic and felsic gneisses present within this area.

**Sharbot Lake Domain**

Three main rock packages are present within the Sharbot Lake domain in the map area, namely:

1. a greenstone-granodiorite package, present mainly adjacent the RLMZ
2. a marble-syenite package, present in the Fifth Depot to Chippegoo lakes area
3. the Hinchinbrooke pluton.

The greenstone-granodiorite package is best exposed in the south-central part of the area along the Tamworth–Parham road (see Figure 31.2). It consists of approximately 750 to 1000 m of amphibolite and metagabbro, overlying by 350 to 500 m of mafic, intermediate and felsic metavolcanic rocks, including felsic pyroclastic rocks, overlain by 150 to 200 m of metawacke and rusty schist, in turn overlain by an unknown thickness of dolomitic and calcitic marbles. Although these rocks are strongly foliated and primary textures are only poorly preserved, the sequence is similar to that observed in the Darling area of Sharbot Lake domain (Easton 1988a). Structural trends are dominantly northwesterly and northeast-dipping, and several large-scale folds are present (see Figure 31.2). Rocks of this sequence are intruded by both medium-grained granodioritic and monzogranitic rocks. In fact, much of the central part of the map area consists of a complex zone of supracrustal remnants and plutonic rocks. Near Cranberry Lake, the sequence is cut by a coarse-grained syenogranite, the Cranberry Lake pluton, which has an unusual map pattern (see Figure 31.2) in that it seems to both cross-cut regional folds in the supracrustal rocks, yet also appears to be folded. It is correlated with the 1090 to 1070 Ma Skootamatta-Kensington suite, as discussed, along with its tectonic significance, in the section on the RLMZ.

The marble-syenite package underlies the Fifth Depot–Slave–Chippegoo lakes area. It consists of coarse-grained to very coarse-grained calcitic and dolomitic marbles intruded by a variety of syenite, monzonite and granitic plutonic rocks, which texturally, compositionally and geochemically (Wallach 1973) resemble members of the lithologically and chemically similar 1180 to 1155 Ma Ganaouque and 1090 to 1070 Ma Skootamatta-Kensington suites. Without further work, it is not possible to clearly assign these plutonic rocks to either suite. Amphibolite and siliciclastic units are locally present within the marble sequence. In the north, marble and granitoid rocks are separable at 1:50 000 scale, however, southeast of Slave Lake, the plutonic and marble units are intimately mixed, making separation of the 2 rock types difficult except at extremely detailed map scales (see Figure 31.2, interlayered syenite, monzonite and marble unit). The Enterprise complex of Wallach (1973), renamed the Chippegwa pluton by Davidson and Ketchum (1993), is part of this rock package. Chemical analyses of the Enterprise complex from Wallach (1973) indicate the presence of at least 2 discrete compositional phases, a monzodiorite to monzonite phase and a syenogranite phase. Insufficient traverses were conducted to separate individual plutons within this package, or to precisely define the contact between these plutonic rocks and the older granitic-granodioritic rocks present to the west. The relationships between marbles in this package and those associated with the greenstone-granodiorite package is unclear. Marbles in the marble-syenite package could simply be stratigraphically higher equivalents or, alternatively, they could represent a separate carbonate sequence, either in unconformable or tectonic contact with the greenstone-gra-
nodiorite package. Regardless, they appear to have been an effective trap in localizing emplacement of the younger plutonic rocks.

The Hinchinbrooke pluton is a tonalite-granodiorite body, which has been subjected to granite-facies metamorphism. In the Fifth Depot Lake area, a 200 to 400 m wide sliver of thinly layered, mafic and intermediate composition gneisses separate the body from the marble-syenite sequence. This gneissic rim, including amphibolites, resembles the aforementioned marginal phases exposed further west surrounding granodioritic-tonalitic plutons within Mazinaw domain. These gneisses are cut by metamorphosed, northeasting mafic dikes which have sharp contacts and preserved chill margins. These dikes were not observed within the marble-syenite or the greenstone-granodiorite sequence. Bright (1986) reported mafic dikes cutting the Mellon Lake complex, and it is possible these dikes are of the same suite.

Robertson Lake Mylonite Zone

Wolff (1982) mapped the RLMZ from Mountain Grove southeastward to just north of Puzzle Lake (see Figure 31.2, west branch of RLMZ). In contrast to Wolff's (1982) position, Davidson and Ketchum (1993) show the RLMZ paralleling Shibsgau creek, roughly concurrent with the position of the main branch of the RLMZ shown in Figure 31.2. The presence of several prominent linear features in the Mellon Lake–Puzzle Lake area, led Easton (1988b) to suggest that the RLMZ might be present as several splays, rather than a single structure.

Determination of the location of the RLMZ in the area utilizes the key features of the RLMZ observed further north, namely:

1. It is a topographically significant feature, serving to localize glacial deposits and subsequent drainage patterns.
2. It separates higher metamorphic grade rocks to the west from lower grade rocks to the east.
3. It is a shallow-dipping structure.
4. Rocks along the zone are extensively mylonitized or cataclastic for widths of several hundred metres.
5. There is abundant development of hematite and epidote (or chlorite at lower grade).
6. Adjacent structural trends are transposed into parallelism with the RLMZ trend.

All of these features are associated with the main branch of the RLMZ, shown in Figure 31.2. The line shown on the map traces the 30 to 60 m high escarpment that follows the trace of the RLMZ in the area. Mylonitic rocks are present at the base of this scarp, dipping shallowly (15 to 20°) to the east. Rocks at the top of the scarp are strongly hematized, epidotized and cut by quartz veins. Quartz stockworks are locally present. In addition, dominantly northwesterly structural trends within the greenstone-granodiorite package in the southern part of the map area are transposed into north-trending parallelism with the RLMZ (see Figure 31.2). Metamorphic grade of the RLMZ appears to be lower amphibolite facies, in contrast to the middle greenschist facies along the RLMZ in the Darling area (Easton 1988a).

A splay of the RLMZ does strike from the Long Lake map area through Puzzle Lake (see Figure 31.2). However, apart from the development of a narrow (100 m²) fine-grained, black to grey, marble mylonite breccia zone exposed on a few islands and peninsulas on the north end of Puzzle Lake, the RLMZ has little effect on the adjacent orthogneisses present on either shore of Puzzle Lake. Similar breccia is preserved on strike in the Long Lake area to the northeast, within the area mapped by Wolff (1982) as part of the RLMZ. South of Puzzle Lake, this splay of the RLMZ is not easily traced. Poorly exposed gritty carbonates south of Little Gull Lake may represent the southward continuation of this splay.

The Cranberry Lake pluton is altered by the RLMZ, however, plutonic rocks of this pluton are not strongly mylonitized, and locally appear to inject more strongly mylonitized rocks. In addition, as shown in Figure 31.2, the western edge of the pluton appears to be injected parallel to, or transposed into parallelism with the RLMZ. Wolff (1982) observed similar relationships between the 1073 Ma (Davidson and Easton 1994) McLean pluton and the RLMZ. If this is the case, it suggests that the RLMZ may have been active prior to 1073 Ma, even though further north it juxtaposes rocks metamorphosed as young as 1020 Ma against cooler rocks. For this reason, it is suggested that the Cranberry Lake pluton is part of the Skootamatta-Kensington suite, rather than the Gananoque suite.

ECONOMIC GEOLOGY

Overview

Figure 31.3 shows the location of previously recorded mineral occurrences in the area (Hewitt 1964; Maleczak et al. 1985), as well as additional occurrences noted in the assessment files. In addition, soil geochemical anomalies reported in the assessment files are also shown. Figure 31.3 also provides a provisional interpretation of the mineral potential of the area, based on field observations and the distribution of occurrences. Areas of lower mineral potential in the area include the high-grade rocks of Mazinaw domain, and the area underlain by the Hinchinbrooke pluton. Note that the area north of Norway Lake is difficult to access, consequently, the geology of this area could not be adequately evaluated, even though it is likely underlain by both carbonate and metavolcanic rocks.

Moderate to High Potential for Zinc, Molybdenum and Wollastonite

This region corresponds to the area underlain by the marble-syenite package and the marble belt west of the main branch of the RLMZ. The rating of the potential of this package is based on the following reasons:

1. The presence of known zinc mineralization in a 5 km long belt in the Slave Lake area (e.g., Table 31.1, sample 98RMF–0242)
2. the presence of zinc soil anomalies in the Fifth Depot Lake area (Brack 1993)

3. the presence of more marble in the area than shown on previous maps

4. the occurrence of dolomite marble throughout part of this belt and, in the Fifth Depot Lake area, the occurrence of extremely coarse-grained calcitic marbles, similar to those present at the Long Lake zinc mine (Wolff 1982)

5. Easton (1995), on geochemical grounds, suggested that contact aureoles of pluvols of the 1090 to 1070 Ma Skootamatta-Kensington suite were targets for zinc and wollastonite mineralization that formed predominantly in a manto-type setting (e.g., Nelson 1991), although contact skarn deposits may also be present. As noted above, pluvols in this belt are either 1180 to 1160 or 1090 to 1070 Ma in age, or both. The abundance of molybdenite showings associated with these pluvols also suggests interaction between the pluvols and the marbles. In addition, the Chisholm mine is a base metal deposit (pyrite, pyrrhotite) with associated molybdenite, not a molybdenite vein or skarn, consistent with formation as a manto deposit (cf. Nelson 1991). The Chisholm mine is spatially associated with monzogranite of the Chippewa pluvol. Grab samples from the Chisholm mine (see Table 31.1) show more consistent enrichment in copper than in molybdenum.

6. Associated mineralization that might be expected along with zinc, includes wollastonite, molybdenite, tin and fluorite. Tin and fluorite are commonly associated with manto deposits (Nelson 1991), and geochemical comparisons with younger skarn systems (cf. Ray and Webster 1991) suggest that the syenogranitic phases of the Chippewa pluton may host tin in addition to molybdenum mineralization. According to Wallach's (1973) limited chemical data on this body, the syenogranite phase occurs mainly along the eastern margin of the body, south of Chippego Lake. Based on the available geochemical data and field observations, zinc mineralization appears to be associated with monzodioritic to monzonic intrusive phases, whereas known molybdenite occurrences are associated with monzo- to syenogranitic phases.

Note that the area south of Norway Lake is only ranked high potential for zinc, based on previous exploration. This area lacks the associated syenite plutons, and their related mineralizing factors.

**Low to Moderate Potential for Gold and Copper Along the Robertson Lake Mylonite Zone**

Based on the occurrence of gold and tetrahedrite occurrences along the RLMZ to the north, this area might be expected to host similar mineralization. The presence of many gossan zones, quartz veins and quartz stockworks in the mylonitic rocks in the RLMZ, and along the escarp-
Low to Moderate Potential for Massive Sulphides

This area outlines the middle amphibolite greenstone-granodiorite package, which includes both mafic and felsic metavolcanics and rusty schists and gossan zones capping the metavolcanic sequence. Ongoing geochemical characterization of these metavolcanics should help elucidate the true potential of this sequence.

Areas of Abundant Veining

Large sections along the Tamworth–Parham road contain pyrite-tourmaline-quartz veins, which occur in all rock units of the greenstone-granodiorite package. Extensive bleaching of the host rocks to these veins is present. This type of alteration has been observed by the author previously in the Mazinaw Lake area in association with gold occurrences in that region (Easton and Ford 1991). At present, it is unclear if the concentration of these veins along the road simply reflects improved exposure, or whether there is some other factor serving to localize veins in this area. Several assays of these veins are underway.

REFERENCES


32. Project Unit 96–098. Geological Setting and Origin of Offset Dikes of the Sudbury Structure, and Allied Hypervelocity Impact Studies


Impact Geology Group, Department of Geology, University of New Brunswick, Fredericton, N.B.

INTRODUCTION

The Impact Geology Group at the University of New Brunswick commenced studies at Sudbury in 1991. The group now comprises a team of eight: Dr John Spray, Post-Doctoral Fellow Dr James Whitehead, four MSc students (John O’Connor, Alain Murphy, Graham Nickerson and Christina Wood) and two PhD students (Heather Gibson and Ron Scott). The primary focus of the Sudbury work is to establish the geological setting of the radial and concentric dike systems that emanate from the main mass of the Sudbury Igneous Complex (SIC). Many of these dikes have in the past been, or are currently, mined for their Cu-Ni-PGE deposits (e.g., Whistle, Copper Cliff, Foy, Worthington, Frood–Stobie). They, thus, represent an important economic asset as a provincial and national resource. The group is currently working on four offset dikes: Frood–Stobie (South Range Breccia Belt), Hess, Manchester and Whistle–Parkin. Consideration is also being given to the metamorphic history of the Sudbury region, with particular reference to distinguishing pre- and post-impact tectonothermal events (Thompson et al. 1998).

FROOD–STOBIE AND THE SOUTH RANGE BRECCIA BELT

The Frood–Stobie subconcentric offset dike occurrence is anomalous with regard to most of the other Sudbury Offset dikes. It is dominated by Sudbury Breccia (pseudotachylite) rather than quartz diorite dike material. Although quartz diorite does occur at Frood–Stobie, it is typically present as isolated pods and not dike-like sheets and quartz diorite is not found to be continuous for any distance. It is important to realize that Frood–Stobie is part of a much larger breccia belt that forms a subconcentric arc of pseudotachylite (Figure 32.1), ranging from 0.1 to over 1 km in width, and reaching approximately 45 km in length (Scott et al. 1996). We have termed the whole of this arc the South Range Breccia Belt. The South Range Breccia Belt abuts the Sudbury Igneous Complex in the vicinity of the Kirkwood Mine in the east (Garson Township), and at the abandoned Victoria Mine in the west (Denison Township). It also connects to the Sudbury Igneous Complex via an embayment or funnel structure in the Little Stobie and Mount Nickel Mines area (Bleazard Township). The belt is cut by the Copper Cliff radial offset dike. Exposure is relatively good to the east of the Copper Cliff offset dike (where extensive sampling has been carried out for magnetic and petrochemical work as part of our studies), but is poor to the west of it. Outcrop in Graham and Waters townships is typically limited to boulders in low-lying, marshy terrain, as originally noted by Speers (1957). However, the full extent of the belt is well expressed as a 120 nT high in regional aeromagnetic data. The reason for the high is not yet clear, but it is probably due to the paramagnetic susceptibilities of ferromagnesian silicates that are both concentrated and aligned within the pseudotachylite matrix.

The South Range Breccia Belt hosts the world’s largest Ni-Cu-PGE deposit, the Frood–Stobie mine complex, which has been in production for more than a century. Deposits at the Kirkwood, Little Stobie and Victoria mines have also contributed economic significance to the belt. The Ni-Cu ore is commonly associated with quartz diorite bodies that are, in turn, hosted by the pseudotachylite. The ores are primarily composed of massive to disseminated chalcopyrite, pyrrhotite and pentlandite, with minor associations of numerous rare PGE phases (e.g., froodite; PJB2).

The belt consists of subrounded to rounded, clasts and large blocks of metasedimentary and metavolcanic rocks of the 2.49 to 2.2 Ga Huronian Supergroup (Debicki 1990). These inclusions range from several millimetres to tens of metres in size and they are set within a dark, fine-grained crystalline matrix. The inclusions and matrix together comprise Sudbury Breccia, also known as Frood Breccia, which we now know is a form of pseudotachylite (Dressler 1984; Thompson and Spray 1996). The matrix comprises quartz + biotite + plagioclase + potassium-feldspar + ilmenite, with minor pyrite and pyrrhotite. Zircon (predominantly inherited from the wall rocks) is ubiquitous and easily detected via plagioclinic haloes in biotite. The matrix of the pseudotachylite is microcrystalline, with no melt glasses being detected (if ever present). The matrix is locally massive, but is more typically found strongly foliated as defined by biotite. Magnetic fabric studies over the last year indicate that the biotite foliation represents a primary flow structure, and that it is not a metamorphic overprint formed during the Penokean (1.9 to 1.7 Ga) and/or Grenvillian (~1 Ga) orogenies (Scott and Spray, in press). This has interesting implications for the metamorphic history of the South Range. Previous workers have favoured amphibolite facies grade being realized during
the later stages of the Penokean orogeny, following the impact event (Thomson et al. 1985). Our work suggests that the metamorphic grade in the South Range did not exceed greenschist facies grade following the 1.85 Ga impact event (Scott and Spray, in press). Grades higher than this would have reset the fabric of the biotites, and there is no evidence that this occurred. Those amphibolite facies grade assemblages that do occur in the area were

Figure 32.1. Location of offset dikes.
probably formed during pre-impact (2.4 to 2.2 Ga) Bleazardian tectonometamorphism (Riller and Schwertner 1997), or the earlier (pre-impact) stages of the Penokean. Geochronological studies are currently underway to test this. Critically, the pseudotachylite matrix flow fabrics established for the Frood–Stobie ore region indicate that the sulphides were injected subvertically. So far, no preferred sense of direction can be determined, but an upward flow regime is favoured.

The Frood–Stobie pseudotachylite belt was derived by the fractional comminution and melting of wall rocks, most probably during large displacement seismogenic faulting (superfaulting, Spray 1997). Such fault behaviour may be the result of gravity-induced transient cavity collapse, which leads to the slumping of the crater walls into the impact melt sheet.

**HESS OFFSET DIKE**

The Hess offset dike occurs as a subconcentric dike, up to 60 m wide (typically 10 to 20 m), located 12 to 15 km north of the base of the Sudbury Igneous Complex within predominantly Archean granitoids (Wood and Spray 1998). The Hess offset dike comprises quartz diorite, allied to the magmatic rocks of the Sudbury Igneous Complex, and is locally mineralized. The Hess body is mineralogically distinct from the more typical quartz diorite of the Sudbury Camp—it is generally less quartz-rich than South Range quartz diorite. The Hess offset dike extends from at least as far as Clear (formerly Hess) Lake in the west (just east of Cartier), to east of the radially-oriented Foy offset dike, for a total strike distance of approximately 23 km. The dike is commonly found in association with Sudbury Breccia (pseudotachylite). The nature of displacement on the post-impact-reactivated, north-northwest-trending, Fecunis Lake and Sandcherry (Onaping) faults indicates that Hess offset dike dips steeply to the south.

Previous work in the North Range suggested that the Hess offset dike occurs at the northern margin of a pseudotachylite-rich annulus surrounding the Sudbury Igneous Complex (Thompson and Spray 1994, 1996). Critically, this implied that Hess could "line" the annulus margin and hence be more extensive than originally mapped (Card and Innes 1981). This project therefore sought to map and trace out the full extent of the Hess offset dike (Wood 1998), the result being that the Hess offset dike has been extended approximately 8 km west of its previously determined limit (Card and Innes 1981). The common association of the Hess offset dike with Sudbury Breccia (pseudotachylite) indicates that the Hess offset dike does indeed delineate a subconcentric fault system that defines either 1) part of the collapsed and modified northern margin of the transient cavity of the Sudbury impact structure, or 2) the northern margin of the peak ring. The pseudotachylite occurs as discrete veins (up to 0.5 m wide and as millimetre to microscopic veinlets), allied to broader zones of more cataclastic deformation. The larger pseudotachylite veins are typically subparallel to the Hess offset dike, whereas the finer veinlets are more anastomosing.

Major, trace and rare earth element chemistry of selected Hess offset dike samples clearly indicates an affinity with the Sudbury Igneous Complex. The primary mineralogy is quartz + plagioclase (labradorite rimmed with oligoclase) + pyroxene (usually found altered to actinolite) + hornblende + biotite.Opaque phases include pyrite, pyrrhotite, pentlandite and chalcopyrite, with minor amounts of argentiferous pentlandite (FeNiAgS8), michenerite (PbBiTe) and hessite (Ag2Te). Secondary (deuteritic) phases include titanite, epidote, chlorite and actinolite. The magmatic source of the Hess offset dike is unclear: it may be fed laterally from the proximal Foy offset dike (see Figure 32.1), which is directly connected to the Sudbury Igneous Complex, or it may be fed from underneath via a listric fault system which connects at depth to the lower levels of the Sudbury Igneous Complex. Field work indicates that the distal Foy offset dike is in turn fed by the Hess offset dike (see Figure 32.1). The Hess offset dike is of interest because it is the most distal concentric offset dike known. As such, studying it can throw light on the nature of impact-related fracturing and faulting in target rocks well beyond the Sudbury Igneous Complex.

Work on the Hess offset dike was completed in September 1998, with the successful defence of Wood's MSc thesis (Wood 1998). Future work on the Hess–Foy offset dike system will focus on investigating footwall breccia inclusions that occur within the quartz diorite.

**MANCHESTER OFFSET DIKE**

The Manchester offset dike is located 4 to 5 km southeast of the Sudbury Igneous Complex within Huronian metasedimentary footwall rocks. It was originally mapped as a basic to intermediate dike-like body in the 1950s (Thompson 1959), and recognized as a Sudbury Igneous Complex-related offset dike in the 1970s (Grant and Bile 1984). The dike is subconcentric to the Sudbury Igneous Complex and is up to 30 m wide. It strikes discontinuously for at least 10 km and dips about 60° degrees to the southeast. A thicker, better exposed zone, approximately 5 km in length, is centred within the 10 km strip. The Manchester offset dike is hosted by a broad zone of Sudbury Breccia (pseudotachylite), the southern limit of which is, in part, defined by the so-called Falcon Fault, which was either post-impact generated or reactivated. The host pseudotachylite zone is up to 350 m thick. Contacts between the pseudotachylite and the dike are generally sharp, with the dike being slightly chilled against the pseudotachylite. The dike comprises the assemblage quartz + plagioclase + alkali feldspar + amphibole + biotite. Granophyric and myrmekitic intergrowths are particularly common. Clino- pyroxene relics also occur. The dike is a quartz diorite and, although clearly genetically related to the Sudbury Igneous Complex, is more siliceous than many of the other offset dikes of the Sudbury structure. This may reflect the effects of assimilating the predominantly quartzo-feldspathic (Huronian) wall rocks.

Critical to this study is the original mode of connection between the Manchester offset dike and the Sudbury Igneous Complex, as there is no apparent physical
attachment between the two at present exposure levels (O’Connor and Spray 1997). Penokean faulting has dismembered the Manchester offset dike and field studies cannot yet constrain the magma emplacement direction (i.e., whether lateral or vertical). If a lateral connection to a radial offset dike cannot be found, then the implication is that the Manchester offset dike was emplaced from below, or from above, via Sudbury Igneous Complex-concentric listric faulting initiated during collapse of the transient cavity some time after hypervelocity impact.

This last (1998) field season has revealed that numerous faults, related to the South Range Shear Zone, pass through the Manchester offset dike and effectively dismember it. This could account for the limited strike length as currently exposed. These faults appear to have had complicated histories, showing evidence for both strike-slip and normal or thrust displacements occurring before, during and after impact. Some were probably reactivated during the Grenville period. The faults can show ductile deformation, in which case they are more correctly defined as shear zones. This implies that the fault-shear systems were active at depths of 10 to 15 km, within the brittle-ductile transition. In addition, several new outcrops of quartz diorite were found during the 1998 season, thus extending the known length of this poorly understood offset dike.

WHISTLE–PARKIN OFFSET DIKE

The Whistle–Parkin offset dike consists of a pair of radial offset dikes, located to the northeast of the Sudbury Igneous Complex, west and northwest of Lake Wanapitei. The Whistle offset dike is the proximal dike, which connects directly to the Sudbury Igneous Complex via a sublayer embayment. The relationship between the Whistle and Parkin offset dikes is a subject of contention and is one of the focal points of the project. The Parkin offset dike is apparently displaced west of the Whistle offset dike via a fault zone. However, there is some evidence that Parkin offset dike may be a separate radial dike and is not a displaced extension of Whistle offset dike. As with all other offset dikes, Whistle–Parkin occupies a pseudotachylyte-lined fracture-fault system, which was probably formed during the contact and compression stages of the impact event. The project started this year (1998) and is currently the focus of an MSc thesis, which may be upgraded to a PhD within the next year.

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33. Project Unit 97–012. Cu-Ni-PGE Potential of the Nipissing Diabase

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INTRODUCTION

This project is part of a PhD thesis at the University of Western Ontario, under the supervision of Dr. Neil MacRae (University of Western Ontario) and Dr. Reid Keays (Laurentian University). The purpose of the thesis is to compare and contrast mineralized versus nonmineralized portions of Nipissing Diabase (gabbro) to see if it is possible to develop geochemical criteria that might be useful in exploration, as well as to provide insights into the controls over mineralization in these systems. A major reason for determining if the parental magmas to the Nipissing Diabase were S-undersaturated is to establish whether or not these magmas had the potential to form major copper (Cu), nickel (Ni) and platinum group element (PGE) sulphide deposits (cf. Keays 1995). Preliminary results from bedrock mapping and regional (Thessalon to southwest of Lake Temagami) and detailed sampling completed during the field seasons of 1997 and 1998 (Jobin–Bevans, MacRae and Keays 1997) are discussed below.

GENERAL GEOLOGY

The Nipissing Magmatic Province (Lightfoot and Naldrett 1996) outcrops within the Proterozoic Huronian supracrustal belt, has an east-west extent of almost 450 km and a north-south breadth of about 350 km. More than 20% of this area is covered by mafic (dominantly gabbroic), tholeiitic, intrusive rocks referred to collectively as Nipissing Diabase. Between Cobalt and Sault Ste. Marie, intrusions of Nipissing Diabase form undulating sills, cone sheets or lopoliths and dikes. Most Nipissing intrusions show very little differentiation and consist almost entirely of quartz diabase, whereas some of the intrusions are highly differentiated, generally ranging upward in composition from basal quartz diabase, gabbronorite (hypersthene gabbro), vari-textured gabbro, granophyric gabbro, aplitic, granodioritic to granitic rocks and a capping quartz diabase (Lightfoot, de Souza and Doherty 1991). Olivine has been reported from only a few intrusions where it occurs in the basal quartz diabase and hypersthene gabbro units (Lightfoot et al. 1986).

Nipissing Diabase represents a major intrusive event between 2206 and 2223 Ma (Corfu and Andrews 1986; Noble and Lightfoot 1992) and is probably the remnant of an eroded Continental Flood Basalt system (e.g., Lightfoot et al. 1987; Lightfoot, de Souza and Doherty 1991). The possibility that Nipissing Diabase was associated with intraplate volcanism, rifting and basin development, is suggestive of an environment favorable for the formation of significant concentrations of magmatic sulphide minerals rich in Cu-Ni-PGE (Lightfoot et al. 1987). Moreover, there are several geochemical and structural similarities between Nipissing Diabase and the sills and intrusions that host the prolific Cu-Ni-PGE deposits in Noril’sk, Russia (Lightfoot and Naldrett 1996), and the large, basin-related Cu-Ni-PGE deposits found in the Insizwa Complex in southern Africa (Lightfoot, Naldrett and Hawkesworth 1984).

FIELD CHARACTERISTICS

The majority of Nipissing Diabase occurs as undulating sills, consisting of basins and arches, and dikes (Hrskievich 1968; Jambor 1971). In this form, massive and disseminated sulphide mineralization is concentrated within the basin or limb (between arches and basins) portions of the sills (Lightfoot and Naldrett 1996); pods of massive sulphide (pyrrhotite + pentlandite > chalcopyrite) also develop within the arches.

In several locations, elliptical surface outcrop expressions of Nipissing Diabase and structural features of surrounding sedimentary rocks suggest inward-dipping or lopolithic forms. Jambor (1971) and Lovell and Caine (1970) interpreted large arcuate to circular outcrop patterns in the Cobalt–Gowganda area as patterns derived by erosion of an undulating sheet or sheets, or cone-shaped intrusions. Results of this year’s mapping suggest that this same model may be applicable to the Nipissing Diabase that outcrops between Lake Wanapitei (Rathbun Township) and the Sturgeon River (Janes and McNish townships). In this region, circular outcrop patterns suggest cone-in-cone intrusion patterns. In cone-shaped intrusions, immiscible sulphide liquids would have settled toward the base of the upper sill-like portions of the cones (e.g., Kukagami Lake intrusion, Kelly Township, and Rathbun Lake intrusion, Rathbun Township) and within the lower conduits or feeder system of the intrusions (e.g., Chiniguichi River intrusion, Janes Township). In addition, physical traps for sulphide mineralization would likely have developed within the irregular portions or embayments of the feeder system.
CURRENT RESEARCH

A major part of the 1998 field season was spent conducting detailed lithochemical traverses across several sills of Nipissing Diabase and examining sulphide occurrences in areas southwest and northeast of Sudbury. Approximately 135 samples were collected and submitted for analysis.

A total of 6 traverses, chosen according to documented mineralized occurrences, were completed. The Basswood Lake intrusion is a relatively large and unminalized body of Nipissing Diabase located north of Thessalon on Highway 129 and within Bridgland, Wells, Kirkwood and Day townships. This traverse will augment previous work by Lightfoot et al. (1986, 1987) by examining the variations in S, Se and PGE across the intrusion.

A northwest-trending traverse was completed through the central part of the Chiniugu River intrusion (Janes Township), located about 51 km northeast of Sudbury, east of Murray Lake and southwest of the Chiniugu River. The intrusion has an irregular shape with a lopolithic or stock-like form and hosts several significant Cu-Ni-PGE sulphide occurrences.

Additional locations of traverses include the Charlton Lake intrusion, an east-trending sill exposed along the north shore of Charlton Lake (Curtin Township), the Bell Lake area, located across a northeast-trending sill, north of Bell Lake (Lorne Township), and two north-trending traverses across the eastern and western extents of an east-trending sill (Kukagami Lake sill) in the northern part of Kelly Township, about 45 km northeast of Sudbury.

In addition to the traverses, a detailed mapping and sampling project was completed on the Rauhala occurrence, located north of Makada Lake in Waters Township, southwest of Sudbury. Included in the sample set from this property is diamond-drill core that intersected a Cu-Ni-PGE sulphide occurrence in quartz-bearing gabbro (Nipissing Diabase).

GEOCHEMISTRY

During the 1997 field season, more than 80 rock samples were collected and analyzed from across the Nipissing Magmatic Province; Clement Township (east-central area), Chiniugu River and Kukagami Lake intrusions (central area), Louise and Waters townships (west-central area) and the Basswood Lake intrusion (western area).

Major Elements

Eighty samples of medium-grained gabbro from the 1997 sample suites have major oxide ranges of 45.01 to 56.84 weight % SiO₂, 0.34 to 3.54 weight % TiO₂ and a magnesium number of 31 to 74. The highest concentrations of SiO₂ (53.05 to 56.84 weight %) were recorded in samples from Clement, Louise and Waters townships, and the Basswood Lake intrusion. The highest concentrations of TiO₂ (2.51 to 3.54 weight %) and the lowest average magnesiu numbers (Mg# 44-58) were recorded in samples from Clement and Waters townships and the Basswood Lake intrusion. Conversely, the lowest concentrations of SiO₂ and TiO₂ were recorded in rocks from the Chiniugu River intrusion and the Kukagami Lake sill and the highest average magnesium numbers (Mg# 64 to 70) were from the Kukagami Lake sill, Chiniugu River intrusion and Louise Township.

Trace Elements

Primitive mantle-normalized (Sun and McDonough 1989) spider diagrams for the six sample sets all show moderate light rare earth element (LREE) and middle rare earth element (MREE) enrichment that is about 5 to 50 times primitive mantle. LREE and MREE enrichment is even more pronounced in samples of coarse-grained to pegmatic gabbro. Similar results are reported by Lightfoot and Naldrett (1996). The trends are typical of fractionated tholeiitic basalts in that they are enriched in the most incompatible elements. In addition, all of the sample suites have negative niobium anomalies and have relatively restricted La/Sm ratios ranging from 2.25 to 4.32—features that are typical of a magma that has interacted with a crustal reservoir or of young rocks derived from ancient regions of mantle lithosphere that contain recycled continental crust (Lightfoot, de Souza and Doherty 1993).

Copper, Nickel and Platinum Group Elements

Platinum group element values for this project were determined using the Ni-sulphide fusion technique and inductively coupled plasma mass spectrometry (ICP-MS) (R.R. Keays, Laurentian University, personal communication, 1998). Sulphide occurrences that have been identified and sampled, to date, can be classified into three main types, based on their stratigraphic position, structural setting and ratio of copper, nickel and platinum group elements: 1) massive sulphide mineralization in upper parts of the stratigraphic succession, with a high nickel:copper ratio and very low platinum group element contents; 2) disseminated sulphide mineralization in lower parts of the stratigraphic succession, with a low nickel:copper ratio and anomalous platinum group element contents; and 3) contact-related disseminated to semimassive sulphide mineralization in lower parts of the stratigraphic succession, with a low nickel:copper ratio and highly anomalous platinum group element contents.

Massive (>80% total sulphide) sulphide mineralization is noted from both the Waters Township and the Louise Township localities. Mineralization occurs primarily in medium-grained gabbro where it forms large massive sulphide pods comprising pyrrhotite > pentlandite > chalcopyrite. Although it is difficult to be sure of their exact location within the stratigraphic succession, most of the pods appear to be located either in the arch or limb portions of undulating sills (cf. Lightfoot and Naldrett 1996, Figure 1b). The sulphide pods are typically surrounded by several metres of medium-grained gabbro that contains 2 to 5% blue quartz phenocrysts (blue quartz eyes) and 2 to 5% disseminated and bleb sulphide. The blue
quartz eyes, coupled with the presence of actinolite and feldspar textures, suggest hydrothermal alteration of the rocks. In this project, we have determined maximum abundances of 50 ppb Pd, 27 ppb Pt, 63 ppb Au, 0.10% Cu and 0.39% Ni from Louise Township and 15 ppb Pd, 10 ppb Pt, 289 ppb Au, 0.30% Cu and 0.45% Ni from Waters Township.

Disseminated (2 to 5% total sulphide) sulphide mineralization occurs within the lower part of the stratigraphic succession in the Kukagami Lake sill. Disseminated and bleft sulphide mineralization (chalcopryite > pyrrhotite > pentlandite) occurs within massive hypersthene gabbro at greater than 50 m above the basal quartz diabase unit and the base of the intrusion. We have determined maximum abundances of 1930 ppb Pd, 440 ppb Pt, 120 ppb Au, 0.63% Cu and 0.28% Ni. Lightfoot, de Souza and Doherty (1991) reported similar findings with maximum abundances of 4160 ppb Pd, 1100 ppb Pt, 600 ppb Au, 1.10% Cu and 0.39% Ni.

Contact-related disseminated (2 to 10% total sulphide) to semimassive (>25% total sulphide) sulphide mineralization occurs within the lower part of the stratigraphic succession and generally within 100 m of the footwall contact. This type of mineralization has thus far only been recognized in the Chingiuchi River intrusion where it is generally confined to the lower hypersthene gabbro and gabbronorite units. Sulphide mineralization is both net-textured and blybby and consists of chalcoprite > pyrrhotite > pentlandite. Surface exposures of semimassive mineralization occur as small (<1.0 m diameter) pods rich in chalcopyrite, pyrrhotite and pentlandite. Distribution of these pods may be in part structural, whereby disseminated sulphides have become concentrated along localized fractures and/or faults and/or joint planes. We have determined maximum concentrations of 7393 ppb Pd, 1300 ppb Pt, 720 ppb Au, 1.71% Cu and 1.14% Ni.

Sulphur and Selenium

Sulphur/selenium ratios are useful to determine whether or not sulphide metals are of magmatic origin or are the result of remobilization through processes such as weathering and alteration and whether or not the S in the system has been externally derived (i.e., a sedimentary rock source) or is entirely magmatic in origin (Eckstrand and Hulbert 1987; Eckstrand et al. 1989).

The majority of samples from the 6 sample suites have S/Se ratios between 1000 and 7300, which is well within the range of uncontaminated magmatic sulphides (1000 to 10 000; Goldschmidt 1954); primitive magmas have a S/Se ratio of about 2700 (Buchanan 1988). Seven samples have S/Se ratios that are less than 1000, which is indicative of sulphur loss due to small-scale redistribution of the chalcophile elements during weathering or metamorphism (Reeves and Keays 1995).

COMMENTS

Preliminary field and geochemical evidence suggest that the magmas that formed the Nipissing Diabase were indeed capable of forming significant Cu-Ni-PGE sulphide deposits. Sulphide mineralization appears to be controlled by magma chamber processes and/or structural features such as the proximity to footwall contact and localized jointing and/or faulting. Field and geochemical evidence suggest that the criteria for mineral exploration in Nipissing Diabase (cf. Lightfoot and Naldrett 1996, Table 1) are a reliable means of target assessment at the outset of a mineral exploration program. Still at issue is whether or not the magmas were S-undersaturated prior to emplacement and whether or not an external sulphur source was required to achieve sulphur saturation of the magma.

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34. Project Unit 93-12. Structural Patterns and Kimberlite Emplacement

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BACKGROUND

The kimberlite project is designed to characterize known kimberlite pipes province-wide, provide a summary of the geology, provide a limited amount of microprobe data on indicator minerals, outline the exploration history of each pipe and develop suggestions for kimberlite exploration within Ontario. This article deals mainly with regional structural patterns and kimberlite emplacement.

GENERAL PATTERN

Kimberlite clusters within Ontario occur mainly in the northern portion of the province and establish a broad regional trend oriented at 325° (Figure 34.1). This trend can be extended southeast to kimberlite dikes near Belleville, Ontario, and beyond into northern New York State. This northwest trend overprints a variety of geological structures and regions. The trend is subparallel to regional-scale swarms of Late Archean to Early
Proterozoic diabase dikes. Within Ontario, the regional trend of kimberlite intrusion is roughly orthogonal to the Kapuskasing Structural Zone, the Grenville Deformation Zone, Cape Henrietta Marie Arch and the Fraserdale Arch. Whereas the regional pattern defines a northwest trend, individual pipes or clusters are spatially associated with more local faults, linears and lithologic contacts oblique to the northwest trend.

**COBALT–NEW LISKEARD AREA**

Kimberlites occur on both flanks of the Lake Timiskaming Structural Zone in this region (Figure 34.2). The Guigee pipe in Quebec occurs immediately west of the Quinze Dam Fault and just east of the Hudson Bay Paleolineament (Kutina and Fabbri 1972), close to the contact between the Baby metavolcanic rocks (south) and Pontiac metasedimentary rocks (north). Lineament trends intersect at or close to the site of emplacement. The Notre-Dame-Du-Nord (NDN) pipes at the north end of Lake Timiskaming have been emplaced in a wedge-shaped block bounded on the west by the Blanche River Fault and the Hudson Bay Paleolineament on the east (Kutina and Fabbri 1972) (see Figure 34.2).

Near Cobalt and New Liskeard, numerous kimberlite pipes occur where more conspicuous northwest-trending faults are intersected by local northeast-trending cross faults. Mapping by Thomson (1956, 1960) and Russell (1984) suggests that the bedrock in this region is broken into many blocks defined by these two fault trends. The more westerly pipes in this area (95-1, 95-3, 96-1) are located near the northwest extension of the 320° trending South Montreal River Fault.

Geophysically, the Cobalt–New Liskeard kimberlite pipes display both normal and reversed magnetic polarity. Geochronological studies indicate that kimberlite emplacement took place between 125 and 155 Ma, with younger pipes displaying a reversed polarity. Kimberlite intrusion spans a polarity reversal in the Earth's magnetic field at approximately 140 Ma. Kimberlites on the east side of the Lake Timiskaming Structural Zone are dominantly hypabyssal facies while those on the west are diatreme.

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**Figure 34.2.** Location of kimberlites in the region of Cobalt and New Liskeard.
facies, suggesting that the east side represents a deeper structural level. This implies post-Jurassic east-side-up movement across the Lake Timiskaming Structural Zone. All kimberlites examined displayed a mild postplacement brittle fracturing cemented by white carbonate. The Lake Timiskaming Structural Zone remains seismically active (Bent 1996).

KIRKLAND LAKE AREA

The kimberlites of the Kirkland Lake area occur along the northwest trend of the Lake Timiskaming Structural Zone between the Larder Lake–Cadillac Shear Zone to the south and the Destor–Porcupine Deformation Zone to the north (Figure 34.3). In the Kirkland Lake region, the A-1, AM47 and A-4 kimberlite pipes lie above a topographic linear in the bedrock beneath the Munro Esker (Fortescue et al. 1984). This topographic feature, known as the Victoria Lake lineament, lies where inflections of isomagnetic contours imply the presence of faulting.

West of the Victoria Lake lineament, and at approximately the same distance from the lineament, lie the Morriset Creek and B-30 pipes. These intrusions may lie along a subparallel structure but there is no geological or geophysical evidence for its existence. Mapping by Rupert and Lovell (1970) and Jensen (1972) indicates that extensions of northwest-trending faults pass through or close to the kimberlite pipes. These unnamed local faults trend at 300° (Morriset Creek) and 310° (B-30), somewhat oblique to the 325° northwest trend postulated for the Lake Timiskaming Structural Zone.

East of the Victoria Lake lineament, the C-14 pipe in Clifford Township lies close to or along the 045° trending Murdoch Creek–Kennedy Lake Fault (see Figure 34.3). Mapping by Jensen (1974) indicates that the area of kimberlite intrusion contains a number of minor northwest- and northeast-trending faults.

The Diamond Lake kimberlite lies beneath the Misema Esker in western McVittie Township and occurs along the flank of a diabase dike. Fortescue et al. (1984) have indicated that many eskers complexes follow topographic lows that may represent faults or lineament structures. There is no geophysical or geological evidence of such a bedrock feature below the Misema Esker. West of the Diamond Lake kimberlite pipe in eastern Gauthier Township, Thomson and Griffis (1941) identified faulting at 300° and in the western portion of McVittie Township, Thomson (1941) recognized 015° trending faulting. The northwest-trending fault in Gauthier Township is part of the Larder Lake–Cadillac Shear Zone. The Diamond Lake kimberlite occurs just north of the Larder Lake–Cadillac Shear Zone along the southwest extension of a northeast-trending fault in western McVittie Township.

There are no known kimberlites between Kirkland Lake and the region just east of Matheson where a number of kimberlite pipes are present in Guibord Township, and kimberlite dikes are known as far east as Garrison Township (Figure 34.4). These pipes and dikes occur where the east–west trending Destor–Porcupine Deformation Zone intersects the northwest extension of the Lake Timiskaming Structural Zone. All of the identified kimberlite pipes along this east–west trend occur in Guibord Township, within or proximal to the Destor–Porcupine Deformation Zone. All were located as a result of gold exploration.

ATTAWAPISKAT AREA

Early Jurassic kimberlite pipes occur along the geophysically defined Winisk River Fault separating the Sachigo and Winisk subprovinces of the Superior Province. These kimberlites are located where the geophysical data suggest that the Winisk River Fault splays to the east, just southeast of a geophysically inferred, northeast-trending Precambrian diabase dike. Dikes of similar trend near Wawa are Precambrian in age (Sage 1994).

Within the Sachigo subprovince, KWG Resources Inc. has identified kimberlites of Precambrian age beneath the Paleozoic cover. The distribution of the Precambrian kimberlites, so far, appears more or less random and restricted to the Sachigo subprovince. The most intensely investigated Precambrian kimberlite is the diamond-bearing Kyle Lake No. 1 dated by the Rb–Sr method as 1100 ± 40 Ma (L. Heaman, University of Alberta, unpublished data, 1996).

PSEUDOKIMBERLITES

Intrusions equivalent or slightly older in age and closely comparable in composition and appearance to kimberlite have been outlined in two areas of Ontario by diamond drilling completed by Selco Exploration Corporation Limited in the early 1980s. In the Sextant (Coral) Rapids area within the Kapuskasing Structural Zone, a cluster of mafic to ultramafic diatremes occurs where the northwest regional trend of kimberlite emplacement intersects the Kapuskasing Structural Zone. This cluster occurs along the crest of the Fraserdale Arch and where the extension of the Gravel River Fault intersects the Kapuskasing Structural Zone. East of Sextant Rapids in the Hearst area, a second cluster of mafic to ultramafic diatremes occurs that displays a crude northeast-elongated pattern enveloping the Maritson Carbonatite on the southwest, northwest and northeast flanks. The mafic to ultramafic diatremes located in these two areas have been classified as alkaline diatremes, carbonatites and alnoites (Reed and Sinclair 1991).

Some of these intrusions, particularly those in the Sextant Rapids area, contain a complete suite of kimberlite indicator minerals, complicating the identification of true kimberlite. The geochemistry of these indicator minerals is not particularly favorable for diamond preservation when applying current ideas on the relationship of mineral chemistry to diamond preservation (Fipke, Gurney and Moore 1995). None of these intrusions is known to contain diamond. In the Hearst cluster, intrusion 425-12 returned a K-Ar isotopic age of 152 ± 8 Ma and intrusion 423/6-12 returned an age of 180 ± 9 Ma (Reed and Sinclair 1991).

The Jurassic isotopic age on the 425-12 pipe is similar to ages obtained on kimberlite pipes of the Lake Timis-
Figure 34.3. Sketch map showing location of kimberlite pipes in the region of Kirkland Lake.
Figure 34.4. Kimberlites of Guibord and Michaud townships, Matheson, Ontario.
kaming Structural Zone (Sage 1996) and the early Jurassic age on 42J/6-12 is similar to those obtained on Attawapiskat kimberlites (Sage 1997). In the Sextant Rapids area, a U–Pb isotopic age on the indicator-mineral rich I5-10 pipe returned a value of 235.6 ± 2.2 Ma (L. Heaman, University of Alberta, unpublished data, 1998). I5-10 is thus Triassic and older than most of the kimberlites found along the northwest trend across northern Ontario. Additional U–Pb isotopic dating is required to establish whether the two areas of pseudokimberlite are equivalent in age and if they are coeval with kimberlite emplacement.

CONCLUSIONS

1. Kimberlite occurrences within Ontario occur along a northwest trend that is not defined by a specific geological or geophysical feature.

2. Major structural features along this trend are the Precambrian Winisk River Fault Zone and the Proterozoic Lake Timiskaming Structural Zone. The development of these structures predates the Jurassic kimberlites.

3. Kimberlites occur where the broad northwest regional zone is intersected by lineaments and faults or major lithologic boundaries.

4. There is no direct relationship between the development of the structural patterns and kimberlite emplacement. The structures have provided a locus of weakness that guided the emplacement of kimberlite magma into the crust.

5. Kimberlites on the east side of the Lake Timiskaming Structural Zone may represent exposure at a deeper structural level than those on the west side. This suggests postkimberlite east-side-up movement across the zone.

6. Geochronological studies indicate that kimberlite emplacement along the Lake Timiskaming Structural Zone took place between 125 and 159 Ma. In the Attawapiskat area, a limited number of age dates indicate that emplacement took place around 180 Ma.

7. Geochronological and magnetic studies suggest that kimberlite emplacement along the Lake Timiskaming Structural Zone took place across a polarity reversal in the Earth’s magnetic field at approximately 140 Ma.

8. All kimberlites display weak postemplacement brittle deformation.

9. The regional northwest trend of kimberlite emplacement is approximately orthogonal to regional structural trends of the Kapuskasing Structural Zone, the Grenville Deformation Zone, Cape Henrietta Marie Arch and the Fraserdale Arch.

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35. Project Unit 93–12. The “Sandor” Diamond Occurrence, Michipicoten Greenstone Belt, Wawa, Ontario

R.P. Sage

Ontario Geological Survey, Precambrian Geoscience Section

BACKGROUND

A dike was observed by veteran prospector Sandor Surmacz and geologist Marcelle Hauseux along Highway 17 in Lalibert Township north of Wawa on June 22, 1993 (S. Surmacz, Prospector, personal communication, 1997). Exploration was being conducted in the area under an Ontario Prospector Assistance Program (OPAP) grant. Diamond indicator minerals were not found at the site but a representative sample was taken for geochemical analysis. This analysis failed to disclose a geochemistry favorable for the presence of diamond and the work was suspended (S. Surmacz, Prospector, personal communication, 1997).

After examining a diamond-bearing dike containing crustal xenoliths on the Thirsty (Aklulik) diamond project in the Northwest Territories with Complex Minerals Corporation in 1995, the Wawa dike was revisited (S. Surmacz, Prospector, personal communication, 1997).

A bulk sample collected from one dike in 1995 was sent to the Saskatchewan Research Council’s diamond-recovery laboratory in Saskatoon and 6 gem diamonds were recovered (press release, KWG Resources Inc., June 6, 1996; S. Surmacz, Prospector, personal communication, 1997). A permit to prospect was obtained from the Algoma Central Railroad, ground was claimed and an option agreement was made with Spider Resources Inc. conditional on securing an agreement with the Algoma Central Railroad (S. Surmacz, Prospector, personal communication, 1997). The agreements with the Algoma Central Railroad were concluded in 1997 and prospecting in the region by Spider Resources Inc. began. Thomas, Novak and Janse (1998) provide a brief history of the discovery, testing and geology of the “Sandor” diamond occurrence.

The “Sandor” diamond occurrence is the first discovery of a bedrock diamond source in the Michipicoten region since veteran prospector “Mickey” Clement first reported in 1993 the presence of alluvial diamond in the Michipicoten River to the Ontario Geological Survey. This latter discovery prompted the Ontario Geological Survey to conduct a kimberlite indicator heavy mineral survey near Wawa, which disclosed kimberlite indicator minerals in alluvium (Morris, Murray and Crabtree 1994). Prospecting for kimberlite and diamond in the region intensified and has resulted in identifying an ultramafic dike with mantle xenoliths (Sage and Crabtree 1997) and the announcement of kimberlite discoveries in the area (press release, Canabrava Diamond Corporation, February 1998). The “Sandor” diamond occurrence does not contain kimberlite indicator minerals, so while the presence of indicator minerals can be assumed to have stimulated much prospecting activity in the region, it did not play a direct role in the “Sandor” discovery.

Higgins (1986) examined a xenolith-bearing, diamond-absent dike exposed in a roadcut on Highway 17 in north Lalibert Township and classified it as a spessartite lamprophyre. A spessartite lamprophyre is defined as being composed of phenocrysts of green hornblende or clinoxyroxene in a groundmass of sodic plagioclase with accessory olivine, biotite, apatite and opaque oxides (Johannsen 1970, p.191; Bates and Jackson 1987, p.632) and a comparable definition is given by Rock (1990) and LeMaitre (1989).

The following discussion is focussed on the field relations and classification of the dike rocks hosting diamond in order to stimulate exploration. These observations are a selected presentation of data.

FIELD OBSERVATIONS

Distribution

Dikes petrographically similar to the diamond-bearing dike are found within a restricted area in the northwest part of the belt; specifically west of the former Magpie Mine in Leclaire Township and east of the Dickenson Lake Syenite Stock in Lalibert Township. They are also present in the northern third of Menzies Township (S. Surmacz, Prospector, personal communication, 1996). The occurrences are centred on Lalibert Township (Figure 35.1). Dikes considered as a potential host for diamond have not been observed in the granitic terrane external to the Michipicoten Greenstone Belt. A diatreme event postdates emplacement of the lamprophyres and a later period of biotite-rich lamprophyre diking cuts the diamond-bearing lamprophyres.

Regional Structural Observations

The lamprophyre dikes display structural fabrics conformable to regional-scale structural trends indicating that they were deformed along with the enclosing supracrustal rocks. Some dikes appear conformable to stratigraphy and it is only upon detailed inspection that crosscutting relations can be established. The diamond-bearing dikes appear restricted to a D1 recumbent nappe structure.
Figure 35.1. Sketch map of the Michipicoten Greenstone Belt showing where inclusion-bearing, possibly diamond-bearing, lamprophyre dikes are found.
located in the west-central portion of the Michipicoten Greenstone Belt (Arias and Helmstaedt 1989; Sage 1994; Arias 1996). The nappe structure represents north-south shortening and is imbricated by south-verging reverse faults (Arias 1996). The direction of nappe transport is from south to north (Arias and Helmstaedt 1989; Arias 1996) and it lies immediately north and adjacent to the Kapuskasing Structural Zone Uplift (Percival and West 1994).

**Outcrop Appearance**

Most of the dikes favourable for the presence of diamond are fine grained and contain isolated xenoliths of actinolite + talc + carbonate up to 1 m in size. The xenoliths are rounded, weather in relief and may display a dark, mica-rich envelope several centimetres in width that weathers depressed relative to matrix or xenolith. Several dikes consist of closely packed subrounded to subangular xenoliths that may be composite in nature. One dike, which was diamond-bearing, had closely packed angular xenoliths imparting a tuffaceous appearance to the dike. The dikes favourable for the presence of diamond are cut by a diatreme event in which angular blocks of dike material are enclosed and by a later period of mica-rich lamprophyre dikening. Most dikes are only a few metres in width even though wider dikes have been observed in Leclaire Township. The xenoliths contain no primary mineralogy. Most xenoliths are fine-grained actinolite with or without talc and some display compositional zoning from talc core to actinolite rims. The matrix displays a metamorphic texture and kimberlite indicator minerals are either absent or display compositions atypical of kimberlite.

**DIAMOND**

Sampling of the dike matrix by the Ontario Geological Survey indicates that diamond is present in the matrix. The xenoliths were not tested for diamond. Most of the dikes do not contain diamond but, when present, the diamonds consist of very complex, multiple, octahedral twins.

**GEOCHRONOLOGY**

The diamond-bearing dikes were interpreted to be Archean during regional mapping based on regional fabrics cutting through the dikes (Sage 1994). To provide an absolute age of the dikeing a sample of the dike matrix at the discovery outcrop was collected and a U-Pb isotopic age of 2703 ± 42 Ma (discordance 7.3%) was obtained using rutile and sphene (L. Heaman, University of Alberta, unpublished data). This age is slightly older than the oldest U-Pb isotopic age of 2701 ± 7.7 Ma obtained on felsic metavolcanic rocks in the upper volcanic cycle of the Michipicoten Greenstone Belt near McCormick Lake, Lalibert Township (Sage 1994). This suggests that the diamond-bearing lamprophyre dikes were emplaced at the close of mafic volcanism or at the start of felsic volcanism within the youngest cycle of Archean volcanism in the greenstone belt. The dikes containing the diamond display a mildly alkaline composition on the QAPF diagram of LeMaitre (1989) (Figure 35.2). Archean alkaline volcanics were not recognized within the supracrustal sequence but

![Figure 35.2. QAPF diagram for normative mineralogy of dike matrix.](image-url)
two alkaline intrusions have been outlined (Sage 1994). The silica-saturated Dickenson Lake Syenite Stock (2677 ± 4.5 Ma) occurs at the western limit of distribution of the dikes favorable for diamond. The dikes occur close to the eastern margin of the stock but are absent adjacent to the faulted western contact. In the eastern part of the greenstone belt, the silica-undersaturated Herman Lake Nepheline-cancrinite Syenite Stock (2711±10 Ma) is spatially separated from any dike known to host diamond (Sage 1994). There is a similarity in rare earth element (REE) distribution pattern between the dike matrix and the Dickenson Lake stock but not the Herman Lake stock (Sage 1994, Figure 22-8) (Figure 35.3). The radiometric age and geochemistry suggest the possibility that the diamond-bearing lamprophyre diking and the emplacement of the Dickenson Lake stock may be closely related in space and time.

PETROGRAPHY

The dike matrix consists of actinolitic amphibole, biotite and albite (10.70 to 11.31% Na₂O) plagioclase. The texture is very fine to fine grained equigranular granoblastic to deccusate. The dikes have been regionally metamorphosed and subjected to hydration and carbonitization along with the enclosing supracrustal rocks. Except for regional metamorphic effects, the dikes have not been subjected to a metasomatic event that would significantly alter original compositions. The metamorphism of the dike matrix and total alteration of the xenoliths indicate classification should be based on normative mineralogy.

INDICATOR MINERALS

Minerals commonly used to interpret the presence of kimberlite, the dominant host rock for diamond, are absent from these diamond-bearing lamprophyres. Two almandine garnets, commonly from a crustal source, were recovered from the discovery outcrop but no chrome pyrope garnets or chrome diopsides (Overburden Drilling Management Ltd.). Chromite was recovered from the discovery outcrop as well as from a second diamond-bearing dike located in northeastern Menzies Township. This chromite is low magnesium (less than 2.0%) and high chrome (some grains are over 60%). Geochemically, Cr₂O₃ versus MgO plots are atypical of kimberlite chromite. The chromites are zinc-rich (2.0 to 4%), which is also atypical of kimberlite chromites. While the chromite chemistry is not comparable with chromite from kimberlite, the low-magnesium, high-zinc characteristic suggests that it could be used to outline down-ice dispersion patterns in soils to assist in locating unexposed dikes.

Ilmenite is less common than chromite and plots as being oxidized on Halliday's (1975) parabolic curves and Halliday and Tompkins (1984) hematite-geikielite-ilmenite plots. In the discovery dike, all the ilmenites are oxidized and plot outside the field of kimberlite ilmenite compositions (Halliday and Tompkins 1984); however, in the second dike in northeastern Menzies Township two ilmenite populations are indicated. One of these ilmenite populations is much less oxidized and plots within the field of kimberlite ilmenite compositions as outlined by Haggerty and Tompkins (1984). The highly variable ilmenite compositions and relative low volume of material make this mineral an unlikely candidate to define down-ice dispersion patterns.

Chromite and ilmenite are present only in the diamond-bearing dikes. The presence of these minerals in a dike suggests it is a likely diamond host.

NOMENCLATURE

Due to metamorphism and alteration of both the dikes and the xenoliths, normative mineralogy has been used to classify the dike. Major element chemistry of the dike matrix and of the xenoliths was plotted on a Jensen (1976) diagram (Figure 35.4). The matrix has a high-iron tholeiitic to high-magnesium basalt composition (a) and the xenoliths all plot in the ultramafic komatite field (b). Normative mineralogy has been plotted on the QAPF diagram of LeMaitre (1989) (see Figure 35.2). This plot indicates the "Sandor" discovery dike to be slightly alkaline (star; foid-bearing trachyte); that the diamond-absent dikes (open circle) fall on the Or + Ab-An join and that the late biotite-rich dikes that cut the xenolith-bearing/diamond-bearing dikes are silica oversaturated (solid circle; quartz latite). The diamond-bearing dikes may display a mildly alkaline nature since the discovery dike plots in the foid-bearing trachyte field. Normative minerals plotted on a Hy-Di-Ol diagram (LeMaitre 1989) indicate a matrix composition equivalent to websterite, olivine-websterite or olivine clinopyroxenite and on a Di + Hy -Ol-An diagram a composition approximating gabbro to gabbronorite to norite-olivine gabbro (LeMaitre 1989; Figure 35.5a). A plot of normative mineralogy for the xenoliths on the above triangular graphs indicates pyroxenite compositions (LeMaitre 1989; Figure 35.5b). Rock geochemistry and normative mineralogy support classification of the lamprophyres as spessartites and it is suggested that the diamond-bearing dikes may be mildly alkaline.
CONCLUSION

The Archean diamond-bearing lamprophyres of the Wawa area are likely spessartites that have incorporated large fragments of pyroxenite.

RECOMMENDATIONS TO THE PROSPECTOR

The diamond-bearing lamprophyre dikes of the Wawa area represent relatively small exploration targets. The area is thickly mantled with bush and it is highly probable that many undiscovered diamond-bearing dikes remain to be found. Exploration will be difficult due to the general lack of minerals of the kimberlite heavy mineral indicator suite. The presence of low-magnesium, high-zinc chromite in the diamond-bearing dikes suggests that this mineral may be a good mineral to use to define down-ice dispersion patterns from favorable dikes. Being restricted to diamond-bearing dikes makes it particularly good mineral for tracing purposes. Since both dikes and xenoliths contain actinolitic amphibole it may be worth experimenting with the use of actinolite as a tracer mineral for these dikes; however, this mineral will not discriminate between diamond-bearing and diamond-absent dikes.

Figure 35.4. Jensen (1976) cation plot of major element chemistry for dike matrix and xenoliths.
Figure 35.5. Normative mineral Hy - Di - Ol and Di + Hy - Ol - An plots for matrix and xenoliths.
REFERENCES


Before kimberlite was known in Ontario it was a long
held belief that the float diamonds found in the American
Midwest originated in the Canadian Shield (Hobbs 1899).
Efforts to locate the source of the float diamonds over the
years in Canada were dilatory but became increasingly
focussed in the Lake Superior region as geologic knowl-
edge of the Canadian Shield expanded. Efforts in the Lake
Superior region focussed on two areas; i.e., Michipicoten

Figure 36.1. Major regional structures and alkalic rock intrusions within eastern Canada (all alkalic rock and carbonatite complexes are not shown) (Modified from Lumbers 1978, Brummer 1978 and Sage 1991).
Figure 36.2. Schematic diagram showing the location of alkalic rock and carbonatite complexes at the northern end of the Ottawa–Bonnechere Graben, Lake Nipissing Area (modified from Ayres et al. 1970, Lumbers and Milne 1979 and Sage 1991). 2–Brent Crater, 3–Callandar Bay Alkalic Complex, 4–Manitou Island Alkalic Complex, Burritt Island Alkalic Complex, 6–Iron Island Alkalic Complex, 7–Lavergne Carbonatite Complex, 8–Spanish River Carbonatite, Nemag and Lusk Lake Fenites, Allen Lake Carbonatite Complex and 52–Sullivan Island Carbonatite Complex.
Figure 36.3. Schematic sketch map indicating the location of alkaline rock and carbonatite intrusions along the northern part of the Trans-Superior Tectonic Zone (TSTZ) or Accommodation Zone (AZ) in the region of Marathon (modified from Sage 1991). The major carbonatite, diatreme and alkaline rock intrusions and major region faults are: 30-Chipman Lake fennites and carbonatite dikes, 31-Killala Lake Alkaline Rock Complex, 32-Prairie Lake Carbonatite Complex, 33-Port Coldwell Alkaline Rock Complex, 36-Slate Islands Diatremes (cryptoblastion structure), 47-Dead Horse Creek Diatreme, 48-McKellar Creek Diatreme, 49-Gold Range Diatreme, 50-Neys Diatreme, A-Michipicoten Island Fault, B-Accommodation Zone and its extrapolated extension, Killala Lake Deformation Zone.
and Marathon. This is the result of the recognition of the
presence of deep crustal fracturing (Hinze et al. 1966),
which is parallel to the trend of the Lake Nipissing Structural Zone.

The Marathon area can be compared with the regional
structural patterns of the Lake Nipissing area. In the Lake Nipissing area, the Ottawa–Bonnechere Graben splits into
the northwest trending Lake Timiskaming Structural Zone
and an east trending series of faults (Sage 1991) (Figures
36.2 and 36.3). Carbonates are commonly associated
with extensional rifting (Bailey 1964, 1974) and the
east-west trend of carbonate intrusions in the Lake Nipissing area suggests that rift-like tectonics may exist at
depth even though it is not recognized at surface. The
east-west trend is subparallel to the Grenville Front Tectonic Zone whereas the Lake Timiskaming Structural Zone is roughly orthogonal to it. Deformation along the
Grenville Front Deformation Zone may have had some
influence in modifying any east-west rift structure that
may have existed leaving the Lake Timiskaming branch
relatively unmodified. If the east-west trend can be
assumed to be a modified rift, an analogy can be made to a
triple junction in response to mantle plume activity as
relates seismic activity along the Ottawa–Bonnechere
Graben System to active rifting in response to hot spot
(mantle plume) activity and Mereu et al. (1986) have
interpreted seismic data to indicate mantle material has
been emplaced into the graben system. In spite of this
tectonic activity, arching or doming that commonly
accompanies rifting has not been recognized (Doig 1970).

Marathon occurs on the northern flank of the Midcon-
tinent Rift at the apex of a flexure in the regional trend (see
Figure 36.1). From Marathon one branch of the rift trends
southwest along Lake Superior towards Duluth, Minnesota
and the other branch south along Lake Superior towards
Sault Ste. Marie. Early in the history of Lake Superior Basin studies it was recognized that these two rift segments
were separated by north trending faulting (Hinze et al.
1966), which bisected the rift structure. This fault has been
subsequently referred to as the Big Bay–Ashburton Bay
Fault (Sage 1978) or the Thiel Fault (Klasner 1982) and
recently as an Accommodation Zone (Cannon 1994). This
Accommodation Zone separates the Lake Superior Basin of the Midcontinent Rift into two blocks with contrasting
structural styles. West of the Accommodation Zone the
structural style in the Proterozoic is one of thrusting and
tight folding and east of the zone, strike slip faulting and
open folding (Cannon 1994; Mariano and Hinze 1994).
Such contrasting structural styles across a major fault is
analogous to Archean subprovince boundaries (Card and
Cielański 1986) however this contrasting structural style
cannot be traced north of Marathon on the flank of the
Midcontinent Rift and within the extension of the Accom-
mmodation Zone. North of Lake Superior the Accommoda-
tion Zone is represented by faulting that can be traced at
least 140 km north to the Chipman Lake area in the
Geraldton–Longlac region (Sage 1978, 1985) (see Figure
36.3). The presence of alkalic rock–carbonatite complexes
and breccia pipes with carbonate geochemistry along
these faults suggests these faults penetrate to the lower
crust–upper mantle region (Sage 1982, 1991; Smyk et al.
1993). A gravity survey of the Port Coldwell Alkaline
Complex that lies within this trend confirms it is rooted in the mantle (Mitchell et al. 1983).

The analogy of a regional fault separating two contrasting structural domains as being favourable to kimberlite emplacement can be made for the Winisk River Fault, which separates the Winisk and Sachigo Subprovinces. The kimberlites of the Attawapiskat area occur within the geophysically inferred Winisk River Fault, thus deep penetrating crustal faulting as well as rifting is favourable for kimberlite emplacement in Ontario.

The Midcontinent Rift is the product of mantle plume (hot spot) activity (Hutchinson et al. 1990; Nicholson and Shirley, 1990; Allen, Hinze and Cannon 1992) however it lacks a clear triple junction that characterizes plume activity (Burke and Dewey 1973). Burke and Dewey (1973) proposed that the Kapuskasing Structural Zone was a failed arm, Weiblen (1982) proposed that extension of the Accommodation Zone in the Marathon area was the failed arm and Franklin et al. (1980) proposed the failed arm to be in the Lake Nipigon region at the northwest corner of Lake Superior. The Kapuskasing Structural Zone is a cratonic uplift representing 27 km of brittle upper crustal shortening (Percival and West 1994), the area north of Marathon is characterized by faulting, which locally appears strike-slip, and not rifting. The Nipigon area displays rift-like faulting in which the epiclinal Sibley Group sediments are preserved (Coates 1972) and the model of Franklin et al. (1980) suggesting the Nipigon area to be the failed arm is likely correct.

The interpretation that the rifting of the Lake Nipissing area and the development of the Midcontinent Rift is the result of plume activity is of importance to diamond exploration. Helmstaedt and Gurney (1995) model plume activity as destructive of the mantle-root and not favourable for the preservation of diamond. The relatively low diamond content of kimberlites along the Lake Timiskaming Structural Zone can be explained by this model (Sage 1996). The diamond content of a kimberlite approximately 20 km west of the Lake Timiskaming Rift is reported to have higher grade diamond mineralization (19 macro and 33 micro diamonds in 2400 pounds of material) (press release Sudbury Contact Mines Ltd., February 12, 1997). Diamond contents may be improving as one moves away from the rift and presumed mantle plume influences of rift formation. Applying these observations and the model of Helmstaedt and Gurney (1995) to the Marathon area, one would prospect north from Lake Superior along the extension of the Accommodation Zone to the area of Geraldton–Longlac. This location would distance oneself from the mantle plume influences that formed the Midcontinent Rift. Kimberlites could still be present in the Marathon area but their proximity to the Midcontinent Rift may mean relatively low diamond content.

**Spatial Distribution**

In the Lake Nipissing area alkalic rock-carbonatite complexes occur along the Ottawa–Bonnechere graben and east-west along an inferred rift at depth (see Figure 36.2). Kimberlites occur along the Lake Timiskaming Structural Zone (Rift) and with the exception of small carbonatite dikes (Bennett 1978) in the Temagami area, the kimberlite intrusions are spatially separate from the other types of alkalic intrusion. This observation predicts that any kimberlites along the extension of the Accommodation Zone from Marathon to the Geraldton–Longlac area will display no spatial relationship to other forms of alkalic magmatism found in the region (see Figure 36.3).

**Geochronological Patterns**

The carbonatites of the Ottawa–Bonnechere Graben and the east-west inferred rift are dominantly late Proterozoic to early Cambrian and one, Sullivan Island, is Proterozoic (Keweenawan) in age (Gittins et al. 1967; Sage 1991). The kimberlites along the Lake Timiskaming Structural Zone are Jurassic (Sage 1996) and much later in emplacement than other types of alkalic magmatism. Thus there need not be a close secular relationship between kimberlite intrusion and other forms of alkalic magmatism. The common relationship is emplacement into a pre-existing zone of structural weakness.

Applying the observations of the Lake Nipissing area to the Marathon area and north to the Geraldton–Longlac region, kimberlite intrusion need not be coeval with the Midcontinent Rift magmatism at 1100 Ma. An interesting observation is that extending the trend of faulting associated with the Accommodation Zone north from the Geraldton–Longlac area into the James Bay region one observes that the diamond-bearing Proterozoic age Kyle Lake No. 1 kimberlite (1100 ± 40 Ma, Heaman 1996, University of Alberta, unpublished data) lies along this extension (see Figure 36.3). At present there is no geological or geophysical justification for such an extension but the isotopic age of the intrusion indicates that diamond-bearing kimberlite emplacement did take place during the time period that the Midcontinent Rift developed and exploration should not be focussed on Jurassic age kimberlites.

**Cryptoexplosion Features**

At the junction of the Ottawa–Bonnechere Graben, the Lake Timiskaming Structural Zone (Rift) and the east-west faulting in the Lake Nipissing area lies the Brent Cryptoexplosion structure (see Figure 36.2). Near Marathon, the Slate Island Cryptoexplosion structure occurs at the junction of the north margin of the Midcontinent Rift and the Accommodation Zone trending north from the Rift (see Figure 36.3). Both features have been interpreted as meteorite impact scars due to the presence of shock metamorphic features in spite of their relatively unique structural setting (Shafiqullah, Tupper and Cole 1968; Dressler and Sharpont, 1997). The presence of shock metamorphic planar deformation lamellae in quartz have been used as one of the main criteria in proposing that these two sites are impact sites, however, planar deformation lamellae similar to those used to establish impact have been described in xenocrystic quartz from kimberlite (Dawson et al. 1997).

The presence of cryptoexplosion structures in a structural zone being considered as an exploration target
for kimberlite can be interpreted as a positive feature. These features represent rapid, explosive emplacement by a volatile-dominated magmatic system. Kimberlite dikes are rapidly emplaced and if diamond is to be preserved during transport from its point of origin where it is in chemical equilibrium to the surface where it is in disequilibrium, this transport needs to be rapid (Harris 1985; Eggle 1989). Cryptoexplosion features indicate that structural conditions exist for rapid emplacement of kimberlite magmas.

CONCLUSIONS

The Geraldton–Longlac region is a favourable target area for kimberlite exploration.

1. The Geraldton–Longlac area satisfies the conditions of "Clifford's Rule" that a thick stable crust is required for commercially viable diamond-bearing kimberlite pipes (see Janse 1991 for a review).

2. The Geraldton–Longlac area contains a proposed extension of deep faults that have controlled emplacement of lower crust–upper mantle alkalic magmas. These faults are the extension of faults associated with the Accommodation Zone that bisects the Midcontinent Rift.

3. The Geraldton–Longlac area lies along strike from a cryptoexplosion structure implying that the geological structures are appropriate for rapid emplacement of volatile-rich kimberlite magmas.

4. The Geraldton–Longlac area lies some distance from the margin of the mantle plume formed Midcontinent Rift and away from the mantle root destructive influences of the plume. Diamond preservation should be enhanced.

5. Kimberlites along this zone need not be coeval with other alkaline rock magmatic events in either space or time.

Recommendations to the Prospector

1. Regional reconnaissance investigation of alluvium for the presence of kimberlite indicator minerals in the down-ice region of Marathon (see Figure 36.3). Chrome-bearing pyroxene occurs in non-kimberlite dike rocks of this area (Laderoute 1987) so the presence of chrome diopside should be interpreted with extreme caution in establishing the presence or absence of kimberlite.

2. Define or outline kimberlite indicator mineral dispersion patterns to define target areas for detailed prospecting.

3. Detailed geophysical investigation of promising areas outlined by kimberlite indicator mineral dispersion patterns. The targets will not likely be exposed, relatively small in size (few hundreds of metres) and could display either reverse or normal polarity such as occurs with kimberlites along the Lake Timiskaming Structural Zone (Sage 1996).

4. Testing of outlined targets by either diamond or reverse circulation drilling to establish the presence of kimberlite.

5. By analogy with the Lake Timiskaming Structural Zone (Sage 1996) kimberlites display a preference to occur where the dominant structural trend is obliquely intersected by linear features, faults or major lithologic contacts.

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INTRODUCTION

A multi-year project to produce a 1:1 000 000 scale metamorphic map of the Province of Ontario was initiated by the Ontario Geological Survey in 1994. In part, this project resulted from a request by the Geological Survey of Canada for Ontario to provide data for an updated metamorphic map for the Canadian Shield (Fraser et al. 1978) and to be used in producing a metamorphic map of the world (in conjunction with the Commission for the Geologic Map of the World). It is also an attempt to integrate metamorphic information with the tectonic synthesis produced as part of the Geology of Ontario project, with the aim of developing improved mineral exploration models.

The metamorphic map of Ontario is designed to complement the set of 1:1 000 000 bedrock, surficial geology, geophysical, and tectonic assemblage maps and time-space charts released in 1991–92 as part of the Geology of Ontario series. In addition, the GSC will produce a metamorphic map of the Canadian Shield at roughly 1:2 500 000 scale, incorporating data compiled during this project.

COMPILATION PROCESS AND STATUS REPORT

The map is being produced digitally, using the Tectonic Map of Ontario as a base map. Table 37.1 outlines the major steps involved in the project; note that the steps are

<table>
<thead>
<tr>
<th>Step</th>
<th>Description</th>
<th>Processes Involved</th>
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<tbody>
<tr>
<td>1</td>
<td>Compilation of isograds, P-T data, mineral assemblage data, timing of plutonism, identification of contact aureoles, compilation of metamorphic geochronology.</td>
<td>Compiled at 1:250 000 scale on paper base maps, linework digitized using a CAD program, point-source information put into database.</td>
</tr>
<tr>
<td>2</td>
<td>Identification of metamorphic map units, using information compiled in Step 1, as well as geophysical and geological interpretation. During this stage, information compiled in Step 1 is verified and updated.</td>
<td>Compiled at 1:250 000 scale on paper base maps derived from Geology of Ontario tectonic map. Map unit polygons then digital encoded to corresponding metamorphic map units, existing polygons commonly need to be redrawn and modified during this process. Digital work done using ArcView™ software.</td>
</tr>
<tr>
<td>3</td>
<td>Proofing and correction of map units identified in Step 2.</td>
<td>This is a multi-stage process, first involving in-house editing, and then editing of polygon maps produced by the GSC at their map scale using their map legend; repeated as necessary.</td>
</tr>
<tr>
<td>4</td>
<td>Identification of base map units. Both GSC and OGS base maps will use a simplified rock unit base map. This requires translation of existing units to simplified scheme.</td>
<td>Compiled at 1:500 000 scale on paper base maps derived from Geology of Ontario tectonic map after assigning metamorphic map plutonic units to appropriate base map codes.</td>
</tr>
<tr>
<td>5</td>
<td>Proofing and correction of map units identified in Step 4.</td>
<td>This is a multi-stage process, first involving in-house editing, and then editing of polygon maps produced by the GSC at their map scale using their map legend; repeated as necessary.</td>
</tr>
<tr>
<td>6</td>
<td>Proofing and correction of metamorphic and base map units along provincial borders (Manitoba, Minnesota, Quebec, New York) so that map units mesh as smoothly as possible.</td>
<td>This is a multi-stage process, first involving in-house editing, and then editing of polygon maps produced by the GSC at their map scale using their map legend; repeated as necessary.</td>
</tr>
<tr>
<td>7</td>
<td>Upgrading of P-T data to same standards.</td>
<td>If time is available, P-T data will be recalculated to standard calibrations to allow improved comparison of data of different ages and sources.</td>
</tr>
<tr>
<td>8</td>
<td>Displaying digitally compiled metamorphic age information, such as Ar/Ar cooling ages on the metamorphic map to assist with interpretation.</td>
<td>If time is available, information of timing of metamorphism and cooling history will be displayed on the map to assist in regional interpretation and to determine if such information can locally be displayed on the map.</td>
</tr>
<tr>
<td>9</td>
<td>Overview article to be produced outlining key decisions and assumptions made with respect to portrayal of metamorphic information, and to serve as &quot;marginal notes&quot; to the map.</td>
<td>CD-ROM will contain compiled information (such as isograds) that may not be displayed on paper version of the map for presentation reasons.</td>
</tr>
<tr>
<td>10</td>
<td>Release of GSC Map and accompanying CD-ROM of compiled information.</td>
<td></td>
</tr>
</tbody>
</table>
not entirely sequential, and that work can progress on several steps in the production process simultaneously. Step 1 is completed, although digitization of linework into a CAD program and input of point-source information into a database is continuing. Delays in data input in Step 1 are related mainly to hardware and software difficulties involving the CAD program, largely the result of the large file sizes involved in this project. Step 2 has been completed, and Step 3 is underway, as are Steps 4 to 6.

In the case of the Grenville Province, information was compiled on both the Tectonic Map of Ontario base, and an orogen-wide base map derived from the geologic compilation map of Davidson (1996). The latter base was used, at the behest of the GSC, as it provided a seamless geologic base covering Ontario, New York state, and Quebec. Further, also at the behest of the GSC, in addition to Ontario, metamorphic data for western Quebec and the Adirondack area in New York state was also compiled, in part because of the similarity in geology between Ontario and these regions. This serves not only to reduce map unit changes across geopolitical boundaries, but will also assist in geological interpretation and synthesis that will be a feature of the latter stages of this project.

Two special issues of *The Canadian Mineralogist* on “Tectonometamorphic Studies in the Canadian Shield” are being produced as part of this project. The first, released in February 1998 (Berman and Easton 1997) includes the results obtained from a variety of ancillary metamorphic studies related to the GSC compilation effort. The second issue, which is in the process of being compiled, will include regional overviews related to the metamorphic map of the Canadian Shield, as well as additional, detailed metamorphic studies.

**DISCUSSION OF RESULTS**

Limited interpretation of the compiled data has occurred to date, since at the time of writing, the GSC was producing the first province-wide version of the map showing near-final map units, colours and patterning. A few limited observations can be made at this time.

As noted briefly in Thurston et al. (1998), a characteristic geophysical signature of granulite facies terranes in the province are paired Bouguer and vertical gradient gravity highs. This signature seems to be more characteristic than the “bird’s eye” pattern seen on total field aeromagnetic maps, particularly in areas subjected to low-pressure granulite facies conditions.

There are three main periods of metamorphism in the Superior Province. The first, an event from 2860 Ma to 2810 Ma, is best exposed in the greenstone belts in the North Caribou Lake area, although the event has been identified in localized areas throughout the Superior Province where older crustal remnants are preserved. This earlier event has largely been overprinted by a major event at ca. 2705 to 2690 Ma that mainly affects the granite-greenstone terranes throughout the Superior Province. This event youngs from north to south across the Superior Province. Finally, a third event affects mainly the metasedimentary and gneiss terranes at ca. 2685 to 2665 Ma. What is not clear is if the 2706 to 2690 Ma and 2685 to 2665 Ma events are 2 separate events, or if they simply reflect the same event affecting different crustal levels, that were subsequently juxtaposed through uplift and tectonism.

The Grenville Province consists of two main metamorphic regimes. First, the area that includes both the Central Gneiss Belt and much of the Central Metasedimentary Belt, has been subjected mainly to upper amphibolite facies metamorphism during the period 1070 to 1040 Ma. Although small areas of higher and lower metamorphic grade occur in this area, as do areas showing older and younger periods of metamorphism, this 1070 to 1040 Ma event is the dominant tectonometamorphic event in this part of the orogen, and corresponds to the main period of NW-directed thrusting and tectonism in the orogen (cf. Carr et al. in press). Second, the area including both Frontenac domain, the Adirondack Lowlands and the Adirondack Highlands was subjected to an low-pressure upper amphibolite to granulite facies metamorphic event at ca. 1170 Ma, although the main metamorphic features presently observed in the Adirondack Highlands, namely the high-pressure granulite facies metamorphism, developed between 1070 to 1040 Ma. The tectonic significance of these two distinct metamorphic regimes in the southwestern Grenville Province is elaborated upon in Carr et al. (in press).

**REFERENCES**


Sedimentary Geoscience Section
38. Project Unit 98–050. Aggregate Resources Inventory of the Highway 17 Corridor Between the Wahnapatia Area and Sturgeon Falls

D. Rowell
Sedimentary Geoscience Section, Ontario Geological Survey

INTRODUCTION

During the summer of 1998, an aggregate resources assessment of the Highway 17 corridor between the eastern boundary of the Regional Municipality of Sudbury and Springer Township (District of Nipissing) was undertaken by the Sedimentary Geoscience Section. The study area covers 13 townships or approximately 130,000 ha (Ontario Ministry of Municipal Affairs and Housing and the Association of Municipal Clerks and Treasurers of Ontario 1997) (Figure 38.1). The area is covered by parts of the Sturgeon Falls (31L/5), Marten Lake (31I/12), Coniston (41L/7), Verner (41I/8), Glen Afton (41I/9) and Capreol (41I/10) 1:50 000 scale map sheets of the National Topographic System (NTS).

The purpose of the investigation was to delineate the aggregate deposits within the study area and to assess the quality and quantity of the sand and gravel resources. This information is required for infrastructure development and general construction applications, as well as for land use planning.

BEDROCK GEOLOGY

Most of the study area is underlain by biotite gneisses derived from a relatively thick sequence of thinly bedded greywacke and argillite. These rocks are part of the Grenville Province of the Canadian Shield. Lumbers (1975) described these rocks as a light to dark grey, fine to coarse-grained, biotite-quartz-feldspar gneiss that consists of 10 to 35% biotite, up to 15% garnet, less than 10% hornblende and generally more plagioclase than potassic feldspar. A migmatic biotite gneiss is also located in the study area, Lumbers (1975) describes this rock as similar to the biotite gneiss but veined by 10% or more granitic material.

The northwestern corner of the study area is underlain by arkose, arkosic wacke, wacke and conglomerate rocks belonging to the Huronian Supergroup of the Southern Province (Dressler 1984). The Grenville and Southern provinces are separated by the Grenville Front Boundary Fault.

In the southwestern corner of Awrey Township is the Red Deer Anorthosite. It consists of gneissic anorthosite.
gabbro and gabbroic anorthosite with some metagabbro and metadiorite units near the eastern margin (Lumbers 1975).

The Markstay Pluton is located near the town of Markstay in Hagar Township. It is a small structure of gneissic tonalite, minor gneissic diorite and pink to grey gneissic quartz syenite and monzonite intruding metasediments (Lumbers 1975).

Located in Badgerow and Hugel townships is the Badgerow Complex containing sodic syenite and monzonite. The complex is moderately deformed, with a narrow gneissic margin and a massive to slightly foliated interior. Lumbers (1975) describes these rocks as pink, gneissic to rarely massive, ferro-hastingsite bearing sodic syenite to quartz syenite.

OUTWASH FEATURES IN THE REPORT AREA WERE DEPOSITED UNDER SUBAREAL CONDITIONS BY MELTWATER FLOWING FROM THE ICE MARGIN. THE OUTWASH IS DIVERSE, UNIFORMLY BEDDED SAND AND GRAVEL. OUTWASH IS ONE OF THE MOST WIDESPREAD GLACIAL WEBSITES IN THE STUDY AREA AND HAS BEEN A TRADITIONAL SOURCE OF AGGREGATE MATERIAL.

As the ice margin receded, glacial meltwater was dammed against the wasting ice margin and formed post-glacial Lake Algonquin. As a result, glaciolacustrine sediments are widespread throughout the study area. In the deeper waters of the glacial lake, massive and/or varved silt and clay were deposited. Glaciolacustrine sediments consisting largely of silty fine sand were deposited in the shallow areas of the lake. These glaciolacustrine sands are generally too fine for most aggregate uses. In places, glaciolacustrine material overlies ice-contact sediments.

The thickest and most extensive fine-grained glaciolacustrine sediments were deposited in glacial Lake Nipissing. These flat-lying deposits along the Veuve River and near Lake Nipissing are generally located below 215 m.

In contrast to these flat-lying glacial Lake Nipissing deposits, gently rolling glaciolacustrine deposits, characterized by an irregular surface, are present in various parts of the study area at elevations between 215 and 245 m. These were deposited in glacial and post-glacial lakes older than glacial Lake Nipissing. They consist of pebbly to bouldery, massive clay with variable amounts of interstratified varved clay and sand and gravel.

The final draining of the glacial lake from the study area probably began with the opening of the Mattawa outlet. Tiny islands would have formed and nearshore deposits would have been created around bedrock knobs. Glaciolacustrine raised beaches occur as discontinuous sand and gravel deposits. Drainage of the lake water would have occurred through numerous small bedrock channels. North of Highway 17 just east of Verner, a few strand lines exist below 215 m. However, most beach deposits in the study area appear to be closer to the 245 m elevation.

Erosion of till from rocky uplands by glacial lake water is very pronounced and in general the terrain in the study area is characterized by abundant wave-washed bedrock knobs with widespread glaciolacustrine deposits in intervening low areas. Awrey Township consists of up to 70% bare rocky ridges (Lumbers 1975).

Erosional activity has been minimal since the disappearance of the ice sheet and the lowering of glacial lake water to present day levels. Organic deposits have been developed in depressions in the land surface. Alluvium has been deposited along the courses of existing creeks and rivers, particularly the Sturgeon River near Field. Present day Lake Nipissing is the largest remnant of the glacial and post-glacial lakes in the area.

The highest part of the study area is along the west side of Street Township at about 300 m. The eastern and central parts of the study area are dominated by the western part of Lake Nipissing which has a surface elevation of 196 m.
The lake is less than 18 m deep. Local relief near the lake is less than 15 m. North of Lake Nipissing the land surface rises to between 245 m and 275 m.

AGGREGATE POTENTIAL

Highly weathered, brittle and friable Precambrian bedrock, which appears acceptable for low-specification aggregate use, is common within the report area. There are also many areas that are underlain by more massive, hard and durable rock which appears suitable for a variety of aggregate applications.

The gneissic rocks in the report area are also potential sources of quality aggregate. These rocks are usually hard and relatively homogeneous, but brittle varieties can occur and should be avoided in aggregate use. Massive, coarse-grained felsic plutonic and gneissic rocks with high mica, feldspar and quartz contents may have bonding problems. The smooth cleavage and fracture surfaces of the minerals hinder the adhesion of asphalt and Portland cement concrete mixes. This problem can often be circumvented by weathering the rocks for a period of time in stockpiles or by adding chemicals (anti-stripping agents) which erode the smooth surfaces and allow better adhesion. Rogers (1985) reports that some granitic rocks can slowly react with alkalies from Portland cement concrete resulting in concrete deterioration.

Within the report area, considerable latitude exists in choosing sites for potential bedrock extraction as there are extensive areas where bedrock is exposed at or near the surface.

Although the Precambrian bedrock in the area may meet MTO specifications for concrete aggregate, it may not be accepted by the MTO for use in Portland cement concrete which will be exposed to de-icing salts. Radioactive mineralization may also occur locally within some rock types in the area and these rocks should be avoided during extraction.

Any site proposed for quarry development should be thoroughly tested before extraction commences. The Precambrian rocks may vary in quality over relatively short distances and may be alkali-reactive with Portland cement concrete mixes.

In the study area excellent sources of sand and gravel aggregate material are eskers, other glaciofluvial ice-contact features and outwash deposits. The townships in the northern part of the study area contain potential supplies of aggregate. Some of the sand and gravel material cannot be used in Portland cement concrete mixes because of an alkali-reactivity problem. The southern townships have less aggregate potential because of the large surface expanses of fine-grained glaciolacustrine sediments.

In the southern part of the study area small sand and gravel pits have been developed in leeside cavity fill deposits. These deposits have formed on the southern side of bedrock topographic highs and have provided granular material for small, local projects.

Till is usually not well-suited for aggregate use as it often contains excess fines and abundant cobbles and boulders. However, it may be a suitable source of fill in some locations.

REFERENCES


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INTRODUCTION

The Vermilion River has been investigated for its potential to host deposits of placer gold several times in the past (Gracey 1898, Coleman 1901, Prest 1949). The abundance of colours in pans (Gracey 1898, Coleman 1901, Prest 1949) has maintained the interest in placer gold along the Vermilion River throughout the past 100 years. The current project’s main goals are to re-evaluate the placer gold potential and locate possible sources for the placer gold within the Upper Vermilion River Watershed of northeastern Ontario. The project is also evaluating the use of GIS to predict sample site locations and interpret and present survey results. The methodology used in this project, could be applied to stream sediment sampling surveys in the search for and evaluation of greenstone belts in more remote regions of the Province.

The Upper Vermilion River is a typical stream system in the Canadian Shield where bedrock topography creates numerous lakes and swamps along the drainage network. In regard to stream sediment sampling, these features act as sedimentation traps and prevent or hinder the transport of heavy mineral grains along the stream system. In addition, bars, which are common stream sediment-sampling sites along rivers in other terrain, are poorly developed along this river system. As a result, riffles provided one of the prime sediment sampling sites.

METHODOLOGY

The GIS was used to assist in developing a sediment sampling strategy. A hydrologically correct digital elevation model (DEM) with a spatial resolution of 30 m was created using ArcInfo software from 22.1:20 000 scale digital Ontario Basic Mapping (OBM) maps with a contour interval of 5 m. Watershed basin analysis was performed using various GRID programs. The OBM maps also provided drainage, transportation and other cultural feature information. In addition, polygon and point data from bedrock and Quaternary geology maps as well as mineral occurrences are being incorporated as coverages into the GIS to aid in site selection and later, data interpretation. The DEM will help in both analyzing and in the visual representation of the watershed.

Various queries were run to determine locations for sampling to optimize time in the field and reduce field-related expenses. Important factors considered at a regional or system scale included sub-basin representa-

tion, geological materials being eroded and length of stream above a lake or swamp. And at a local scale important factors include: stream gradient and the locations of riffles/rapids, deltas and bars (obtained from air photographs, OBM maps and/or Landsat TM) were considered.

Once constructed the GIS will also be used in the interpretation of results obtained from the analysis of the stream sediment samples.

It was also decided to characterize and sample the glacial sediments that occur within the drainage basin. Till reflects the character of the debris being transported by ice into and within the basin. The properties of sediments being deposited within glacial meltwater conduit systems can be determined from sampling eskers. The major tracts of outwash sediments are the products of glacially-fed braided streams and provide possible targets for placer gold exploration.

UPPER VERMILION RIVER WATERSHED

The Upper Vermilion River watershed is approximately 660 km² in area and can be subdivided into 67 sub-watersheds.

The watershed is underlain by rocks of Precambrian age. These rocks have been mapped by Meyn (1970, 1971, 1973, 1976) and Dressler (1982) and a short summary taken from their works is presented below.

Archean metavolcanic and metasedimentary rocks, consisting of mafic and felsic flows, iron formation and metasedimentary schists and conglomerates are the oldest rocks exposed within the watershed. Granitic rocks (granite, quartz monzonite, granodiorite, and pegmatite) have intruded these rocks. Mafic dikes have subsequently intruded all of these older rocks.

Proterozoic (Huronian Supergroup) metasedimentary rocks unconformably overlie the above-mentioned rocks. The Huronian Supergroup includes: the Mississagi Formation (quartzites); the Bruce Formation (paraconglomerate); the Gowganda Formation (argillite, siltstone, minor quartzite and paraconglomerate); and the Lorrain Formation (quartzite) (Meyn 1973).

The entire sequence has been intruded by dikes and sills of Nipissing-type quartz diabase, the Sudbury Nickel Intrusive and olivine diabase. All the rocks have been folded and faulted.
Upper Vermillion River Watershed
Heavy Mineral Sample Sites

Legend
■ ice-contact stratified sediment
○ outwash
● stream sediment
▲ till
▲ till (flowtill)

Figure 39.1. Location map of samples taken for heavy mineral analysis.
Rocks of the Sudbury Igneous Complex and the Whitewater Group underlie the very southern part of the watershed. Gold has been reported to occur at the contacts of Nipissing gabbro with Huronian metasediments (Dressler 1982) and within shears in Nipissing-type diabase (Meyn 1971).

The bedrock lithology and structure played a key role in landscape evolution. The landscape is strongly bedrock controlled with glacial sediments commonly occurring as thin discontinuous deposits. The exceptions are along bedrock-controlled valleys where large braided streams draining meltwater from the receding ice margin deposited thick deposits of glaciofluvial outwash sediments. In places these deposits exceed 5 m in thickness.

Evidence of former ice-marginal positions is present, but limited; the exception being the Cartier I Moraine that forms the southern boundary of what is referred to herein as the Upper Vermilion River and its watershed. The study area is included on the surficial geology map of Boissonneau (1965) and engineering terrain geology maps of Gartner (1980a, b). Prest (1949) and Bajc (1997) have mapped the Quaternary geology of the southern part of the area.

**Sampling**

Sampling took place during the month of August. In total, 102 samples were collected and will be processed for their heavy mineral content (Figure 39.1). Of the 102 samples, 54 were of modern stream sediments. Samples were taken from several depositional environments within the stream. In the smaller tributaries, samples were commonly collected in the lee of large boulders, longitudinal bars and point bars and from the base of channels or chutes. In wide stretches of the Vermilion River, samples were taken from riffles and longitudinal and transverse bars. And where the tributaries entered lakes of the Vermilion system, delta topsets, foresets and distributary channel sites were sampled.

Thirty-three samples were taken from exposures of outwash sediment along the length of the watershed. Samples were predominantly of massive, imbricate boulder gravels, a facies that proved to be very common in the project area. Samples were also taken from trough crossbedded gravel and sand facies. Several samples were taken at 3 selected sites, to determine the in-site variation in heavy mineral content.

Eight samples were taken from boulder gravel facies of eskers and 7 samples were collected of till.

Samples were sieved in the field and the minus 7 mm portion sent for heavy mineral processing. In addition, samples from till, outwash and esker sample sites were taken for grain size analysis and for pebble lithology studies.

**Conclusions**

The prediction of sample site locations using GIS was very helpful in the initial stages of project planning. It provided targets for which daily field traverses could be planned. However, site prediction was only as good as the quality of information on the digital base maps used. Most predicted sites that proved to be of no use, were the result of the origin map information being in error, or not consistent with present-day conditions. Several of the other predicted sites were not sampled due to time or access constraints, however, for many, alternative sites were selected following the same general principles.

**Acknowledgments**

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**References**


INTRODUCTION

Increased exploration activity north and west of Thunder Bay has been attributed to the recent recognition of a new gold belt traversing the eastern portion of the Shebandowan greenstone belt (Lavigne and Scott 1994). This 75 km long by 6 km wide gold belt, informally referred to as the “Matawin gold belt”, has yielded over 50 new gold occurrences within the last 15 years. Exploration activity prior to this period focussed primarily on the belt’s iron ore potential. The recent discovery of “high-grade”, base metal float in northeastern Aldina Township also resulted in much of the southeastern portion of the belt being staked and evaluated for its base and precious metal potential (Lavigne and Scott 1997).

Much of the exploration activity within the Shebandowan belt has currently subsided, largely in response to the depressed prices of precious and base metals and the inability of junior mining companies to raise funds for exploration. In addition, after almost 3 decades of operation, the INCO Limited Shebandowan nickel mine was decommissioned during the second quarter of 1998.

In response to this recent decline in mining and exploration activity within the belt, the Ontario Geological Survey initiated a 2 year program of Quaternary geology and drift geochemistry to further evaluate the mineral resource potential of the Shebandowan belt. This was accomplished by detailed (1:50 000 scale) Quaternary geology mapping and regional humus and c-horizon till sampling. Quaternary mapping was undertaken to document the distribution and character of the various Quaternary deposits within the study area and to reconstruct the ice flow history associated with these deposits. This information is essential to better understand how the various surficial materials can be utilized for mineral exploration programs.

During the 1998 field season, Quaternary mapping and regional humus and c-horizon till sampling were conducted within the area covered by the Sunshine (52A/12) and Kakabeka Falls (52A/5) 1:50 000 scale National Topographic System (NTS) map sheets (Figure 40.1). Detailed, property-scale soil sampling was also conducted over and adjacent to the Bylund gold occurrence which is situated along Highway 11/17 at the west end of the Dawson Road Lots (see Figure 40.1). This orientation survey was undertaken to characterize glacial dispersal from a known gold source in terms of magnitude, shape and composition. This information will be applied to the interpretation of the regional sampling dataset. Quaternary mapping and regional drift geochemistry of the Shebandowan (52B/9) and Burchell Lake (52B/10) 1:50 000 scale NTS map areas to the west will be undertaken during the 1999 field season (see Figure 40.1).

BEDROCK GEOLOGY

A comprehensive summary of the bedrock geology of the study area is contained in papers by Sutcliffe (1991), Williams (1991) and Williams et al. (1991) and as well as in numerous Ontario Geological Survey maps and reports. The study area contains 3 distinct bedrock domains. The centrally-located Wawa Subprovince, within which the Shebandowan greenstone belt occurs, is fault-bounded to the north by metasedimentary and felsic intrusive rocks of the Quetico Subprovince and unconformably overlain to the south by Paleoproterozoic metasedimentary rocks of the Gunflint and Rove formations.

Metasedimentary rocks of the Quetico Subprovince consist primarily of turbiditic wackes, arkoses and quartz arenites and their associated paragneisses and migmatites. Post- to synsedimentary felsic plutons consisting of feldspar-megacrystic granite, granodiorite to tonalite and monzonite comprise a notable proportion of this domain. Narrow, elongate lenses of tholeiitic, mafic metavolcanic rocks and associated gabbroic rocks occur locally (Brown 1995).

Metasedimentary rocks of the Gunflint and Rove formations belong to the Animikie Group and were deposited within a large basin that extends into parts of Ontario, Minnesota, Wisconsin and Michigan (Sutcliffe 1991). The Gunflint Formation contains a basal unit of conglomerates that rest unconformably on the Archean basement and are overlain by interbedded argillites, argillite tuffs, cherts, algal cherts, jasper, carbonates, ferruginous carbonates, hematite, magnetite taconite and silicate taconite. The overlying Rove Formation consists of black, locally pyritic shales and grades upward into interbedded black shale and arkosic greywackes. Mesoproterozoic “Logan” diabase sills intrude the Rove Formation and form the resistant cap rocks of
large mesas in the southern part of the map area. Silver mineralization occurs locally along normal faults that intersect the contact between Rove Formation shales and Logan diabase sills. Notable silver deposits occur in the southeastern corner of the Kakabeka Falls NTS map area in the vicinity of Badger and Beaver Mountains (Creswel or Rabbit Mountain group of deposits) and Silver Mountain. Deposit summaries are presented in a paper by Franklin et al. (1986).

The Shebandowan greenstone can be subdivided into 2 contrasting suites of volcanic and sedimentary rocks: 1) an older suite of mafic to felsic volcanic rocks defined by the Burchell and Greenwater assemblages; and 2) a younger suite of sedimentary and volcanic rocks referred to as the Shebandowan assemblage. The Greenwater and Burchell assemblages each consist of 3 oppositely-facing, bimodal volcanic cycles typically consisting of a lower sequence of tholeiitic basalts and an upper sequence of calc-alkaline andesite, dacite and rhyolite (Williams et al. 1991; Rogers and Berger 1995). The younger, Shebandowan assemblage unconformably overlies (i.e., fault contact) and intrudes the Greenwater and Burchell assemblages. The Shebandowan assemblage consists of alkaline metavolcanic and intrusive rocks (tuff, breccia, syenite, lamprophyre, quartz and feldspar porphyry, granodiorite and diorite) as well as metasedimentary rocks (arkoses, wackes, conglomerates and oxide-facies iron formation) (Schneiders et al. 1998). This assemblage is interpreted as representing an intracratonic basin assemblage similar to the Timiskaming assemblage of the Abitibi subprovince. Granitoid intrusions belonging to the Northern Light-Perching Gulf lakes batholith complex occur along the southern margin of the Shebandowan belt. Most of the gold occurrences within the Matawin belt display a close spatial association with the Shebandowan assemblage. The style of gold mineralization is not unlike that which occurs in such prolific mining camps as Timmins and Kirkland Lake. Notable exploration programs within the current study area are highlighted in the last 5 annual volumes of the Resident Geologist’s Report of Activities. Schneiders et al. (1998) presents a succinct list of exploration criteria for programs of gold exploration within the Matawin/Shebandowan greenstone belt.

**GLACIAL GEOLOGY**

The study area is characterized by several contrasting physiographic regions, each defined by a unique suite of overburden conditions. Bedrock-dominated uplands are situated in the northwest corner of the Sunshine NTS map area and the northwest and southeast corners of the Kakabeka Falls NTS map area. Aside from local

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**Figure 40.1.** Regional geologic setting of the Matawin–Shebandowan study area.
accumulations of thick drift in morainic belts and narrow, structurally-controlled depressions, these uplands are essentially bedrock-dominated terrains.

Thicker deposits of till and/or glaciofluvial and glaciolacustrine sediments occur within 2 isolated basins. The northern basin, here referred to as the “Kaministikwia basin”, occupies the valleys of the Oskondaga, Matawin, Shebandowan and Kaministikwia rivers. The southern basin is centred on the valleys of the Whitefish and Kaministikwia rivers. In the northern basin, thick deposits (several tens of metres) of glaciofluvial and glaciolacustrine sediments drape a rugged Archean bedrock surface producing a gently rolling landscape with sporadic bedrock outcrops. The low relief observed within the southern basin is a reflection of not only the flat-lying Proterozoic sediments that underlie the region, but the extremely thick glaciolacustrine deposits that have infilled deep valleys along the Whitefish and Kaministikwia rivers. Overburden thicknesses along these valleys exceed 60 m in places.

The erosional and depositional features preserved within the study area record a complex history of ice flow events associated with the Wisconsinan glaciation. The entire study area was initially affected by southward flowing “Patrician” ice. Striae associated with this event range between 170° and 220°. In the southern half of the Sunshine map area, there is widespread evidence for an older ice flow event towards 160°. The age and significance of this event is not known.

Pauses in the general retreat of the “Patrician” ice lobe resulted in the formation of large, arcuate, recessional moraines in northwestern Ontario. The Brule Creek Moraine, which traverses Aldina and Marks townships adjacent to the Boreal Road, represents one of these stillstand positions. The moraine, which reaches widths of several kilometres in places, is defined by a discontinuous belt of morainic ridges and associated ice-contact stratified deposits. Zoltai (1965) correlates this feature with the Eagle-Finlayson Moraine to the west. The eastern extension of the Brule Creek Moraine beyond Marks Township has been obscured by a younger advance.

A second major moraine of northern affinity (Dog Lake Moraine) is situated in the northeastern corner of the Sunshine map area in the vicinity of Dog Lake. Unlike the Brule Creek Moraine, this moraine was constructed following a minor readvance of the “Patrician” ice lobe. Glaciolacustrine sediments underlying till north of the moraine support a readvance. In the vicinity of Dog Lake, the moraine is fronted by a lobate, flat-topped, glaciofluvial plain believed to be deltaic in origin. The elevation of this plain (460 m asl) defines the level of Glacial Lake Kaministikwia, a large glacial lake which occupied the Kaministikwia basin. Distinctive red clays, associated with this glacial lake can be found throughout the basin up to, but not above, this elevation.

Glacial Lake Kaministikwia was topographically-supported on its southern flanks by the large upland in the northwest corner of the Kakabeka Falls area as well as by the large, northeast-trending Marks moraine that diagonally bisects the map area. This moraine is correlative with the Dog Lake Moraine and represents the terminal position of a lobe of ice originating from the Lake Superior basin that flowed northwestward and westward into the Whitefish-Kaministikwia lowland. Striae associated with this advance range from 340° in the northeast to 240° in the southwest attesting to the lobate nature of the Superior lobe. This advance is correlated with the Marquette stadial, a late-glacial surge event that resulted in much of the Superior basin being reoccupied by glacial ice about 9.9 ka BP. With the exception of the northeast corner of the Kakabeka Falls area, the Superior lobe advanced into a glacial lake, referred to as “Glacial Lake O’Connor” (Zoltai 1963). Numerous sections along the Whitefish River expose thick sequences of glaciolacustrine rhythmites beneath a fine-textured Superior lobe till. Shorelines of this glacial lake have not been identified.

Till of northern (Patrician) provenance has been observed throughout the entire study area. On Archean terrane, beyond the limits of the Superior lobe, the till has a silty sand texture with low to moderate stone-content. Unoxidized till generally has an olive-grey to buff-grey colour. The till often contains a high proportion of clasts that are locally-derived. Faceted and striated clasts within the till indicate deposition by ice flowing towards the south-southwest. Many exposures of this northern till contain remnant clasts of Sibley Group metasediments. These distinctive, reddish-coloured siltstones occur on the Sibley Peninsula and presumably, beneath Lake Superior, 75 km east and southeast of the study area (see Figure 40.1). Their occurrence may indicate an older ice flow event out of the Lake Superior basin, the debris of which has been recycled by northern ice. Alternatively, their occurrence may indicate the presence of, as yet undiscovered, outliers of Sibley Group metasedimentary rocks north of the study area.

The till of northern provenance is buried beneath variable thicknesses of glaciolacustrine deposits and Superior lobe till within the Whitefish-Kaministikwia lowland. The till matrix is generally of a sandy silt to clayey silt texture, owing its fineness to the underlying shales and argillites of the Rove and Gunflint formations. Stony till, of presumed northern provenance, exposed at water level along Whitewood Creek in east-central O’Connor Township, contains abundant striated and faceted boulders indicating ice flow toward the southwest.

Superior Lobe till is easily recognized by its fine-texture, and high proportion of Animikie Group metasedimentary clasts in the pebble fraction. Even in areas underlain by Archean terrane, as occur in the northeast corner of the Kakabeka Falls area, Superior lobe till contains only a small proportion of the underlying rock types in the pebble fraction. The colour of the unweathered till ranges from dark grey, to brown to reddish-brown. Sections of till, containing layers with all 3 colours have been observed in Oliver and Gillies
townships and probably reflect the melting out of debris bands derived from different bedrock sources. Superior lobe till also varies markedly in texture. In Oliver Township, near the community of Murillo, Superior lobe till is silty and charged with shales of the Gunflint Formation. The till in this region is fluted and has been observed to reach thicknesses of 7 to 8 m. Similarly, to the south in Scoble and Gillies townships, Superior Lobe till is silty and charged with clasts of Rove Formation shale and Logan diabase. Along the axis of the Whitefish River, the till is silty to clayey and contains less than 5% clasts. This is primarily due to the thick deposits of glaciolacustrine silts and clays that were overridden and incorporated into the basal debris layers of the ice. It has yet to be determined how well the Superior lobe till reflects local bedrock sources within the study area. Pebble counts, mineralogy of the silt and sand fraction and geochemistry may assist with this determination.

REGIONAL HUMUS AND TILL SAMPLING

A total of 219 humus and 288 c-horizon till samples were collected as part of the regional sampling program. Most samples were collected from the Shebandowan greenstone belt. Sampling density within this area is approximately 1 sample per 3 km². Sample distribution is erratic in places, limited by access as well as the nature of the surficial deposits within various areas. Samples were collected at a much lower density over the southern margin of the Quetico subprovince as well as over areas underlain by Archean felsic intrusive rocks and Proterozoic metasedimentary rocks.

Samples were collected primarily within areas accessible by roads and bush trails. A short, helicopter-assisted sampling program was undertaken to obtain samples from less accessible areas and to fill in the regional sampling grid. Till was collected primarily from shallow (~1 m deep), hand-dug test pits, roadcuts, trench exposures and small borrow pits. Samples were collected from the till-bedrock interface where possible, or from as deep as possible in areas of thick till. Samples were screened in the field to remove the +5 mm fraction; this fraction was saved for lithologic determinations. At each sample site, 2 samples of horizon till were collected for analysis: 1) a 10 kg sample for heavy mineral concentration and subsequent gold grain, kimblerite indicator mineral (KIM) and metamorphic massive sulphide indicator mineral (MMSIM) determinations; and 2) a 2 to 3 kg sample for -63 mm fraction geochemical determinations. Humus samples were collected from all undisturbed sample sites as well for geochemical analysis.

High density soil sampling was undertaken over and adjacent to the Bylund occurrence in Dawson Road Lots. Fifty-three soil samples were collected. A sample was taken every 100 m on grid lines spaced 100 m apart. This orientation survey was undertaken in an attempt to characterize and define the magnitude and shape of dispersal trains originating from a known gold source. Mineralization at the Bylund occurrence is contained within a 125 m wide, east-trending zone of carbonate-altered, mafic to ultramafic metavolcanic rocks truncated by a stockwork of northeast-trending quartz-carbonate veins (Schneiders et al. 1998). Channel sampling over the main occurrences have returned values of up to 4.1 g Au over 14 m.

CONMEE TOWNSHIP FLOAT OCCURRENCE

Mineralized float was encountered frequently as part of the regular mapping and sampling program. Many of these were collected and submitted for assay. Of particular interest was the discovery of several large boulders of sulphidized iron formation within a gravel pit in Conmee Township, on the west side of Highway 11/17, 600 m north of the 3rd Concession. One of the boulders measured over 1 m in diameter and consisted of massive pyrite and magnetite with 10 to 15% sphalerite disseminated in pyrite-rich sections. Sphalerite was also concentrated along fractures and adjacent to quartz veinlets throughout the rock. Nodules of pyrite up to 1 to 2 cm in diameter were observed in some of the finer-grained zones of the rock suggesting a primary rather than secondary origin for the pyrite.

Two samples from the pyrite-rich zones returned values of: 1) 5.13% Zn, 18 ppm Cu, 19 ppm Pb, 260 ppb Au and 0.5 ppm Ag; and 2) 2.85% Zn, 16 ppm Cu, 20 ppm Pb, 245 ppb Au and 0.5 ppm Ag. A sample from the magnetite-rich zone returned values of 850 ppm Zn, 25 ppm Cu, 5 ppm Pb, 25 ppb Au and less than 0.2 ppm Ag. A second sulphidized iron formation boulder from the same pit, measuring approximately 0.5 m in diameter and consisting almost exclusively of pyrite, returned values of 140 ppm Zn, 8 ppm Cu, 11 ppm Pb, 710 ppb Au and less than 0.2 ppm Ag.

The boulders were found in a deposit of ice-contact stratified drift derived from the Superior lobe. Streaks in the immediate vicinity of the pit are oriented at 320 to 330°. Assuming that the boulders are derived from an Archean source, it may be that the boulders were either eroded from a source less than 3 km southeast of the pit (i.e., Proterozoic rocks are present beyond this point) or that the boulders were initially eroded from a source north-northeast of the pit by "Patricia" ice then remobilized by the Superior lobe.

Exploration work during the early 1900s along the lower reaches of Brule Creek, 4 to 5 km north-northeast of the gravel pit, by B.L. Morrison, the Davis Sulphur Company and General Chemical Company resulted in the discovery of "seven lenticular masses of brecciated, banded iron formation, in which pyrite has replaced a considerable part of the rock" (Carter 1990). The largest of these masses has a maximum width of 23 m and is 244 m long. Other discoveries include a 21 m wide body of pyrite containing magnetite and pyrrhotite and a 9 m wide by 15 m long zone of magnetite-pyrite-jasper ironstone. It is possible that the boulders found within
the gravel pit were derived from this area and that sphalerite was not recognized in the rock. It is not yet clear whether the sulphides indicate proximity to a VMS style zone of mineralization. Further work is required to assess the mineral potential of this area.

REFERENCES


INTRODUCTION

Kimberlite is recognized as the primary host rock for diamond. A suite of heavy minerals (chrome pyrope garnet, chromite, magnesium-rich ilmenite, chrome diopside) is commonly associated with kimberlite and are referred to as kimberlite indicator minerals (KIMs). A Quaternary geology mapping and modern alluvium sampling program conducted by the Ontario Geological Survey (Morris 1990, 1991, 1992a, 1992b, 1994a, 1994b, 1995, 1996, 1997a; Morris et al. 1994a; Morris et al. 1997) in the Wawa area, northeastern Ontario, was successful in defining the types and distribution of KIMs. The Kapuskasing Structural Zone (KSZ) underlies the Wawa region and is thought to be a suitable structure for hosting kimberlite (Boland and Ellis 1989). The KSZ extends northeast from Wawa through the Kapuskasing area and into James Bay (Figure 41.1).

During the summer of 1997, Quaternary geology mapping and a modern alluvium sampling program was completed in the Woman Falls–Wakusimi River area (see Figure 41.1). This project was successful in establishing the occurrence of KIMs, magmatic massive sulphide indicator minerals (MMSIMs®; Registered trademark of Overburden Drilling Management Limited, Nepean, Ontario) and gold grains within this area (Morris 1997b, Morris 1998, Morris et al. 1998). The relative proximity to source of indicator minerals was determined through evaluation of the pebble component of samples and limited geophysical data. A relationship between the occurrence of KIMs and bedrock structure was also established. From this work, several exploration targets for kimberlite and base metals were determined.

In the summer of 1998, modern alluvium sampling and overburden mapping was extended west to the Opataska–Roche Lake area (see Figure 41.1). This was done to determine if the occurrence of KIMs and their relationship to bedrock structure continued west across the Kapuskasing Structural Zone. Modern alluvium was the principal material sampled. A few coarse-grained glaciolacustrine samples were collected for reference. In addition, a limited number of till samples were collected to establish glacial dispersal characteristics from a mafic-ultramafic pluton.

The KIMs, MMSIMs and gold grains are being picked from heavy mineral concentrates. The MMSIMs (anthophyllite, chalcopryite, chromite, gahnite, hypersthene, olivine, red epidote, red rutile, ruby corundum, sapphireine, spessartine, staurolitie) are indicators of polymetallic deposits associated with migmatized terrain. Also, gahnite is used to identify rare element pegmatites.

Quaternary geology mapping provides a framework for understanding the distribution of the data derived from the overburden materials. In addition, mapping may identify aggregate resources critical in the development of road infrastructure for the local forest industry.

The Opataska Lake–Roche Lake area is represented on two, 1:50 000 scale National Topographic Series maps. The western part of the study area is covered by the Roche Lake (42G/4) sheet and the eastern part by the Opataska Lake (42G/3) sheet. The area is bounded by longitudes 84°00’ W and 83°00’ W and latitudes 49°00’ N and 49°15’ N.

PHYSIOGRAPHY

The study area is located north of the Great Lakes–Hudson Bay drainage divide and all surface drainage flows north
into James Bay. There are 4 major drainage basins within the area. Flow from these basins is through the Mattawichiwe-wan, Goat, Missinaibi and Opasatika rivers.

The Opasatika Lake–Roche Lake area is part of the Abitibi Uplands subregion of the James Physiographic region (Bostock 1976). This region, and all of the study area, is underlain by crystalline Archean rocks and has a broad rolling surface that rises gently from the Hudson Bay Lowland.

Much of the topography in the Opasatika Lake area is subtle although local relief is up to 45 m. In the Roche Lake area the topography is more pronounced with local relief of up to 120 m. Bedrock outcrops occur most commonly in the Roche Lake area and along parts of the Missinaibi and Brunswick rivers.

**BEDROCK GEOLOGY**

The bedrock geology of the area consists of 7 bedrock types (OGS 1991). These include: 1) muscovite-bearing granitic rocks including muscovite-biotite and cordierite-biotite granites, granodiorite and tonalite; 2) a foliated tonalite suite consisting of tonalite to granodiorite, foliated to massive; 3) a gneissic tonalite suite consisting of tonalite to granodiorite, with minor supracrustal inclusions; 4) mafic and ultramafic rocks including gabbro and anorthosite; 5) metasedimentary rocks including wacke, arkose, argillite, slate, marble, chert, iron formation and minor metavolcanic rocks; 6) felsic to intermediate metavolcanic rocks including rhyolitic, rhyodacitic, dacitic and andesitic flows, tuffs and breccias, chert, iron formation, minor metasedimentary and intrusive rocks and related migmatites; and 7) mafic to intermediate metavolcanic rocks including basaltic and andesitic flows, tuffs and breccias, chert, iron formation, minor metasedimentary and intrusive rocks, and related migmatites. All units are cut by northwest trending diabase dikes of the Matachewan and Hearst swarms.

The Rufus Lake Fault strikes northeast across the eastern part of the map area. Many of the lake and stream orientations appear to be structurally controlled.

**QUATERNARY GEOLOGY**

All glacial landforms and materials in the map area were likely formed and deposited during the Wisconsin glaciation. Striae, grooves and drumlinoid features define ice flow directions. Three sets of striae were observed: 110° to 162°, 170° to 195°, and 200° to 260°. Striae crossing relationships on 2 outcrops clearly demonstrate that striae oriented southeast (110° to 162°) are the youngest. In the eastern part of the study area, one outcrop surface consisted of striae oriented 135° crossing striae oriented 170° to 195°. On a second outcrop, in the central part of the map area, striae oriented 144° crosscut striae oriented at 210°. The 110° to 162° set of striae is the most prevalent and was observed on numerous outcrops throughout the area. The relative age relationship between the 170° to 195° and 200° to 260° flow events is not clear. Similar striae observations were made to the east in the 1997 survey (Morris 1997b).

Observations on the types and distribution of surficial materials were made at each sample site. These observations are being integrated with airphoto interpretation to establish the types and distribution of overburden materials and landforms associated with glacial and post-glacial events in the area. In turn, these observations will be combined with those from the 1997 work to produce a composite overburden map and accompanying Quaternary geology report.

**PROJECT STATUS**

A total of 123 modern alluvium, 25 till and 11 coarse-grained glaciolacustrine samples were collected during the 1998 survey. Modern alluvium was collected from as many streams and rivers as possible. From this, a good distribution of samples was collected from the entire study area. Till sampling was completed in an area underlain by a mafic-ultramafic pluton. Till samples were collected up-ice, over and down-ice of this pluton to establish glacial dispersal characteristics for the area. Coarse-grained material was collected from as many gravel pits as could be reached.

Currently, modern alluvium and till pebble lithologies are being determined and interpreted. All samples have been submitted to an external lab to isolate any KIMs, MMSIMs and gold grains. This data will be interpreted and presented in a report. Airphotos are presently being interpreted with the aid of field notes. The Quaternary (surficial) map produced from this work will highlight any potential aggregate resources within this area and area mapped in the summer of 1997.

**REFERENCES**


——— 1992b. Quaternary geology of the Wawa area; Ontario Geological Survey, Open File Map 192; scale 1:50 000.


—— 1997a. Results of modern alluvium sampling for kimberlite indicator minerals and garnet, Kinniwabi Lake area, northeastern Ontario; Ontario Geological Survey, Miscellaneous Release-Data 23.


INTRODUCTION

Fieldwork for a high density lake sediment and water geochemical survey of the Garden Lake–Obonga Lake area was carried out between July 6 and August 9, 1998. The survey area is located approximately 130 km north of Thunder Bay (Figure 42.1). The survey completely covered the area defined by National Topographic System (NTS) 1:50 000 scale map sheets 52G/9, 52H/12, 52H/13 and partially covered the area outlined by map sheets 52G/8, 52H/5, 52H/6, 52H/11 and 52H/14.

Lake sediment and/or water samples were collected at 2120 sites for an average of 1 sample per 2.2 km². Stream/river sediments and/or waters were collected from 45 sites. In addition, 15 rock samples were collected and submitted for geochemical analysis.

The Garden Lake area was selected for this type of geochemical survey for several reasons. These include, client interest in the area, a relative deficiency of exploration data for the region and recent speculation that felsic rocks of the same age as those of the Sturgeon Lake belt may occur in the Obonga–Garden Lake area (Tomlinson et al. 1996).

Some of the lakes in the southern half of the study area were sampled during the national geochemical reconnaissance (NGR) lake sediment program that was undertaken during the late 1970s by the Geological Survey of Canada.
BEDROCK GEOLOGY

The most recent bedrock mapping and data compilation of the entire study area was completed by Sage et al. (1974) and Sage (1998). The Garden Lake area was mapped in detail by Milne (1964). Mapping of the Obonga Lake area was completed by Thurston (1967) and Kustra (1966). The Metionga Lake area, in the northwestern portion of the study area, was mapped by Rogers (1964). Refinements and re-evaluations of this early work, augmented by additional lithochemical analysis, has occurred in recent years in an effort to unravel the age and genesis of the supracrustal rocks in the region (e.g., Blackburn et al. 1991, Cortis et al. 1988, Thurston et al. 1987, Tomlinson et al. 1996).

The Garden Lake greystone belt strikes easterly for a distance of approximately 75 km, from Alley Lake in the west to the Gull River in the east (Figure 42.2). The widest portion of the belt lies between the Gull River and Weaver Lake. Here the geology consists of predominantly mafic metatrendic rocks with a thin core of metasediments between Garden Lake and Kears Lake (Sage 1998). Flat lying diabase occurs locally and covers the eastern end of the greystone belt. The greystone belt west of Weaver Lake is thin but reaches a thickness of approximately 5 km at Alley Lake, at the extreme western end of the belt. The predominant rock type in this western portion of the belt is mafic metavolcanic rocks, although outcrop exposure is poor (Sage 1998). A mafic intrusive body occurs at Alley Lake and another at Twining Lake, located to the southwest of the greystone belt at the edge of the survey area (Sage 1998).

Milne (1964) noted a zone of particularly intense shearing within the central part of the belt. Sage (1998) also reported evidence of strong tectonic activity along the granite-greystone boundaries of the Garden Lake belt. Thurston et al. (1987) identified the structure along the core the belt as the Central Garden Lake Shear.

Gold was first discovered during the 1930s in quartz veins east of Conick Lake, near the northern granite-greystone boundary (Milne 1964). Thereafter, sporadic prospecting/exploration activity in the Garden Lake belt has concentrated on gold and associated sulphide occurrences. For example, gold and pyrite in sheared, silicified and carbonatized metavolcanic rocks were reported by Little Long Lac Gold Mines Ltd. in 1946 along the southeast shore of Garden Lake (Milne 1964).

The Obonga Lake greystone belt lies north of the Garden Lake greystone belt and is approximately 32 km long. It consists of an east striking synclinorally folded sequence of mafic and felsic metavolcanic rocks, metasedimentary rocks and mafic to ultramafic intrusions (Sage 1998). The eastern end of the belt is overlain by Proterozoic rocks, predominantly flat lying diabase. A thickening of felsic volcanic rocks occurs toward the western end of the belt, these are thicker in the area north of Tommyhow Lake (Sage 1998). Along the northern margin of the belt, in the vicinity of Chrome and Puddy lakes, a serpentinite body is present.

Intermittent exploration efforts in the Obonga greystone belt have focussed on the PGE-chrome-base metal potential of the mafic-ultramafic intrusive rocks and the base metal potential of the felsic volcanic rocks near Tommyhow Lake. A chromite mine was operated briefly during the 1930s by Chromium Mining and Smelting Co. Ltd. at Chrome Lake (Sage 1998). More recently, Houston Lake Mining Inc. has acquired mining claims at Puddy Lake with the intention of exploring for PGMs (platinum group minerals).

Previous exploration in the Tommyhow Lake area has included airborne geophysics and diamond drilling. Massive and disseminated pyrite has been reported (Sage 1998) but apparently no economic base metals have been found to date. It is not known if the area has been examined for precious metal (gold) potential.

PHYSIOGRAPHY AND QUATERNARY GEOLOGY

Gentle topography characterizes most of the study area with local relief rarely exceeding 50 m. The most rugged topography occurs in the east portion, and northeast part of the study area, where the bedrock is dominated by flat lying Proterozoic diabase. Spectacular cliffs and talus slopes are common in the area near Obonga Lake.

Systematic, detailed Quaternary geological mapping has not been conducted over the study area. Reconnaissance Quaternary mapping (1:506 880) was completed by Zoltai (1965). Engineering geology terrain maps, at a scale of 1:100 000, were produced for the region between 1980 and 1983 (Molland and Molland 1980a, 1980b, 1981, 1983). A more recent regional compilation of the Quaternary geology of the area has been completed by Barnett et al. (1991). These sources indicate that surficial materials within the study area consist predominately of a thin discontinuous veneer of drift (till) over bedrock. Glacio-fluvial deposits occur locally, most notably in the south and southeast of the region surveyed, where short sections of the Lac Seul, Hartmann and Katash moraines cross into the survey area. A short section of the Sioux Lookout moraine is present in the northwestern corner of the survey area. A number of north trending eskers, up to 20 km long, occur in the southern half of the study area. This north to south esker orientation corresponds with the general direction of ice flow retreat from the area.

Modern fluvial deposits occupy major river valleys, such as along the Mooseland, Roaring and Gull rivers. Glaciolacustrine sand, silt and clay occur locally, increasing in extent toward Lake Nipigon in the east.

SAMPLING METHODS

Organic lake sediment samples were collected from a helicopter float using the OGS designed gravity corer. On average, 24 lakes were sampled per hour of helicopter time. In order to avoid anthropogenic influences and
water/sediment interface effects (i.e., increased Mn due to anoxic conditions that result in secondary accumulation of base metals), only deep sediment (>20 cm below the sediment surface) was collected. This sediment better reflects the effects of natural geochemical inputs that may be traced to local geology.

Lake water samples were collected at a depth of 0.5 m from shallow lakes (<3 m deep) and at a depth of 2 m from deep lakes (>3 m). A semi-automated water sampling apparatus, developed by the OGS, was utilized for water collection. The apparatus consists of a submersible pump, a flow cell (for measurement of parameters such as pH, conductivity, oxidation-reduction potential and dissolved oxygen), a sample bottle tray and various hoses and pinch valves. Water is pumped from the lake and allowed to purge the sampling system prior to the collection of a water sample and the recording of water quality parameters. Water samples were kept cool after collection and processed (filtered and acidified) within 6 hours of collection.

River/stream sediment samples were collected by shovel or with a home made scoop fashioned out of a plastic bottle. In some cases, river sediment was collected with the lake sediment sampler from the helicopter float. Where streams crossed a road, the sample(s) were taken approximately 25 to 50 m upstream of the road. On narrow, low energy streams, sediment was collected from the stream bottom and typically had a high organic component. In high energy streams/rivers, samples were collected from several locations including the lee of large boulders, quiet pools (at the tail of bars or outside portion of river bends) and the top surface of gravel bars. In all
cases, an attempt was made to sample active, inorganic, silt/sand sediment.

A GPS receiver was utilized to record accurate sample site positions. A customized database application operated on a pen based computer was used to record sample descriptions and observations. Data from the water quality probe was also monitored and recorded with the pen based computer.

**SAMPLE PREPARATION AND ANALYTICAL METHODS**

Lake sediment samples were placed in breathable fabric bags and allowed to partially air dry prior to shipment to the laboratory. The samples were then freeze dried (to retain volatile elements), partially pulverized in a ceramic ring and puck pulverizer and sieved to obtain the -80 mesh (<177 µm) size fraction. Laboratory analysis includes nitric-aqua regia digestion followed by inductively coupled plasma-mass spectrometry (ICP-MS) to determine approximately 50 trace elements. Mercury is determined by cold vapour-flameless atomic absorption spectrometry. Nitric acid-aqua regia digestion attacks all sample matrix constituents, except for silicate minerals, and therefore is considered a nonselective, relatively strong partial extractant.

Approximately 15 g of sample pulp are pressed into briquettes prior to analysis by instrumental neutron activation analysis (INAA) for Au and a suite of 34 other elements. Quality control will be monitored through the use of sample pulp duplicates and certified reference materials. Loss-on-ignition (LOI) is determined at 500°C, using an automated gravimetric technique.

Water samples are passed through 0.45 mm syringe filters and acidified to 1% ultrapure nitric acid within 6 hours of collection. Analysis of water includes direct aspiration ICP-MS to determine approximately 50 elements including major cation and anion species. Quality of the analyses is monitored through the use of sample duplicates, certified reference standard SLRS-3 and blanks.

Stream sediment samples were placed in breathable fabric bags after collection and air dried. Samples are gently disaggregated and then sieved to retain the -80 mesh size fraction for INAA and ICP-MS analysis. A distinction between organic and inorganic based sediments is necessary for both INAA and ICP preparation (briquettes versus vial encapsulation; partial versus total digestion). This distinction is maintained during data interpretation.

Several samples of inorganic river sediment, weighing between 5 and 10 kg, will be sent for analysis for gold grains, heavy mineral concentrate and kimberlite indicator minerals.

**Preliminary Results**

During the course of helicopter lake sediment sampling a broad (approximately 20 to 25 m wide) gossanous shear zone was observed in the Tommyhow Lake area (see Figure 42.2). The outcrop consists of sheared and altered felsic volcanic rocks or silicic metasedimentary rocks (quartz-sericite schist), mineralized locally with up to 3% pyrite and trace chalcopyrite (as evidenced by minor malachite staining). Some quartz-carbonate veining with trace pyrite was also observed. Although traces of previous work (old flagging tape) was present, it is not known if this site was ever examined for gold potential. This gossan and sulphide occurrence is not shown on the geology maps available for the area (i.e., Thurston 1968, Sage et al. 1974, Pye and BaiUie 1965). Nine bedrock samples were collected from this zone for analysis.

During the course of river sediment sampling on the Mooselander River (see Figure 42.2), a strongly sheared and carbonatized mafic volcanic bedrock outcrop containing trace pyrite was observed. This may be evidence of an eastward continuation of the Central Garden Lake shear along the core of the greenstone belt. Numerous pieces of quartz-carbonate vein float were also found along the river bed. This vein material contained approximately 5% hematite and trace pyrite. Samples of each of these materials have been sent for analysis.

As of September 1998, much of the ground in the Garden–Obonga Lake study area, including the locations mentioned above, were open for staking.

**REFERENCES**


Thurston, P.C. 1968. Kershaw Lake; Ontario Department of Mines, Map P.458, scale 1 inch to 1/4 mile.


43. Project Unit 98–59. Field Data in Support of Electrochemical Transport of Elements Through Thick Glacial Overburden

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INTRODUCTION

This article describes fieldwork carried out during 1998 that was part of ongoing investigations by the OGS into geochemical exploration techniques that can be applied in thick overburden areas (Jackson 1995; Hamilton et al. 1995; Hamilton et al. 1997; Baje 1998; Hamilton, 1998). Geochemical methods capable of penetrating thick glacial overburden would be a useful advent for explorationists in Ontario because of the difficulties faced in many parts of the province arising from thick drift.

The 1998 work involved self potential surveys, limited geochemical sampling and a small drilling program at the Shoot Zone, a gold property in Taylor Township owned by St. Andrew Goldfields Ltd. Self potential surveys were also completed at Victoria Creek, a gold property near Kirkland Lake owned by Sudbury Contact Mines. This fieldwork follows extensive geochemical sampling at both sites and a drilling program at the Shoot Zone.

The purpose of the 1998 fieldwork was to field-test a theoretical model proposed by Hamilton (1998) that infers the presence of an electrochemical “cell” in overburden that could mobilise and transport metals relatively quickly through thick (>20 m), young (8 to 10 Ka) glacial overburden. The model was advanced to explain the presence of surficial geochemical anomalies spatially related to mineralisation known to exist at these two sites and others.

BACKGROUND

Researchers from the OGS (Jackson 1995; Baje 1998) and other authors (e.g., McClenaghan et al. 1997, Gleeson et al. 1988, Clark 1993, M. Fedikow, Manitoba Energy and Mines, personal communication, 1997) have reported the presence of surficial geochemical anomalies spatially related to mineralization or other bedrock features in areas of thick glacial overburden. The genesis of these anomalies is problematic because of the young age and thick nature of the glacial deposits. Any model that attributes surficial geochemical anomalies in soil to the presence of underlying mineralization must involve an element dispersion mechanism that could operate in the 8000 to 10,000 years since glaciation ended.

Mechanisms such as groundwater transport, chemical diffusion and transport by gaseous carriers have been proposed by others to explain the presence of surficial geochemical anomalies in exotic overburden overlying mineralization. These were investigated by the OGS and ruled-out as major contributors to the formation of geochemical anomalies over mineralization at the Shoot Zone. Given the thickness and hydro-stratigraphy of the overburden, both groundwater transport and chemical diffusion are too slow to have penetrated overburden in the time period since deposition. The overburden at the Shoot Zone is continuously saturated to ground surface and as a result gaseous diffusion would require prohibitively large quantities of gas to maintain a separate gas phase below the water table. A detailed groundwater and overburden stratigraphic investigation has never been carried out at Victoria Creek but, because of the similarities in the morphologies of the anomalies at the 2 sites, a similar transport mechanism has been inferred.

Many authors (e.g., Govett 1973, 1976; Bolviken and Logn 1975; Thornber 1975; Smee 1983; Govett et al. 1984) have attributed the development of surface geochemical anomalies overlying mineralization to electrochemical processes. However, the mechanisms they proposed cannot explain anomalies related to mineralization in young, thick, glacial overburden as occurs at the Shoot Zone and Victoria Creek. The OGS carried out an extensive theoretical investigation into existing models and the principles of electrochemical transport in a groundwater electrolyte and advanced a new theoretical model (Hamilton, 1998) that can potentially account for the presence of the observed anomalies. The purpose of the 1998 fieldwork was to test this model.

THEORY

This section presents a very brief account of the theory being investigated as part of the 1998 fieldwork. For a more detailed treatment see Hamilton (1998).

The presence of an electronic conductor such as graphite or metallic sulphides in bedrock provides a shortcircuit route for electrons from reducing agents at depth to react with more abundant oxidizing agents in basal overburden (Sato and Mooney 1960). This voltage differential (self potential) between the top and bottom results in the flow of electrical current in the conductor. The upward movement of electrons is followed by the
consumption of oxidizing agents in overburden overlying mineralization. A stronger redox gradient exists, therefore, between mineralization and ground surface than exists between bedrock and ground surface in surrounding areas. This high electrochemical gradient results in the rapid outward and upward migration of reduced, negative charge carrying species into groundwater saturated overburden at a rate that far exceeds that of simple chemical diffusion. Simultaneous inward and downward dispersion of positive charge carrying species prevents macroscopic charge imbalances in the groundwater solution. Over time, the outward migration of reduced species will result in the upward propagation of the negative redox anomaly above bedrock surface. A reduced “column” should therefore develop in groundwater saturated overburden immediately overlying mineralization.

The reduced column or “cell” (Figure 43.1) should influence the movement of all charged species in its vicinity. This mechanism alone should result in geochemical anomalies in near-surface soils as mobile ions move in response to it. Relatively rapid transport of charged species between mineralization and ground surface should also be possible.

**SITE CONDITIONS**

For the most part, the Shoot Zone ore body occurs in an approximately 45° south-dipping metasedimentary package that subcrops parallel to and 75 m from subcrop of the similarly dipping Porcupine–Destor Fault Zone. The metasedimentary package comprises interbedded arkose and graphitic argillite and the subcrop varies in surface width between 15 and 30 m. The hanging wall and footwall rocks consist of green-carbonate (fuchsitic) altered ultramafic flows. Results from surface electromagnetic (EM) surveys show the mineralized zone to be a moderately strong conductor (K.A. Jensen, St. Andrew Goldfields, personal communication, 1997). Microscopic gold is dispersed throughout the sediments and occasionally in the adjacent green-carbonate. Visible gold is rare.

Hamilton et al. (1995) and Bajc (1998) describe the surface and subsurface Quaternary geology of the Shoot Zone. Mineralization occurs under the western flank of the south-trending Taylor Esker. The sand unit overlying till is a wedge-shaped glaciolacustrine deposit that thickens to the east, toward the esker. Both the clay and the till thin, and effectively pinch out at the esker to the east and thicken, relative to the sand, to the west. A peat bog

![Figure 43.1](image-url)
overlies the mineralization on the west flank of the esker. In the bog, ground is fully saturated to within 30 cm of the surface. Geophysical line 1000W (discussed below) is located entirely within the bog. Sand is exposed on surface to the east and clay on surface to the west of the bog.

The mineralized zones at Victoria Creek occur largely in 64° north-dipping felsic volcaniclastic rocks flanked to the north and south by felsic to intermediate volcanicastic rocks. These units form part of the Gauthier Group, which is in disconformable contact with underlying Kinlochewan mafic volcanic rocks approximately 150 m north of the centre of the gold mineralized zones. The mineralized zones and contacts between the major units strike roughly east. The majority of the ore is confined to the felsic pyroclastic unit although mineralization also exists in the felsic to intermediate units to the south and north. The gold is associated with pyrite. The projected subcrop of the higher-grade mineralized zone is approximately 40 m wide but lower-grade lenses extend the subcrop zone to approximately 150 m. The felsic rocks in the gold-mineralized zones contain galena, chalcopyrite and sphalerite. Graphite tends to be concentrated in the lower part of the mineralized zones and results in a weak to moderate EM anomaly. The zone is a strong induced polarization (IP) target.

The Victoria Creek study area also occurs under the flank of an esker, in this case the east flank of the Munro Esker, and mineralization is overlain by about 30 m of overburden. The overburden material, however, is nearly all sand overlying up to 2 m of till. Up to 1.5 m of peat and/or living vegetation was encountered during probing of the swamp in 1997 and in most cases sand was encountered beneath this. Large parts of the bog comprise a floating vegetative mat above up to 1 m of water. The bog is not treed. Geophysical lines 200E and 500E (discussed below) overlie mostly dry, sandy soils whereas line 315E overlies the bog except at the extreme north end of the line.

FIELDWORK

The object of the fieldwork was to allow the collection of both vertical and horizontal redox and self potential (SP) data along a line transecting mineralization. Surface SP surveys were run to determine if any surface evidence exists for an electrically reduced area over mineralization. A small drilling program was carried out at the Shoot Zone in order to allow testing for the presence of a reduced column in groundwater saturated overburden between mineralization and ground surface.

Five approaches were used to acquire SP and redox data:

1. Installation of platinum (Pt) tipped wires at multiple levels in boreholes. The object was to provide a series of points between which voltage differences could be measured against a Cu-CuSO₄ surface electrode.

2. Lowering of a platinum-tipped wire down plastic cased monitoring wells. This was used on the centre PVC pipes in the boreholes drilled this season to measure SP against a Cu-CuSO₄ surface electrode. This technique was also used in 10 of the existing 18 plastic multi-level monitoring wells installed in 1995.

3. Surface SP surveys using a millivoltmeter, standard electrodes, a 250 m reel of wire and a central base station.

4. The N₂ glove-box measurement of Eh in fine-grained sediment samples collected with a split-spoon during the drilling.

5. The extraction of groundwater from monitoring wells and the measurement of groundwater Eh.

Drilling

A CME track mounted geotechnical drill rig was used to drill 3 boreholes through overburden and the upper few metres of bedrock. Hollow-stem augers (4" ID) were used to penetrate most of the sediments but where this became impossible due to boulders, NW-sized casing was used. Split-spoon samples of overburden materials were taken every 1.5 to 3 m depending on the material. Samples were split open and immediately placed inside a N₂ filled glove box for the measurement of redox and pH. Only fine-grained (clay and silt) samples could be measured in this fashion. NQ-sized core was taken in bedrock to a depth of 3 to 6 m.

Of the 3 holes that reached rock, only the first, 98-1 (Figure 43.2), encountered mineralization and this penetrated only the lowermost 15 cm of graphitic argillite before encountering footwall green-carbonate rocks. Borehole 98-2 (see Figure 43.2) was drilled about 15 m to the north of 98-1 and also penetrated footwall rocks. Borehole 98-4 was drilled 110 m to the east and also failed to intersect mineralization but based on SP readings, and lithological information is likely to have encountered footwall rocks near to mineralization. Borehole 98-3 was drilled 45 m north of TW-1 and split-spoon samples were taken of the clay that was encountered below the surface peat and above the sand aquifer. This hole was discontinued at the base of the clay layer.

In 98-1, 98-2 and 98-4, a 1" (ID) rigid PVC monitoring well with internal threaded couplings was installed in bedrock. A seal between bedrock and overburden was established using an NQ-sized self-setting, bentonite-filled, gum-rubber packer on the end of each pipe and below this the hole in bedrock was essentially left open (i.e., no filter-pack). Additional groundwater sampling points were added in 98-1 and 98-4 by the attachment of ½" (ID) low-density polyethylene tubes to the monitoring well. The lower end was slotted and wrapped in nylon mesh to provide a sampling point. An additional 5 points were added in 98-1 and 2 in 98-4. In both 98-1 and 98-4, bundles of 5 wires with platinum-tipped electrodes were permanently attached to the monitoring well at various intervals and left exposed to the groundwater environment.

After installation of the wells, they were purged using a jet of N₂ gas. Wells were sampled to the extent possible using a peristaltic pump. Low flow rates and the 10 m negative pressure maximum of peristaltic pumps limited the amount of sample recovered.
SP Survey Program

The SP surveys were carried out using standard SP electrodes (Burr 1982) that consist of a copper rod in a porous pot filled with a saturated CuSO₄ solution. On each survey, a base station was selected near the middle of the area to be surveyed, a stationary pot was placed at the base station and a 250 m wire was drawn out first in one direction and then in the opposite. A millivoltmeter was connected between the stationary pot (negative) and the mobile pot (positive). With this arrangement, negative readings are indicative of Eh conditions around the mobile pot that are more chemically reducing than those around the base station, and conversely, positive readings are indicative of more oxidizing conditions around the mobile pot. Duplicate readings were taken at each station; the first as the wire was being drawn out and its duplicate as the wire was reeled back. Stations were located 10 or 12.5 m apart depending upon whether the picket spacing was 20 or 25 m.

Three sources of error were repeatedly encountered. One occurred due to variable moisture content of the soil and pertained only to SP readings taken in swampy peat terrain. Variations of up to 30 mV were recorded in a radius of less than 2 m when the electrode was moved from wet (visible water) spots to dry hummocks or spruce root

![Shallow Groundwater Geochemistry - Line 1000 W](image)

**Figure 43.2.** Self potential voltages in overburden versus shallow groundwater geochemistry – Shoot Zone, line 1000W. Data from wells 95-10, 95-02 and 95-16 were projected eastward along strike to line 1000W distances of 140, 70 and 70 m respectively. All other boreholes are located on the line.
mounds. Errors due to this problem could be minimised by careful site selection during the survey. A second source of error, perhaps related to the first, occurred due to locating stations in bare, dry areas such as drill-pads or roads where the moist humus layer had been stripped away leaving only B-horizon soils. This error could not always be prevented but it was accounted for after the fact by using data from a short parallel line run 10 to 15 m away in undisturbed areas and comparing, and if necessary, correcting the response obtained over the disturbed area. Positive errors up to 60 mV were noted due to removal of the A-horizon. The final source of error was by far the most significant and occurred due to the presence of metallic drill-casing penetrating overburden. Negative variations of up to 1000 mV were noted in the immediate vicinity of drill casing. Experimentation determined that the area of influence of 45° dipping drill casing was approximately 4 m in a down-dip direction and 3 m in an up-dip direction in sand and about 1 m in both directions in clay. If survey points were selected beyond this, they did not appear to be influenced by the presence of drill casing. If points were inadvertently collected near boreholes they were removed from the data prior to plotting.

At the Shoot Zone 10 SP lines were run; 7 at right angles to the strike of mineralization and three approximately parallel to it. The results are complicated by 3 different terrain types; peat, sand and clay, which clearly have different background SP. The results of this survey are currently being analyzed and are not a subject of discussion herein.

At Victoria Creek, SP surveys were run over 3 lines parallel to mineralization and 1 line perpendicular to it. Only 2 terrain types were encountered here, peat and sand, and although they too had different background SP, the data correlate well with geochemical data. Selected results are presented below in the context of the earlier geochemical data from the same lines.

Down-hole SP was measured in 10 of the existing multi-level monitoring wells and the new wells 98-1, 2 and 4. A single station was selected on surface and a 250 cm insulated wire was stretched out from this base to each borehole. For overburden measurements in 98-1 and 98-4 the wire was successively connected to each of the down-hole wires leading to the permanent Pt electrodes. For shallow bedrock measurements in 98-1, 2 and 4 and for all measurements in the 10 existing wells, a separate Pt-tipped wire was reeled down each well into the well-screen and connected to the base station wire.

The geochemical data shown below for Victoria Creek are for humus samples collected in 1993 on lines 200E and 500E, and for shallow groundwater samples collected in 1997 on line 315E. Both sets of samples were analyzed at the Geological Survey of Canada Laboratories in Ottawa. For the humus elements discussed, Cu and Ni were analyzed by ICP-ES aqua regia; Cr, LREEs (light rare earth elements) and Ti were analyzed by ICP-MS aqua regia; and Sr was analyzed by INAA. All the elements discussed for groundwater were analyzed by ICP-MS except Cr, which was analyzed by ICP-ES.

PRELIMINARY RESULTS AND DISCUSSION

The following is a presentation of selected preliminary results of the drilling and SP programs. Considerably more data were gathered and will be published at a later date.

The drilling showed stratigraphy in the 4 boreholes very similar to the stratigraphic interpretations from the 1995 drilling program (Hamilton et al. 1995; Bajc 1998). Boreholes 98-1, 2 and 3 were drilled in the swamp and encountered 3 m of peat overlying 8.4 to 10.7 m of clay. Boreholes 98-1 and 2 were continued to bedrock encountering respectively, 11.7 and 9.1 m of sand overlying 5.0 and 7.0 m of till which directly overlies bedrock at depths of 28.2 and 29.9 m. Borehole 98-1 penetrated 15 cm of the graphitic argillite that hosts gold mineralization and this was first presumed to be a small boulder. Below this, 2.1 m of footwall fuchsite “green carbonate” metamorphosed ultramafic rocks were drilled. Borehole 98-2 penetrated 2.0 m of green carbonate bedrock. Borehole 98-4 was drilled over the esker 110 m east of 98-1 and encountered 27.7 m of sand overlying 2.5 m of boulder rich till and encountered green carbonate footwall rocks at a depth of 30.3 m. Coring in 98-4 penetrated 4.3 m of bedrock.

Figure 43.2 shows a transect of subsurface overburden SP data collected at the Shoot Zone following the drilling program. There are obvious similarities between these data and the theoretical model of Hamilton (1998), shown in Figure 43.1. An electrically reduced column exists in overburden above mineralization that appears to extend to the shallow subsurface. Furthermore, 1997 peat-interstitial groundwater geochemical data from samples collected at approximately 1m depth show anomalies in a number or elements (Ba, Ca, Mg, Mn, Na and Sr) that correlate very well with this reduced column. A number of other elements (H+, Ti and Al – not shown) show a negative anomaly over the reduced column. It should be noted that 98-4, located near but likely north of mineralization shows a similarly reduced zone through what is mostly sand. The contour lines shown in Figure 43.2 are for overburden and ignore data for bedrock. Data for bedrock show a similar trend of increasing values toward mineralization but the much lower resistivity (higher electrical conductivity) of overburden results in a wider area of influence in overburden due to mineralization than occurs in the underlying silicate bedrock.

The presence of an electrochemically reduced column in overburden as is shown in Figure 43.2 must inevitably produce multi-element geochemical anomalies in the shallow subsurface due to changes in speciation of mobile elements (Hamilton, 1998). Elements that are particularly immobile in reduced form may be expected to accumulate in solid media but be depleted in aqueous media over the column. The opposite should be true for elements that are particularly immobile in oxidized form. In addition, the high electrochemical gradient should induce the movement of charged species, also resulting in anomalies (Hamilton, 1998). Positive carrying ions should move inward and downward and negative charge carriers outward and upward. Therefore, it is theoretically possible
Figure 43.3. Surface self potential versus concentration for lines 200E, 315E and 500E, Victoria Creek. Lines 200E and 500E show data for humus whereas line 315E shows data for shallow groundwater.
that some elements could have migrated from bedrock to surface along the high electrochemical gradient in this area but these would have had to be transported as reduced species (e.g. HS⁻, FC⁻²⁻).

Figure 43.3 shows 1998 SP data plotted against selected elements along lines 200E, 315E and 500E at Victoria Creek. Lines 200E and 500E show data from the analysis of humus collected as part of 1995 fieldwork. Line 315E shows peat-interstitial groundwater geochemical data from samples taken at approximately 1 m depth in 1997. The data clearly show a direct correlation with SP for some elements and an inverse correlation for others. Depending on the element, this can result in either apical (single peak) or “rabbit-ear” (double peak) anomalies centred over mineralization. This is consistent with the theory of electrochemically-induced mobility of charged species (Hamilton, 1998) and the idea that negatively and positively charged species will behave in opposite ways.

From Figure 43.3 it is also apparent that, for a given element, the correlation between SP can be direct on one line and inverse on another. This suggests either a change in speciation of elements possibly induced by soil pH differences, or an entirely different Eh regime on the different lines. Further work is clearly required to resolve these issues and soil pH data collected during fieldwork should help. However, the correlation between SP and soil chemistry is so good that we cannot help but conclude that electrochemical phenomena play an important role in the formation of the observed surficial geochemical anomalies.

Although a 3-dimensional SP program has not been carried out at Victoria Creek, preliminary data suggest that a reduced column may also exist at that site. This inference is based on surface SP data and on measurements taken relative to a central base station on line 500E down 4 boreholes cased into bedrock and one monitoring well finished in overburden. The data respectively show a moderately high background SP of -150 mV and -270 mV in shallow bedrock 180 m north and 150 m south of the subcrop of mineralization. A considerably lower SP of -520 mV exists in upper bedrock at a distance of about 90 m north of the subcrop of the north-dipping mineralization. The SP troughs that occur in bedrock and at ground surface near mineralization suggest the presence of a reduced column in that area similar to that documented at the Shoot Zone.

CONCLUSIONS

The above data are consistent with the electrochemical model of Hamilton (1998) and suggest the following:

1. A reduced column exists at the Shoot Zone above mineralization and likely also at Victoria Creek.
2. Geochemical anomalies in shallow surficial materials at both sites are related to self potential phenomena and are, by inference, genetically related to electrochemical processes.

3. Despite the good agreement between data and theory, a number of key issues remain to be resolved:
   - The electrochemical gradients observed at both sites are theoretically sufficient to move many charged species to ground surface in less than 1000 years. This assumes that the overburden behaves as a pure electrolyte and this has not been proved. The presence of solid-phase minerals in overburden, including clay minerals, may either enhance or retard the movement of charge and mass between bedrock and surface. This issue must be further investigated before surface geochemical anomalies in thick overburden can be assumed to be representative of bedrock chemistry.
   - Elements deposited in humus appear to be behaving as charged species but it is not immediately clear which are moving as reduced and which as oxidized species. An investigation into element speciation will help to identify which elements are best used as indicators of bedrock chemistry.
   - The presence of a reduced column will create geochemical anomalies without considering mass transport from mineralization. A method will have to be devised to differentiate this type of anomaly from those due to mass transport from depth and perhaps from mineralization.
   - The cause of the reversal in correlation with SP of some elements between line 200 and 500 at Victoria Creek must be resolved.

REFERENCES


INTRODUCTION

Joints are the dominant structural features in the Paleozoic rocks of south-central Ontario and the best potential indicator of structural history in the area. Ongoing research has established a similarity in orientation between major regional joint sets in the Paleozoic and Precambrian rocks of south-central Ontario (Andjelkovic, Cruden and Armstrong 1996). Trajectories of both regional and local joint sets continue smoothly across the Precambrian–Paleozoic unconformity, suggesting that fractures in the Paleozoic cover are inherited from structures in the Precambrian basement. Aeromagnetic data also reveal a general correlation between the trajectories of the major north-northeast–northeast-trending joint set in Paleozoic rocks and the structural grain of the underlying Precambrian rocks (Andjelkovic, Cruden and Armstrong 1997). Furthermore, lineaments detected on LANDSAT TM and RADAR SAT SAR satellite data can be traced from the Precambrian basement rocks into the sedimentary cover, which also indicates a link between structures in the basement and cover rocks (Andjelkovic and Cruden 1997).

Field observations were collected in the Kingston–Tweed area from selected outcrops in which the Paleozoic–Precambrian unconformity is exposed. The purpose of this work is to better understand the mechanism of basement-cover fracture inheritance.

BASEMENT-COVER RELATIONSHIPS

The trends of magnetic anomalies in southern Ontario are interpreted to reflect the orientation of geological structures in the metamorphosed basement rocks, such as schistosity and foliation, folds, shear zones and faults (Carter, Trevail and Easton 1996). Traces of these geological structures in regional aeromagnetic data continue beneath the Paleozoic cover. This cover is effectively magnetically transparent due to the very low magnetic susceptibility of Paleozoic rocks. Trends between major aeromagnetic anomalies and trajectories of the north-northeast–northeast joint set in Paleozoic rocks are remarkably similar (Figure 44.1).

Two joint trends dominate the study area, north-northeast in the western part and northeast in the eastern part. Both the aeromagnetic data set and joint data exhibit similar behaviour, with a shift from north-northeast directions in the west to northeast directions in the east across an approximately northeast-trending boundary that continues into the Paleozoic cover. The transition between the domains containing these 2 trends can be drawn approximately along the Elzevir–Maxinaw terrane boundary (Andjelkovic, Cruden and Armstrong 1997).

FRACTURE INHERITANCE

The above relationships suggest that formation of fractures in the Paleozoic strata of south-central Ontario was influenced by, and/or inherited from, structures in the underlying Precambrian basement.

Geological structures are considered to be inherited if they owe their character to conditions or events of a former period. However, there is little understanding of the mechanisms involved in the process of inheritance. We propose that mechanisms that explain the inheritance phenomenon fall into 2 types. The first is a “passive” mechanism that involves the formation of pre-, syn-, and postconsolidation joints due to gravitational, water-escape and compactional regimes (Gabrielsen and Arland 1990). Differences in sediment thickness and type in the vicinity of a buried structure, which is expressed as a basement topographic low or high, can result in lateral variations in vertical and horizontal stresses. These stresses may lead to a local change in topographic relief and surface fracturing. This process can be repeated several times as newly deposited sediments de-water and differentially compact over and adjacent to a buried structure (Figure 44.2a, b). The resulting concentration of joints in the rocks overlying the basement high or low may result in enhanced surface erosion and the formation of a linear topographic feature (lineament).

The second mechanism is “active” and involves normal or reverse basement fault reactivation. This implies an active tectonic regime on a regional scale. Jointing in the overlying rocks can occur by reactivation induced by new, renewed or continued movement. New stresses refer to changes in orientation of the principal stresses that generated the buried structures. Renewed stresses refer to an increase in stress magnitude to values that cause permanent deformation. Continued stresses refer to a stress field with similar orientation to that which generated the buried structure. In all 3 cases, jointing in the overlying cover results from strains associated with displacement on the basement fault (Figure 44.2c, d). If displacement of a basement fault is sufficiently high, new faults will also form in the overlying cover (Horsfield...
Figure 44.1. North-northeast-northeast joint trajectories and major aeromagnetic anomalies in south-central Ontario; dashed box is the Kingston area.
Figure 44.2. Mechanisms of inheritance: a) passive inheritance related to basement high; b) passive inheritance related to basement low; c) active inheritance related to normal faulting in the basement; and d) active inheritance related to reverse faulting in the basement. In each case a topographic feature forms at the surface that may be detectable as a lineament on satellite images.
1977). In south-central Ontario, fracturing may have occurred several times between the Cambrian and Devonian periods during the Taconic, Acadian and Alleghanian orogenies (Johnson et al. 1992). During these orogenies, the study area was subjected to repeated episodes of tectonic uplift along the Algonquin and Frontenac arches, curvilinear basement ridges that formed the structural basement for Ordovician sedimentation (Sanford, Thompson and McFall 1985).

In order to differentiate between the various outlined mechanisms, detailed structural mapping at selected outcrops was conducted in 1998. Field work was focussed on the Tweed–Belleville–Kingston area where the sedimentary bedrock cover is thin (less than 100 m) and numerous exposures of the unconformity, both along the erosional edge of the Paleozoic rocks and above several Precambrian highs, allow direct observation of basement-cover relationships. Results are presented from the Barriefield roadcut in Kingston, where a prominent basement ridge is overlain by spectacularly “folded” Paleozoic rocks. The results are considered to be representative in that they illustrate how joint development in Paleozoic cover may be related to differential compaction over and between basement topographic features.

**BARRIEFIELD ROADCUT**

In the Barriefield exposure, basal carbonate units of the Gull River Formation unconformably overlie foliated Grenville quartzites that form a basement high. The Paleozoic rocks consist of fine-crystalline dolostones in the lower part of the sequence, and supratidal laminated lime mudstones in the upper part (Figure 44.3)(Cowan, Narbonne and Smith 1988).

Measurements of bedding planes were taken from stations along the roadcut together with joint-orientation measurements. Joint-orientation data were collected using the traverse mapping technique, which involves measurement of all visible joints along traverses in an outcrop (La Pointe and Hudson 1985). In this case the traverses are defined by the roadcut faces which are oriented 060° (see Figure 44.3). Joint populations are subdivided into a logical grouping of sets that are defined by ranges of frequently occurring orientations. Joint sets were defined by examination of rose and circular diagrams of joint measurements for the entire outcrop.

The beds define an open shallow fold with limbs dipping to the northwest (286°/17°) and southeast (114°/10°) (Figure 44.4). The subhorizontal fold axis and vertical axial plane of the fold trend north-northeast (021°), subparallel to the strike of the basement ridge, which is exposed in the core of the fold. The strike of the basement ridge appears to be controlled by the orientation of a penetrative foliation within the quartzite that is oriented 032°/90°. Rose diagrams and stereonets of jointing within the roadcut display 2 major peaks, north-northeast (013°) and east-southeast (105°). The north-northeast-trending joint set is subparallel to the fold axial plane and represents “bc” or longitudinal joints (Price 1966). The east-southeast-trending set of joints cut the fold at right angles to the fold axis and is defined as “ac” or cross joints. Besides these 2 major sets, there are also oblique joints with varying dips (50 to 80°). These joints form conjugate pairs and exhibit visible shear displacements and dip to the southeast and west-southwest. They are interpreted to be conjugate shear joints formed by layer-normal compaction within the crest and flanks of the basement high.

**DISCUSSION**

The rose diagram of joints in the Barriefield roadcut appears to be different to that of all joints in the Kingston area (Figure 44.4c). However, the north-northeast peak trend of “bc” joints at Barriefield is also present in the

![Figure 44.3. Barriefield roadcut looking to the southwest. Paleozoic strata define an open fold with an exposure of Grenville-age quartzite in the core (dotted) (Cowan, Narbonne and Smith 1988).](image)
Figure 44.4. a) Equal-area, lower-hemisphere stereogram (Gaussian contoured with $K=100$ and extra-fine counting grid) of poles to bedding in the Barriefield roadcut, showing a plot of the axis of the compactional fold (large dot), the strike of the foliation of Precambrian basement (dotted great circle), and the plane of the fold axis plunging to the north-northeast (solid great circle); b) Equal-area, lower-hemisphere stereogram (Gaussian contoured with $K=100$ and extra-fine counting grid) of poles to joints and a rose diagram of joints in the Barriefield roadcut; c) Rose diagrams of joints in the Paleozoic rocks of the Kingston area: rose diagram on left is created from joints in the immediate Kingston area (box in Figure 44.1) and rose diagram on right represents the broader Kingston area (from Figure 28.2, Andjekovic, Cruden and Armstrong 1997).
regional data, but is suppressed due to a predominance of northeast- and east-northeast-trending joints at the regional scale. Likewise, the east-southeast-trending "ac" joints and transverse shear joints can also be identified as minor peaks in the regional data.

We conclude that joints at the Barrie field roadcut formed due to compaction over a north-northeast-trending basement high, which in turn is controlled by the metamorphic fabric of the basement rock. A similar relationship may apply regionally, in which case the predominant basement structure and joint trend is northeast.

Further work, including detailed levelling of stratigraphic markers across lineaments and buried basement structures, will test the hypothesis that jointing is due to differential compaction over linear basement highs rather than by reactivation of basement faults.

REFERENCES


INTRODUCTION

The purpose of this pilot study is to determine whether airborne hyperspectral images can be used to detect biogeochemical indicators over mineralized sites in the boreal forest areas of the Precambrian Shield. Previous work by Singhroy (1988), Singhroy and Kruse (1991) and Springer et al. (1989) have demonstrated that the reflectance spectra of vegetation, as well as their structure and types, are influenced by the geochemistry of the soil and rock.

Recently developed hyperspectral sensors (e.g., casi, Compact Airborne Spectrographic Imager) should be better able to detect the spectral response of geochemically stressed vegetation than imagery previously used.

Image Acquisition

Hyperspectral imagery from the Compact Airborne Spectrographic Imager (casi) in the visible and near infrared was acquired over the test area on August 19, 1998 in 72 contiguous 8.7 nm wide spectral bands covering a wavelength range from 407 nm to 944 nm. The sensor was flown at an altitude of 3000 m above sea level (asl) and produced a spatial resolution of 3.5 m.

SITES

Two test sites were selected along the northeast rim of the Sudbury basin within the casi coverage (Figure 45.1). One site is located over a mineralized zone (Site A) and the other not (Site B).

The first site (Site A) consists of an undulating area of bedrock-dominated terrain (Bajc 1997) underlain by rocks mapped as the 'sublayer' of the Sudbury Igneous Complex (Dressler 1982, 1984). The sublayer is a breccia that has been described as an inclusion-bearing rock unit associated with the copper-nickel ores of the Sudbury Nickel Irruptive Complex (Dressler 1982). Outcrops at the western end of the site, however, expose medium-grained, equigranular norite.

The cover of glacial sediment is thin to non-existent except for a very small area of the test site that is underlain by glaciofluvial outwash deposits (see Figure 45.1).

Archean granitic rocks underlie the second site (Site B; Figure 45.1). The eastern half of the site is a level area underlain by outwash sediments and the western half is an undulating bedrock-controlled area with a thin and discontinuous cover of glacial sediments (Bajc 1997). Outcrops that expose granite are common along the western edge of the site.

Podzolic soils occur at both sites. The A-horizon is well developed with a layer of leaf litter overlying humus that overlies an eluviated zone of mineral soil (Ae-horizon). B-horizons are present at most localities. Samples were collected of the A-horizon (humus) and B-horizon where it was present. Samples were taken along grid lines at intervals of approximately 15 m (see Figure 45.1). Analysis of both the A- and B-horizon samples will be completed to determine chemical composition.

The natural vegetation at both sites is similar. Birch (betula), poplar (populus) and fir (abies) are the common tree types. A total of 81 reflectance spectra of white birch (betula papyrifera) and trembling aspen (Populus tremuloides) were taken from both test sites during airborne data acquisition using a GER3700 spectrometer which operates between 400 and 2500 nm. The leaves were then dried and processed for later chemical analysis.

FUTURE WORK

During the fall and winter (1998–1999) the casi images will be processed, analyzed and correlated with the field spectral data collected, and the soil and leaf chemistry.

REFERENCES


Figure 45.1. Location of the project area, test sites, and soil and rock samples taken.
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# Conversion Factors for Measurements in Ontario

## Geological Survey Publications

### Conversion from SI to Imperial

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<tr>
<th>SI Unit</th>
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<th>Gives</th>
<th>Imperial Unit</th>
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<td>inches</td>
<td>1 inch</td>
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### Area

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<td>1 square foot</td>
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<td>acres</td>
<td>1 acre</td>
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<td>1 m³</td>
<td>1.308 0</td>
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### Capacity

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### Mass

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<td>1 t</td>
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### Concentration

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<th>ounce (troy)/ton (short)</th>
<th>1 ounce (troy)/ton (short)</th>
<th>34.285 714 2</th>
<th>g/t</th>
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<tbody>
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<td>0.583 333 33</td>
<td>pennyweights/ton (short)</td>
<td>1 pennyweight/ton (short)</td>
<td>1.714 285 7</td>
<td>g/t</td>
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## Other Useful Conversion Factors

**Multiplied by**

- 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.*