ISSN 1916-6117 [online]
ISBN 978-1-4435-1575-7 [PDF]

ISSN 0826-9580 [print]
ISBN 978-1-4435-1574-0 [print]

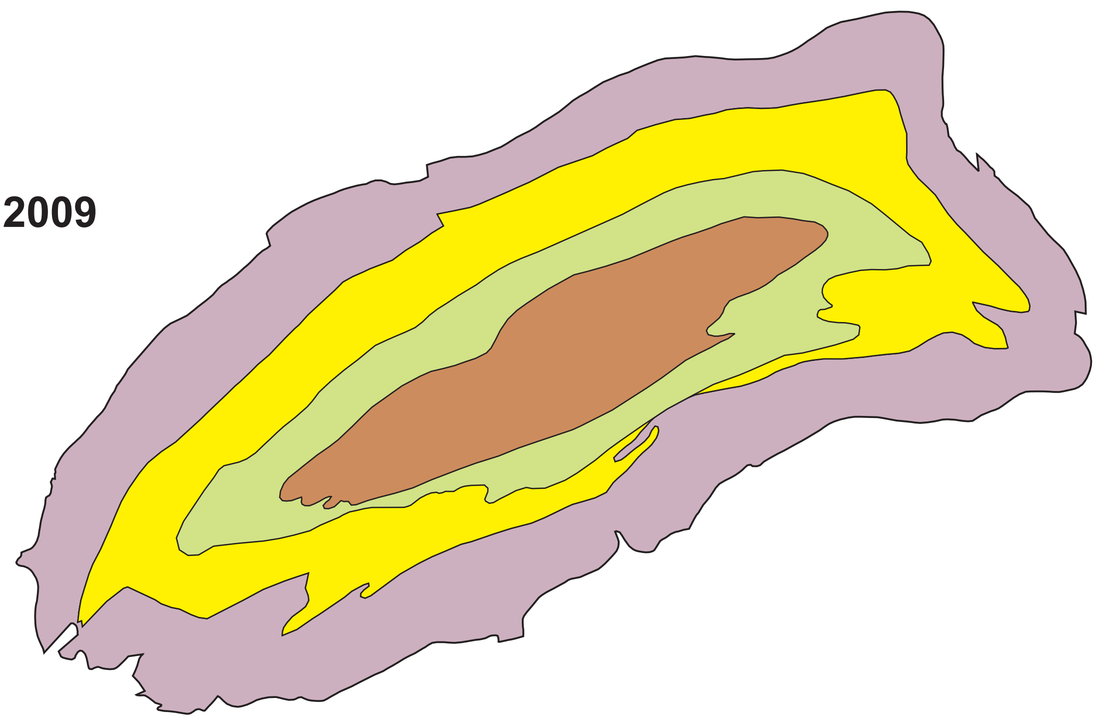
THESE TERMS GOVERN YOUR USE OF THIS DOCUMENT

Ontario

Ontario Geological Survey
Open File Report 6243

A Field Guide to the Geology of Sudbury, Ontario

2009



ONTARIO GEOLOGICAL SURVEY

Open File Report 6243

A Field Guide to the Geology of Sudbury, Ontario

Edited by

Don H. Rousell and G. Heather Brown

2009

Users of OGS products are encouraged to contact those Aboriginal communities whose traditional territories may be located in the mineral exploration area to discuss their project.

Open File Reports of the Ontario Geological Survey are available for viewing at the John B. Gammon Geoscience Library in Sudbury and at the regional Mines and Minerals office whose district includes the area covered by the report (see below).

Copies can be purchased at Publication Sales and the office whose district includes the area covered by the report. Although a particular report may not be in stock at locations other than the Publication Sales office in Sudbury, they can generally be obtained within 3 working days. All telephone, fax, mail and e-mail orders should be directed to the Publication Sales office in Sudbury. Purchases may be made using cash, debit card, VISA, MasterCard, American Express, cheque or money order. Cheques or money orders should be made payable to the *Minister of Finance.*

John B. Gammon Geoscience Library 933 Ramsey Lake Road, Level A3 Sudbury, Ontario P3E 6B5	Tel:	(705) 670-5614
Publication Sales 933 Ramsey Lake Rd., Level A3 Sudbury, Ontario P3E 6B5	Tel: Toll-free: Fax: E-mail:	(705) 670-5691(local) 1-888-415-9845 Ext. 5691 (705) 670-5770 pubsales.ndm@ontario.ca

Regional Mines and Minerals Offices:

Kenora - Suite 104, 810 Robertson St., Kenora P9N 4J2

Kirkland Lake - 10 Government Rd. E., P.O. Box 100, Kirkland Lake P2N 3M6

Red Lake - Box 324, Ontario Government Building, 227 Howey St., Red Lake P0V 2M0

Sault Ste. Marie - Suite 6, 875 Queen Street East, Sault Ste. Marie P6A 2B3

Southern Ontario - P.O. Bag Service 43, 126 Old Troy Rd., Tweed K0K 3J0

Sudbury - Level A3, Willet Green Miller Centre, 933 Ramsey Lake Rd., Sudbury P3E 6B5

Thunder Bay - Suite B002, 435 James St. S., Thunder Bay P7E 6S7

Timmins - P.O. Bag 3060, 1270 Hwy. 101 East, South Porcupine P0N 1H0

This report has not received a technical edit. Discrepancies may occur for which the Ontario Ministry of Northern Development, Mines and Forestry does not assume any liability. Source references are included in the report and users are urged to verify critical information. Recommendations and statements of opinions expressed are those of the author or authors and are not to be construed as statements of government policy.

Cette publication est disponible en anglais seulement.

Parts of this report may be quoted if credit is given. It is recommended that reference be made in the following form:

Long, D.G.F. 2009. The Huronian Supergroup; *in* A Field Guide to the Geology of Sudbury, Ontario; Ontario Geological Survey, Open File Report 6243, p.14-30.

Preface

For over one hundred years, Sudbury has been a destination for practicing geologists, especially those in the mineral industry, as well as students, academics and scientists interested in meteorite impact craters. The Apollo astronauts visited Sudbury in 1971. Numerous field guides have been prepared over the years but they tend to be for specialists and are soon out-of-print. We have prepared a comprehensive field guide, directed toward the general geologist and students, which covers all aspects of the bedrock geology and mineral deposits of the Sudbury area.

Sixteen authors have contributed to the Guide. Most of them have had years of experience in Sudbury geology. However, there are still aspects of Sudbury geology that remain contentious. The Guide is divided into two parts. The first part is an overview of Sudbury geology and consists of thirteen chapters. Some chapters represent a comphrehensive account of a topic, even with new information, while others tend to be summaries. Unless otherwise indicated, all figures and photos are by the authors.

The second part represents a self-guided field trip and comprises fifty stops. Detailed instructions are given to locate each stop including GPS coordinates. Although most stops are roadside rock-cuts, some require walking through the bush.

Those who may wish to pursue topics further are referred to the “References” section, which contains over 380 entries! Even so, the list does not contain every published reference on Sudbury geology. Moreover, new papers and theses appear every year. In 1905 Coleman referred to the literature on Sudbury geology as “very voluminous”. How would he describe it today?

Financial support, from the following sources, is much appreciated.

Centre for Excellence in Mining Innovation, Laurentian University – Dr. Peter Kaiser
Faculty of Science and Engineering, Laurentian University – Dr. Patrice Sawyer
Laurentian University Research Fund for Retired Professors – Dr. Liette Vasseur
Mineral Exploration Research Centre, Department of Earth Sciences, Laurentian University – Dr. Harold Gibson
Vale Inco – Fred M. Stanford
Wallbridge Mining Co. Ltd. – Dr. Bruce Jago
Xstrata Nickel – Mike Romaniuk

The final production of the volume was greatly facilitated by the Ontario Geological Survey. We thank Andy Fyon, Director, for his faith in the project; Thomas Watkins, Team Leader, Publication Services, for support; Alphons Evers for improving the figures and the maps; and Johanne Roux-Guindon, John B. Gammon Geoscience Library, for providing cheerful assistance in locating material.

Lorraine Dupuis of Geo-Cad Services drafted many of the diagrams. Computer problems were expertly solved by Jennie Byron.

Finally, D.H.R. would like to thank his wife Catherine for her tolerance of a “retired” husband who spends much of his time at the office or in the field.

Don H. Rousell and G. Heather Brown, Editors

Authors

1.	Doreen E. Ames	Geological Survey of Canada. dames@nrcan.gc.ca
2.	G. Heather Brown	Ministry of Northern Development, Mines and Forestry. heather.brown@ontario.ca
3.	Ken D. Card	Consulting Geologist (retired, Geological Survey of Canada)
4.	John S. Fedorowich	ITASCA Consulting Canada Inc. jfedorowich@itasca.ca
5.	J. Paul Golightly	Consulting Geologist (retired, Vale Inco). golgeo@sympatico.ca
6.	Michael J. Gray	Pacific International Securities Inc. mgray@pisecurities.com
7.	Richard S. James	Department of Earth Sciences, LU. rjames@laurentian.ca
8.	Scott Jobin-Bevans	TreasuryMetals Inc. scott@treasury metals.com
9.	Darrell G.F. Long	Department of Earth Sciences, LU. dlong@laurentian.ca
10.	Anthony J. Naldrett	Dept. of Geology, University of Toronto. ajn2306@aol.com
11.	Jari J. Paakki	Blackstone Ventures. jari@blv.ca
12.	Edward F. Pattison	Consulting Geologist (retired, Vale Inco). ed.pattison@sympatico.ca
13.	Ulrich Riller	Geography & Earth Sciences, McMaster U. rilleru@mcmaster.ca
14.	Don H. Rousell	Department of Earth Sciences, LU. drousell@laurentian.ca
15.	J.A. Stoness	Consulting Geologist, Vancouver. geojill@canada.com
16.	Brenda Koziol	Science North, Sudbury. koziol@sciencenorth.ca

Contents

PART 1 - SUDBURY AREA GEOLOGY AND MINERAL DEPOSITS

Chapter 1

Geological Setting

D.H. Rousell and K.D. Card

Introduction

Over forty-five years ago Robert Dietz (1962, 1964), prior to even visiting Sudbury, suggested that the Sudbury Basin was formed by meteorite impact and that the Ni-Cu-PGE ores, for which the area is famous, were of cosmic origin. He also correctly predicted that shatter cones, trade marks of astroblemes, would be found in the footwall rocks surrounding the basin. This novel idea sparked new interest in Sudbury geology and expanded and diversified the literature on Sudbury, which, even 100 years ago, Coleman (1905) already considered "very voluminous". Virtually all modern investigators embrace the impact theory (e.g., French 1967; Dence 1972; Guy-Bray 1972; Peredery 1972; Rousell 1981; Morrison 1984; Peredery and Morrison 1984; Lakomy 1990; Grieve et al. 1991; Butler 1994; Grieve 1994; Golightly 1994; Thompson and Spray 1994; Deutsch et al. 1995; Reimold 1995; Ames et al. 1998; Dressler and Reimold 2001; Scott and Benn 2001; Lowman 2002; Therriault et al. 2002; Naldrett 2004; Dence 2004; Mungall et al. 2004; Addison et al. 2005; Ames and Farrow 2007). However, there is a tendency to ignore or force-fit certain aspects of Sudbury geology which are not in exact harmony with the impact model (Muir 1984; Roscoe and Card 1992; Gibbins 1994; Norman 1994; Rousell et al. 1997; Rousell and Long 1998; Cowan et al. 1999; Long 2004). For overviews of Sudbury geology see the following: Bennett et al. 1991; Dressler et al. 1991; Golightly and Rousell 2000; Rousell et al. 2002; Naldrett 2004; Rousell 2006; and Ames and Farrow 2007.

Regional Setting

The Sudbury Basin and Sudbury Igneous Complex (SIC) are part of the Southern Province of the Canadian Shield with the Superior Province to the northwest and the Grenville Province to the southeast (Figure 1.1). The Sudbury area lies within a zone of regional gravity and magnetic anomalies that are possibly attributable to dense, magnetic rocks at depth and to uplifted lower crustal granulites (Card et al. 1984). The area is also cut by several regional fault sets which were active both before and after the Sudbury Impact Event (1850 Ma, Krogh et al. 1984).

Sudbury Structure

The Sudbury Structure (Giblin 1984), formed by the Sudbury Event, consists of three major components: the Sudbury Basin; the SIC which surrounds the basin in the form of an elliptical collar (58 x 28 km) and is geographically divided into North, East and South ranges (Figure 1.1); and an outer zone of locally brecciated footwall rocks, known as Sudbury Breccia, which extends as much as 80 km from the outer margin of the SIC (Simony 1964; Peredery and Morrison 1984).

Rocks of the Whitewater Group, approximately 2900 m thick, occupy the basin and comprise four conformable formations which are, in ascending order, the Onaping, Vermilion, Onwatin and Chelmsford formations (Figures 1.2 and 1.3). The Onaping Formation comprises a basal intrusion overlain by a complex succession of upward-fining breccia units whose origin has been variously ascribed to volcanism (Stevenson 1972), fall-back breccia from meteorite impact (French 1972; Peredery 1972) or to explosive and hydroclastic processes (Gibbins 1994, 1997; Ames et al. 1997; Ames et al. 1998). The Vermilion Formation, too thin to appear on Figure 1.2, consists of carbonate, siltstone and chert breccia (Martin 1957; Stoness 1994). The unit hosts Zn-Pb-Cu mineralization in the former Errington and Vermilion mines (Figure 1.2; Paakki 1992; Gray 1995). The Onwatin Formation comprises carbonaceous mudstone and siltstone, whereas the Chelmsford Formation is dominated by muddy wackes of proximal turbidite origin (Cantin and Walker 1972; Rousell 1972, 1984b).

The SIC (Figure 1.3) consists of four units which are, from outside in and bottom to top: Contact Sublayer, norite, quartz gabbro and granophyre (Naldrett 1984a). The latter three units comprise the so-called Main Mass (Naldrett and Hewins 1984). The Contact Sublayer, a xenolith-bearing norite, is a discontinuous unit which is thickest in depressions in the footwall known as embayments (Souch and Podolsky 1969; Pattison 1979; Naldrett et al. 1984). Concentric and radiating dikes called offsets intrude footwall rocks (Figure 1.2) in zones of Sudbury Breccia or in fractures. The dikes consist mainly of quartz diorite, which is related to the SIC, as well as abundant xenoliths derived from country rocks, basal units of the SIC and hidden mafic bodies (Grant and Bite 1984).

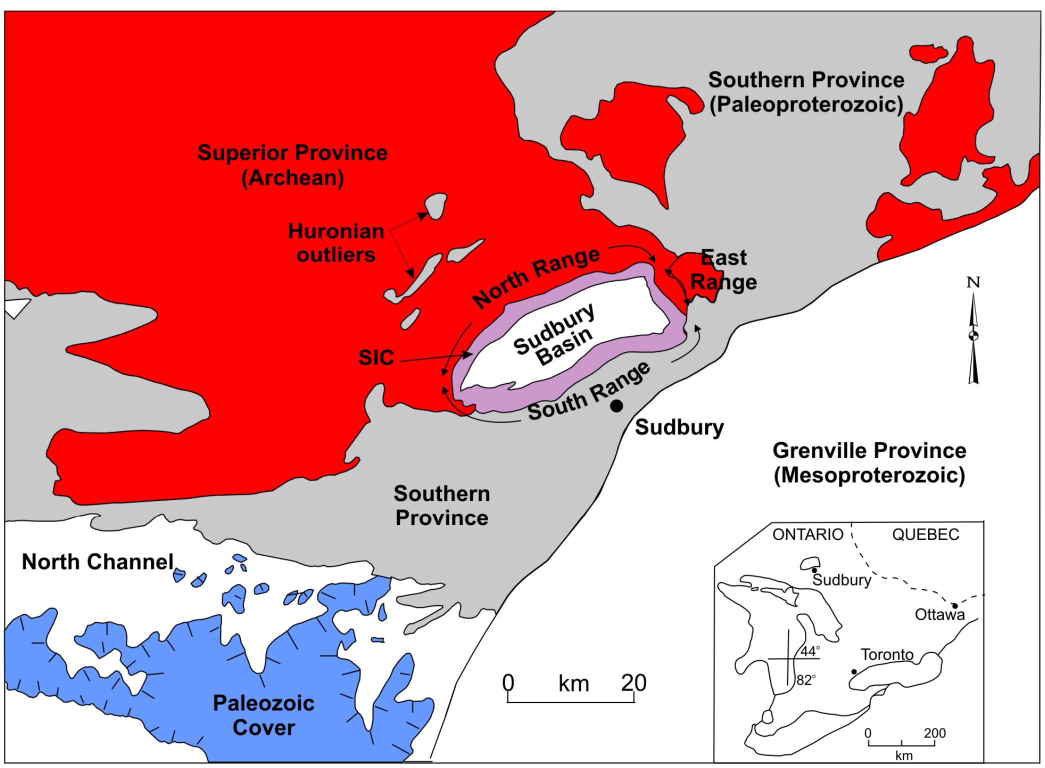


Figure 1.1. Map showing the regional setting of the Sudbury Structure. SIC = Sudbury Igneous Complex.

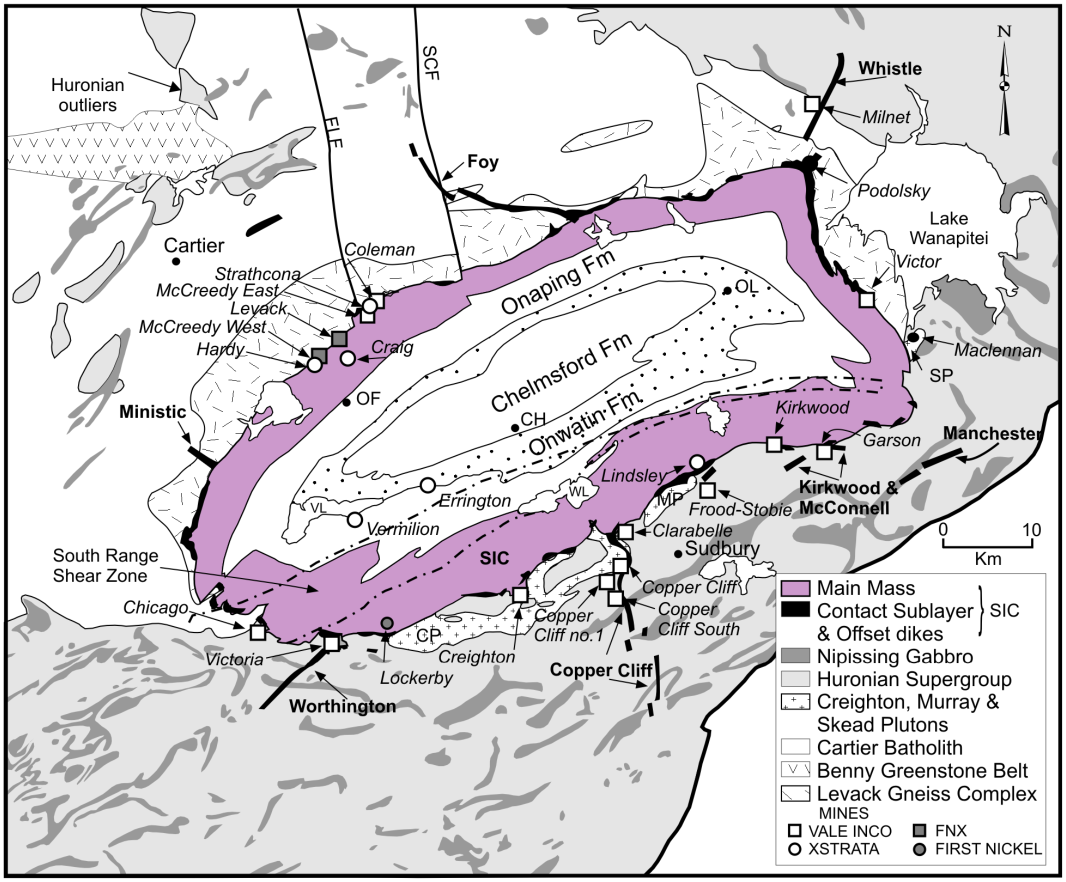


Figure 1.2. Geological map of the Sudbury area with the location of mines and offset dikes. CH = Chelmsford, CP = Creighton Pluton, FLF = Fecunis Lake Fault, MP = Murray Pluton, OF = Onaping Falls, OL = Onwatin Lake, SCF = Sandcherry Creek Fault, SIC = Sudbury Igneous Complex, SP = Skead Pluton, VL = Vermilion Lake, WL= Whitewater Lake.

Sudbury Breccia occurs as irregular bodies or dikes and consists of subrounded clasts, mainly derived from adjacent host rocks, set in a fine-grained to aphanitic matrix (Speers 1957; Dressler 1984a; Müller-Mohr 1992a, 1992b). The breccia seldom comprises more than 5% of the exposed bedrock (Fedorowich et al. 1999). The breccia may be classified into three types based on the nature of the matrix (Rousell et al. 2003). These are: 1) pseudotachylite in the crystalline rocks of the Superior Province; 2) clastic in quartzite of the Southern Province; and 3) microcrystalline in a zone approximately 1.2 km wide in the footwall of the North Range.

The Sudbury Structure hosts some of the world's largest Ni-Cu-PGE magmatic sulphide deposits. The deposits occur at or near the base of the SIC (Figures 1.2 and 1.3). Although the major ore minerals are pyrrhotite, pentlandite and chalcopyrite, there are numerous other minor ore minerals (Hawley 1962; Naldrett 1984c; Dressler et al. 1991). Host rocks for the ores include Contact Sublayer, Footwall Breccia, quartz diorite, Sudbury Breccia and a variety of footwall rocks. Types of deposits, classified in terms of their geologic setting, include: SIC–footwall contact, footwall, offset dike and shear deposits (e.g., Naldrett 1984b; Dressler et al. 1991; Morrison et al. 1994; Golightly and Rousell 2000; Naldrett 2004).

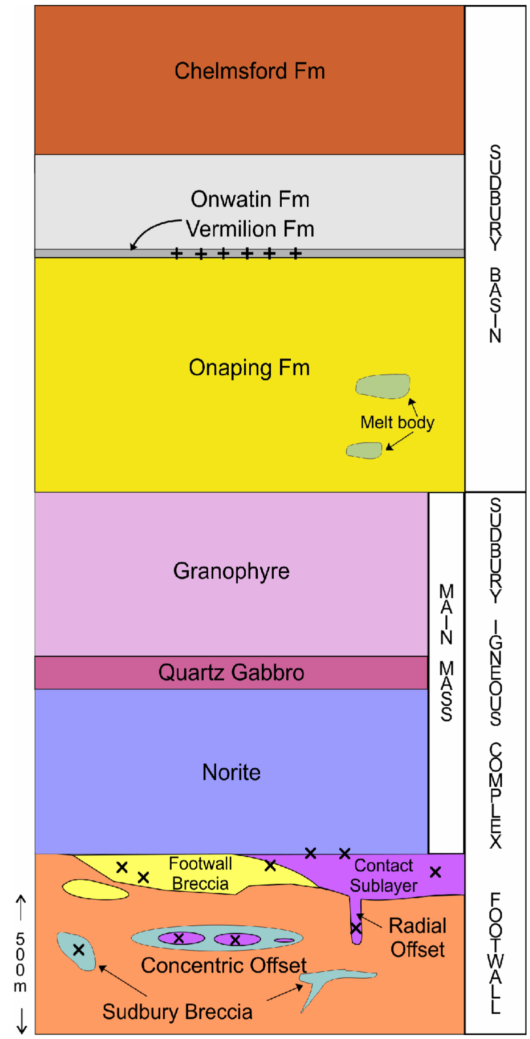


Figure 1.3. Stratigraphic section of the Sudbury Structure. x = Ni-Cu-PGE deposit, + = Zn-Pb-Cu deposit.

Footwall Rocks

Supracrustal rocks of the Huronian Supergroup (2550-2219 Ma, Bennett et al. 1991) are extensively exposed in the Southern Province as well as in a band of outliers in the Superior Province (Figure 1.1). The supergroup is divided into four groups. The lowermost group consists of volcanic and sedimentary rocks, whereas the upper three groups comprise sedimentary rocks only. The Huronian rocks are intruded by dikes and sills of the Nipissing gabbro (2219 Ma, Corfu and Andrews 1986).

Rocks to the northwest and in a narrow belt to the northeast of the basin consist of Archean gneisses, migmatites, granites and volcanic rocks of the Superior Province (>2500 Ma). The North and East ranges are fringed by rocks of the Levack Gneiss complex (Figure 1.2) with a primary isotope age of 2711 Ma (Krogh et al. 1984).

Tectonometamorphism

Prior to the Sudbury Event (1850 Ma), the rocks of the Southern Province and the outliers of the Huronian Supergroup in the Superior Province were folded about east- to northeast-trending axes and locally about north-northeast- to north-northwest-trending axes. The timing of the event is not well constrained as it may have occurred, in part, during or even prior to the emplacement of the Nipissing gabbro at 2219 Ma (Card et al. 1972). This folding may be related to the Blezardian Orogeny (2140 Ma, Stockwell 1982; revised age 2400 to 2200 Ma, Riller and Schwerdtner 1997) or even to an early phase of the Penokean Orogeny (1900-1700 Ma, Bennett et al. 1991).

After the Sudbury Event, the southern and central parts of the basin were folded about northeast-trending axes and more widely spaced northwest-trending axes (Thomson 1957; Zolnai et al. 1984), the South Range was locally displaced by a southeast-dipping zone of reverse shear (Burrows and Rickaby 1930; Shanks and Schwerdtner 1991) and the rocks of the basin were cut by at least five fault groups (Rousell et al. 2002).

The Sudbury area has been affected by several episodes of metamorphism that include contact, regional and shock metamorphism. In the North Range, a zone of contact metamorphism, up to 1.2 km wide, is present in the Archean footwall rocks, with the pyroxene-hornfels facies adjacent to the SIC. The Archean metavolcanic and metasedimentary rocks are metamorphosed to upper greenschist and lower amphibolite grade (Meyn 1970) at circa 2690 Ma (Jackson and Fyon 1991) while those of the Levack Gneiss Complex locally reach the granulite facies (Langford 1960) at 2647 Ma (Krogh et al. 1984). In the Southern Province, regional metamorphism, which ranges from subgreenschist to amphibolite facies (Card 1978a, 1978b), has overprinted planar deformation features in quartz and feldspar (shock metamorphism) and contact metamorphic effects of the SIC.

Regional metasomatism (1700 Ma, Schandl et al. 1994) of Huronian rocks has produced local areas of intense albitization, carbonatization and chloritization, accompanied by the introduction of copper, gold, nickel, cobalt and chromium (Meyer et al. 1986, 1987). A hydrothermal event at 1848 Ma (Ames et al. 1998; Ames, Jonasson et al. 2005) produced semiconformable alteration zones in the Onaping Formation, including silicification, albitization, chloritization, calcitization and feldspathization. The Zn-Cu-Pb mineralization in the Vermilion Formation is interpreted to be a product of this hydrothermal event.

Pre-Sudbury Event Dome

There are several lines of evidence that suggest the Sudbury Basin was the site of a pre-Sudbury Event dome of possible magmatic origin. Cooke (1946, 1948) recognized that the major northeast-trending synclinal axis and the minor northwest-trending synclinal axis in the Paleoproterozoic rocks of the Sudbury Basin overlie a major northeast-trending anticlinal axis and a minor northwest-trending anticlinal axis in the Archean basement. Based on the observation that, in the footwall of the South and East ranges, the oldest rocks are adjacent to the base of the SIC and become progressively younger further away, Speers (1957, p.499) concluded that the basin “had been superimposed on the great dome structure of the underlying rocks”.

Rocks of the Levack Gneiss Complex occur as a fringe around the northwest and northeast margins of the SIC. Metamorphism to the granulite facies at depths between 21 and 28 km (James et al. 1992), followed by uplift to depths between 5 and 11 km during regional amphibolite facies metamorphism, suggests that the complex cored a paleodome (Rousell et al. 1997).

In the footwall of the North and East ranges, diabase dikes range in age from 2600 Ma (Fahrig and West 1986) to 2452 Ma (Heaman 1989). Dikes, located between 10 to 15 km from the outer margin of the SIC, are oriented perpendicular to the adjacent margin of the SIC, which is consistent with dike emplacement during the local magmatic doming of the crust (Rousell et al. 1997).

The voluminous Nipissing gabbro is absent between the northwest edge of the SIC and the Huronian outliers further to the northwest. This suggests little or no deposition of the Huronian sediments, or their complete erosion, implying that the site of the basin was a topographic high during and/or after Huronian sedimentation (Rousell et al. 1997). Three felsic plutons—the Murray (2477 Ma, Krogh et al. 1996), Creighton (2333 Ma, Frarey et al. 1982) and Skead plutons—were intruded into an area of the inferred paleotopographic high and magmatic dome.

Chapter 2

Superior Province

K.D. Card

Introduction

Archean rocks of the Superior Province lie northwest of the Sudbury Basin (Figure 2.1). They are part of the Abitibi Subprovince, a Neoarchean greenstone-granite terrane. Major Archean units in the Sudbury area include supracrustal rocks of the Benny greenstone belt, high-grade rocks of the Levack Gneiss Complex and felsic plutonic rocks of the Cartier batholith. Paleoproterozoic sedimentary sequences of the Huronian Supergroup unconformably overlie Archean rocks and form a number of outliers in the area (Figure 2.1). The Archean rocks are cut by the lower units of the Sudbury Igneous Complex (SIC), intruded by three sets of mafic dikes and were deformed and variably metamorphosed by the Kenoran Orogeny (2500 Ma). These rocks, together with those of the Huronian Supergroup, were also affected by the Blezardian Orogeny (2140 Ma, Stockwell 1982; revised age 2400 to 2200 Ma, Riller and Schwerdtner 1997). Features of shock metamorphism, attributed to the Sudbury Impact Event, include Sudbury Breccia, planar deformation features in quartz and feldspar and kink bands in biotite. Shatter cones are scarce.

Benny Greenstone Belt

The Benny greenstone belt trends east, is approximately 40 x 5 km, and is surrounded and intruded by felsic plutonic rocks (Card and Innes 1981). In general, the belt consists of highly deformed basalt and andesite flows, tuff breccias of tholeiitic and calc-alkaline affinities as well as sulphide-bearing volcanogenic metasediments. The rocks have been regionally metamorphosed under greenschist to lower amphibolite facies conditions.

Levack Gneiss Complex

The Levack Gneiss Complex (LGC) forms a collar, 0.5 to 5 km wide, around the North and East ranges of the SIC. There, it is intruded by the Cartier batholith and SIC (Figure 2.1). Seismic (Milkereit et al. 1992) and gravity (McGrath and Broome 1994) studies suggest that the LGC dips moderately beneath the SIC and steeply beneath the Cartier batholith.

Tonalite-granodiorite orthogneiss forms an almost continuous unit adjacent to the SIC. The unit contains abundant layers, inclusions and boudins of mafic to diorite gneiss, diorite, gabbro, anorthosite, amphibolite, iron formation and pyroxenite. Biotite paragneiss and foliated granodiorite form units to the northwest of the orthogneiss. All three units locally form isolated bodies within the Cartier batholith. There are also several gabbro bodies adjacent to the SIC.

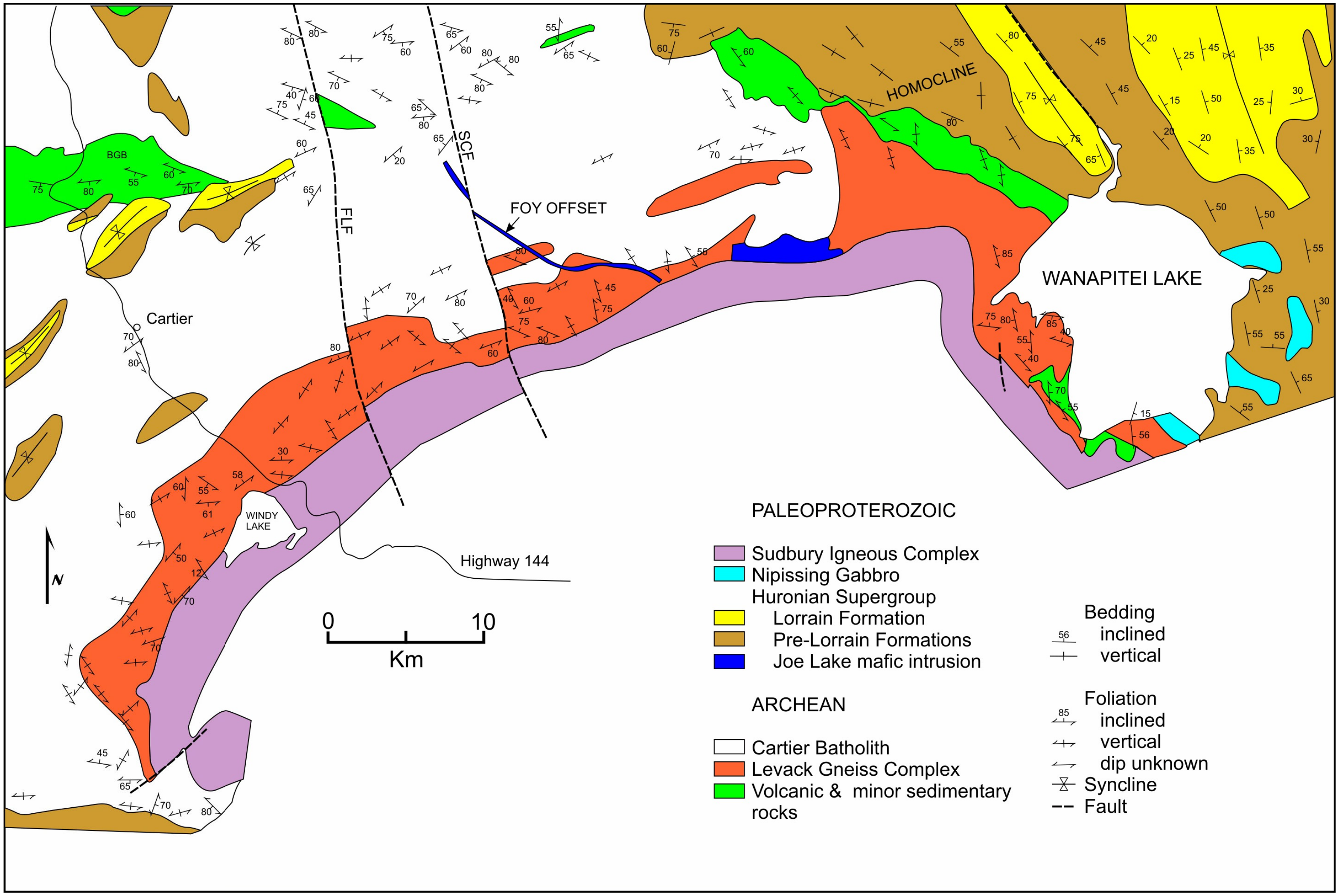


Figure 2.1. Geological map of the footwall of the North and East ranges (*from* Langford 1960; Card 1965, 1994; Card and Meyn 1969; Meyn 1970; Card and Innes 1981; Muir and Taylor 1981; Dressler 1981, 1982, 1984b; Choudhry 1984; Lafleur and Dressler 1985; Legault et al. 2003). Local absence of structural elements is due to a lack of data or the massive nature of the rocks. BGB = Benny greenstone belt, FLF = Fecunis Lake Fault, SCF = Sandcherry Creek Fault.

The rocks of the LGC record a polymetamorphic history (James et al. 1992; James and Dressler 1992; Card and Wodicka, in prep.) including early granulite facies metamorphism, retrograde amphibolite facies metamorphism, a regional greenschist facies overprint and localized shock and contact metamorphism adjacent to the SIC. Rocks of the granulite facies are locally preserved.

Thermobarometric studies (James et al. 1992) suggest that the high-grade metamorphism occurred at pressures of 6.0 to 8.5 kb and temperatures of 750 to 900° C, corresponding to depths of 20 to 30 km. Retrograde amphibolite facies metamorphism, indicated by the breakdown of garnet to plagioclase and biotite, intimates decompression (1.5 to 3 kb) and cooling (500 to 550° C). When and how did the high-grade, lower crustal rocks of the LGC become juxtaposed with the low-grade rocks of the Cartier batholith?

Krogh et al. (1984) obtained U-Pb zircon ages of 2711 Ma for tonalite gneiss and 2647 Ma for mobilizate, which they interpret as a minimum age for the gneiss and the age of high-grade metamorphism, respectively. U-Pb zircon ages obtained by Card and Wodicka (in prep.) indicate three ages of felsic magmatism: 1) early tonalitic at 2711 Ma; 2) granodioritic at 2668 Ma; and 3) monzodioritic at 2657 Ma. According to Card and Wodicka (in prep.) there were at least two periods of high-grade metamorphism at 2661 Ma and 2647 Ma. Attempts to define the post-granulite metamorphic thermal and uplift history of the LGC by means U-Pb and Ar-Ar studies reveal a wide range of values. The geographic distribution of these data is not consistent with models of cooling related to the thermal aureole surrounding the SIC, indicating that the Sudbury Event was not the only, or even the main, cause of uplift of the LGC. The isotopic data, coupled with the evidence for rapid decompression and cooling following high-grade metamorphism, suggests a late Archean period of uplift which broadly coincided with emplacement of the Cartier batholith at 2642 Ma. Uplift of the LGC and emplacement of the Cartier batholith at high crustal levels may have been related to late orogenic extension (Kusky 1993). Following uplift, the LGC probably underwent slow cooling until the early Paleoproterozoic, when the area was intruded by the Matachewan dikes and deformed during the Blezardian Orogeny.

Cartier Batholith

The Cartier batholith is one of a suite of late Archean granitic intrusions, known as Algoma plutons, which form several large batholiths in the southern part of the Superior Province (Card 1979). These intrusions were emplaced as extensive sheets, roughly 5 km thick, at high crustal levels during the late stages of the Kenoran Orogeny (2642 Ma, Meldrum et al. 1997). The Cartier batholith largely consists of coarse-grained, commonly subporphyritic, massive leucogranite. Local porphyritic phases display rapakivi textures. The Algoma intrusions probably formed by the partial melting of tonalitic or granulitic rocks at depth. Their emplacement at high crustal levels apparently signalled the end of major Kenoran magmatism and the beginning of the stabilization of the Superior craton.

Mafic Intrusions

During the early part of the Paleoproterozoic, a suite of gabbro-anorthosite and norite plutons and sills were emplaced in the Sudbury area. Intrusions located west of the Sudbury Basin include the Chicago, Shakespeare–Dunlop (2491 Ma, Krogh et al. 1984) and East Bull Lake intrusions (Peck et al. 1993; 2480 Ma, Krogh et al. 1984). The latter two bodies are surrounded by Archean rocks. The Chicago (at the junction of the North and South ranges), Norduna (at the junction of South and East ranges; ca. 2450 Ma, Prevec 1993) and Joe Lake intrusions (Figure 2.1) are small bodies at the margin of the SIC.

The Archean rocks are cut by mafic dikes belonging to the Matachewan (2452 Ma, Heaman 1989), Nipissing (2220 Ma, Corfu and Andrews 1986) and Sudbury (1238 Ma, Krogh et al. 1987) swarms. Matachewan dikes have two trends: northwest and northeast. Both trends transect the gneissic foliation in the Archean rocks and display chilled contacts. Some dikes display coarse plagioclase phenocrysts. Dikes of the Sudbury swarm consist of olivine diabase. See Chapter 5 of this volume for data on the Nipissing gabbro.

Structural Geology

Lithologic units of the LGC are discontinuous. The trends of gneissic layering are highly variable and are, in general, discordant to the contacts with the SIC and the Cartier batholith (Figure 2.1). Mineral lineations plunge at moderate angles to the northeast. Minor folds with sheared limbs are considered the product of inhomogeneous ductile deformation. Boudins and inclusions with the external foliation deflected around them, and a rotated internal fabric, are common. The LGC apparently underwent high-grade tectonometamorphism at depth during the Neoarchean. The gneissosity is locally overprinted by a northeast-trending cleavage or micaceous foliation which developed under low-grade conditions during the early Paleoproterozoic.

The north-northwest-striking Fecunis Lake Fault and the Sandcherry Creek Fault offset the top and base of the LGC in a sinistral sense (Figure 2.1).

Chapter 3

Felsic Plutons

U. Riller

Introduction

The footwall rocks of the South and East ranges are intruded by three felsic plutons. The Creighton (2333 Ma, Frarey et al. 1982; > 2420 Ma, Smith et al. 1999) and Murray (2477 Ma, Krogh et al. 1996) plutons occur in the footwall rocks of the South Range whereas the small Skead pluton is located near the south end of the East Range (see Figure 1.2 for locations). The plutons intrude volcanic and sedimentary rocks of the Elsie Mountain and Stobie formations of the Elliot Lake Group (lowermost group of the Huronian Supergroup), enclose roof pendants of both formations, and are themselves thermally overprinted by the Sudbury Igneous Complex (SIC). The plutons were emplaced during the Blezardian Orogeny (2140 Ma, Stockwell 1982; revised age 2400 to 2200 Ma, Riller and Schwerdtner 1997) during which time the rocks of the Huronian Supergroup were deformed under amphibolite-facies metamorphic conditions (Card et al. 1972; Card 1978a). Chemical similarities between the Creighton and Murray plutons and the Copper Cliff Formation (Elliot Lake Group), together with similar isotope ages, intimates a relationship between the plutons and the formation (Card 1978b). The Creighton, Murray and Skead plutons differ significantly from each other in terms of size, composition and structure.

Lithology

The Creighton pluton, largest of the three plutons, encloses two prominent greenstone enclaves which lithologically resemble the Huronian Supergroup host rocks (Figure 3.1). The pluton is composed of three granitoid phases: 1) melanocratic quartz monzonite with as much as 20% mafic minerals; 2) leucocratic granite consisting of equigranular quartz (approximately 0.2 mm in diameter), recrystallized plagioclase, perthite and minor biotite; and 3) porphyritic quartz monzonite comprising orthoclase phenocrysts, up to 5 cm long, in a coarse-grained matrix of quartz, plagioclase, perthite and biotite. Phase three is dominant and varies in appearance from magmatic to weakly foliated to augen gneiss, depending on the magnitude of the imposed deformation. Crosscutting relationships suggest that phase one was emplaced first, followed by phase two, then phase three (Riller and Schwerdtner 1997).

The Murray pluton consists largely of equigranular granitic rocks. Based on the orientation of planar mineral fabrics, the pluton can be divided into a southeast and a northwest structural domain (Riller et al. 1996). The southeast domain is characterized by a weakly developed preferred orientation of elongate mafic clots, biotite and euhedral feldspar, which suggest a magmatic origin of the fabric. In the northwest domain, adjacent to the SIC, biotite and amphibole are interlayered with felsic ribbons which consist of recrystallized K-feldspar, plagioclase and elongate quartz. The fabric is interpreted to have formed as a result of an overprint of regional metamorphism. However, the fabric in this domain is also characterized by interstitial K-feldspar and plagioclase associated with strain-free, isometric quartz grains with triple junctions. This attests to a high-temperature thermal overprint, attributed to the SIC, involving the partial melting of granitoid rocks (Riller et al. 1996; Rosenberg and Riller 2000).

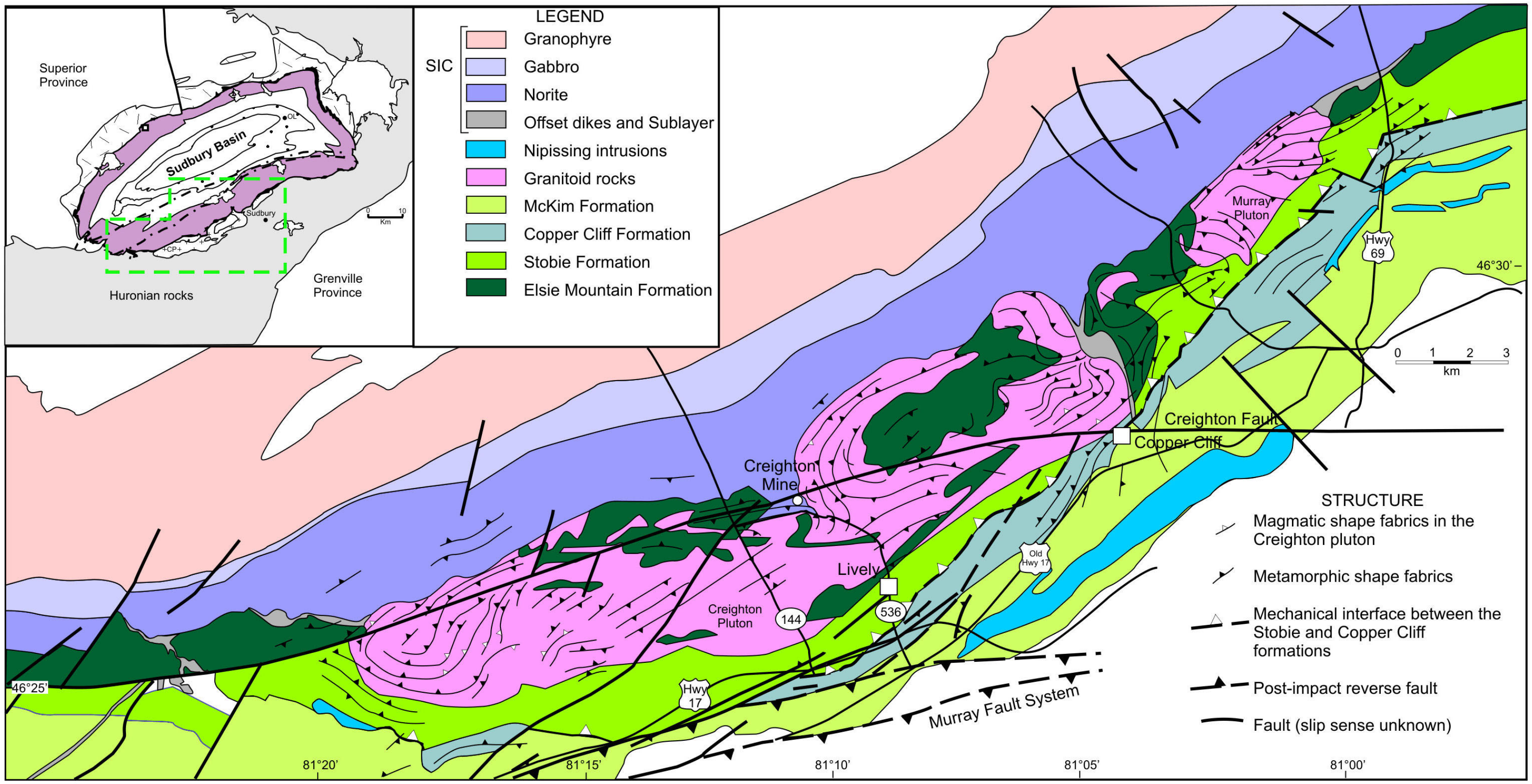


Figure 3.1. Structural map of the Creighton and Murray plutons with their Huronian Supergroup host rocks (*after* Riller and Schwerdtner 1997).

The small Skead pluton is located at the south end of Lake Wanapitei and adjacent to the SIC (Ames, Davidson et al. 2005). Dressler (1982) interpreted the rocks of the Skead pluton as pre-Huronian Supergroup in age, mapped them as granodiorite, diorite and migmatite and equated them with the footwall rocks of the East Range between the SIC and Lake Wanapitei. The latter rocks are currently mapped as part of the Levack Gneiss Complex (Ames, Davidson et al. 2005).

Emplacement of the Plutons

The Huronian host rocks of the Creighton and Murray plutons dip steeply northwest and face southeast. Various emplacement models have been proposed with this structural characteristic in mind. Dutch (1979) concluded that the Creighton pluton was emplaced as a diapir with the present level of exposure near the root zone. Card (1979) argued for magma emplacement during regional metamorphism. The large greenstone enclaves are cited as evidence for extensive block stoping of the roof zone. Syntectonic emplacement of granitoid magma is further supported by the following: 1) concordance of magmatic and metamorphic foliation within the pluton; 2) continuity of foliations across intrusive contacts; and 3) general obliquity between foliation surfaces and pluton contacts (Figure 3.1). However, it is uncertain as to whether pluton emplacement occurred as a consequence of crustal extension during Paleoproterozoic continental rifting (Rousell et al. 1997; Rousell and Long 1998) or during subsequent transpressive deformation with emplacement into the hinge zone of a major fold (Riller and Schwerdtner 1997; Riller et al. 1999). Regardless of the regional deformation regime active during pluton emplacement, the Creighton and Murray plutons have been tilted as a consequence of impact-induced deformation during the crater-modification stage (Riller 2005).

Chapter 4

The Huronian Supergroup

D.G.F. Long

Introduction

The Huronian Supergroup in the vicinity of Sudbury is one of the best-exposed Paleoproterozoic successions in the world (Figure 4.1). It includes a considerable thickness of sandstones, mudstones, carbonates and conglomerates (with minor volcanics) that have been metamorphosed regionally to greenschist facies, and locally to amphibolite facies. The sequence was deposited between 2.45 and 2.22 Ga, and reflects the initiation and development of a continental margin from an early transform margin (marked by left-lateral strike-slip activity), to a passive margin, facing a newly formed Paleoproterozoic ocean (Figure 4.2).

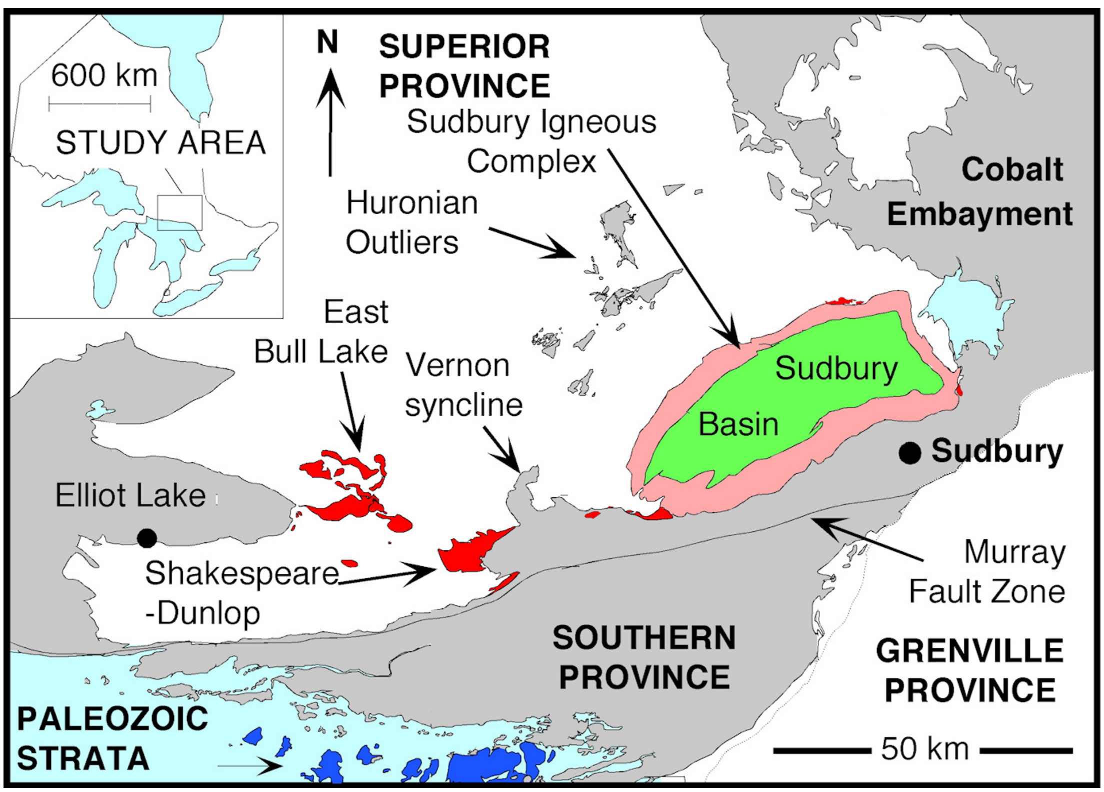


Figure 4.1. Distribution of Huronian strata (grey) in the vicinity of Sudbury. Early Paleoproterozoic intrusions are red, the Sudbury Igneous Complex is pink, Whitewater Group is green, and Paleozoic strata are dark blue.

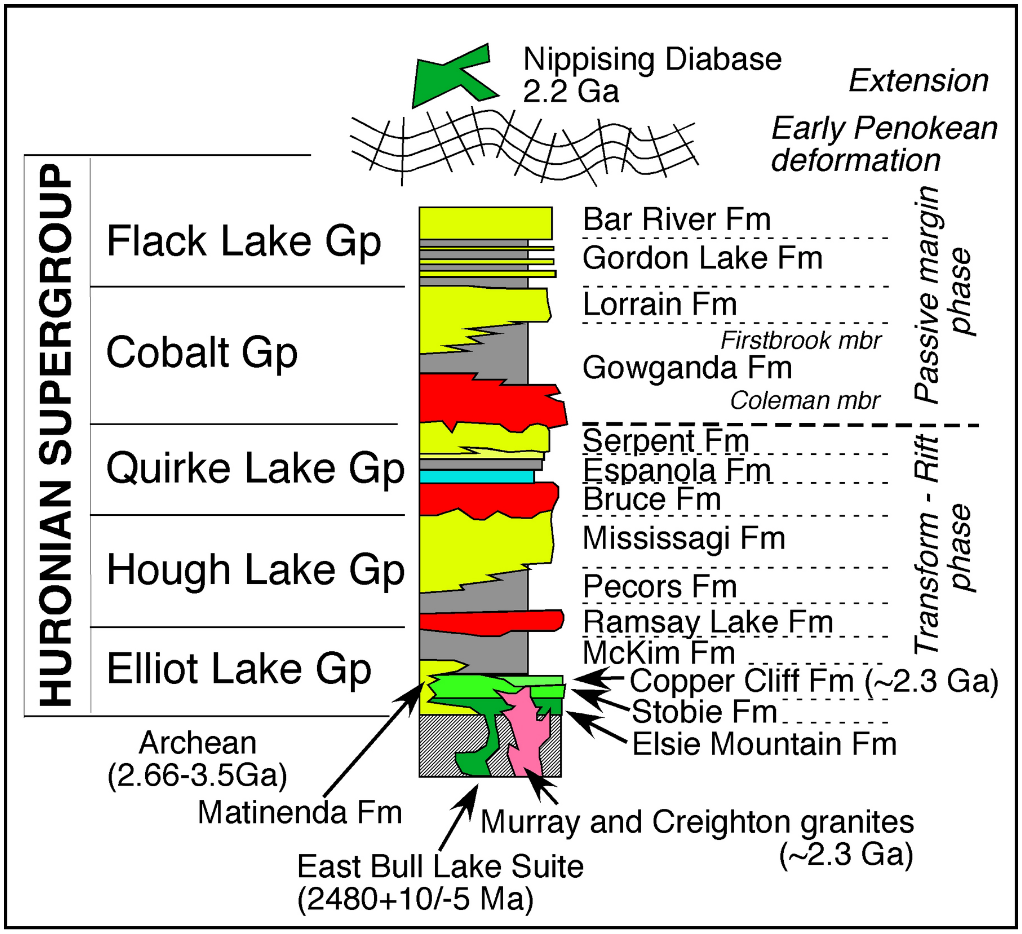


Figure 4.2. Lithostratigraphy and tectonic history of the Huronian Supergroup (*based on* Long 2004). Mudstone is shown in grey, limestone in blue, sandstone in greenish-yellow, and conglomerate in red. Other colours indicate intrusive and extrusive rocks.

The Huronian Supergroup contains evidence of three major glacial events, as well as a record of the transition from oxygen-deficient to oxygen-enriched atmospheres (Roscoe 1969; Young et al. 2001; Long et al. 1999; Long 2004). In the Sudbury area, strata of the Huronian Supergroup occur extensively in the Southern Province, to the south of the Sudbury Igneous Complex, where it is over 12 km thick. Strata are also present to the west (Elliot Lake – Sault Ste. Marie), to the northeast (Cobalt Embayment), and as outliers on the Superior Province, immediately north of the Sudbury Structure (Figure 4.1). Huronian rocks within the Southern Province are considered to be the eastern extension of the Penokean Fold Belt of Michigan, Wisconsin and Minnesota (Young 1983, 1995; Sims et al. 1993). Huronian rocks have been correlated with other Paleoproterozoic successions in places as widely separated as Chibougamau in northern Quebec (Long 1974), Michigan, Wisconsin and Minnesota on the south shore of Lake Superior (Young 1970, 1983), southeast Wyoming (Roscoe and Card 1993) and the Hurwitz Group on the west side of Hudson Bay (Young 1973a; Aspler and Chiarenzelli 1998).

The lower part of the Huronian Supergroup was deposited in an extensional basin, developed over a transform segment of a Paleoproterozoic continental margin, characterized by left-lateral strike slip (Long 2004). Development of this pull-apart basin began around 2.45 Ga, due to the break-up of the Archean supercontinent of Kenorland (Aspler and Chiarenzelli 1998). Transition from a rifted strike-slip margin to a passive margin occurred during deposition of the Gowganda Formation, Cobalt Group (Figure 4.2), and was accompanied by major slope failure and soft sediment deformation south of the Murray Fault (Young and Nesbitt 1985). Within the Southern Province, diabase dikes (Nipissing) were intruded into unconsolidated sediments at considerable depth, producing round-stone breccias, which are texturally similar to Sudbury Breccia (Shaw et al. 1999). Minor compressional folding occurred prior to emplacement of the Nipissing Diabase Suite at around 2.2 Ga (Figure 4.2). The abundance of diabase intrusions near the Sudbury Structure suggests that the area may have been a focal point for magma emplacement prior to emplacement of the Sudbury Structure at 1.85 Ga.

The economic potential of strata within the Huronian Supergroup is limited. Basal strata of the Elliot Lake Group in Hyman Township (west of Sudbury District) were exploited for uranium in the 1960s (Card 1965, 1968), and may contain further deposits at depth (Barnes and Lalonde 1973; Long and Lloyd 1983). The Nipissing diabase has a minor potential for fracture-fill gold deposits (Card and Poulsen 1998). Copper has been found in these intrusions west of Sudbury near Bruce Mines, and silver in veins cutting sills in the Cobalt area, but neither commodity has been found in these rocks in the Greater Sudbury District.

Stratigraphy

Strata of the Huronian Supergroup are discussed here in stratigraphic order (Figure 4.2). More comprehensive reviews of the geology and geotectonic setting of the area can be found in papers by Rousell et al. (1997), Long et al. (1999), Young et al. (2001), and Long (2004), and references cited therein. The oldest preserved strata within the Huronian Supergroup are arkosic and subarkosic sandstones of the Livingstone Creek Formation, which are preserved locally between Thessalon and the area north of Sault Ste. Marie (Bennett et al. 1991; Bennett and Meyer 1995). These may have been restricted to pre-rift basins that occurred prior to emplacement of PGE-bearing mafic intrusions west of Agnew Lake (James, Easton et al. 2002), low-Ti tholeiitic lavas of the Thessalon Formation (and equivalents) west of Elliot Lake (Bennett et al. 1991; Jolly et al. 1992), and mafic to acidic rift-related extrusions in the Sudbury area. Strata in the American part of the Penokean Fold Belt are discussed in Sims et al. (1993).

ELLIOT LAKE GROUP

The Elliot Lake Group in the Sudbury District begins with a thick package of volcanic rocks and deep-water sediments that accumulated within a transtensional (pull-apart) basin caused by left-lateral movement along the precursor of the Murray Fault system (Long and Lloyd 1983; Long 2004). The Elliot Lake Group in this area has been divided into the Elsie Mountain, Stobie and Copper Cliff formations (Figure 4.3). Each of these units contains minor intervals of stratified metasediment (predominantly laminated mudstones and siltstones).

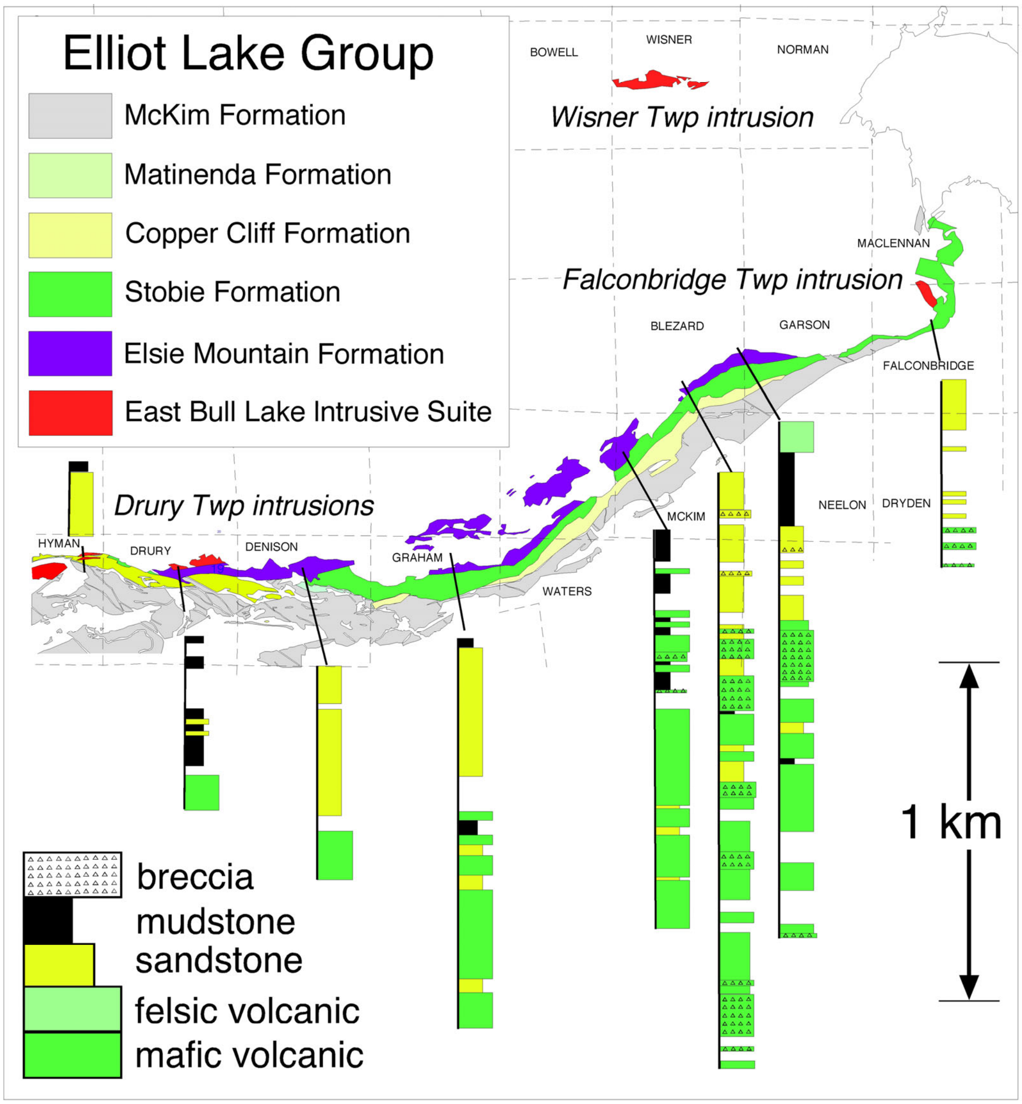


Figure 4.3. Distribution of the Elliot Lake Group in the Sudbury District (*after* Ames, Davidson et al. 2005). Stratigraphy of the lower part of the Elliot Lake Group from Long and Lloyd (1983), and Lloyd (1984). Top of Elsie Mountain Formation is placed at the top of the last mafic flow over 20 m thick.

Elsie Mountain Formation

Iron-enriched tholeiitic basalts of the Elsie Mountain Formation locally exceed 1 km in thickness (Figure 4.3). About 90% of the formation consists of massive and pillowed flows (20 to 90 m thick: Innes 1978a) that were emplaced in less than 1.5 km of water. The thickness of individual flow units thins up-section and to the west (Card et al. 1977). Most flows appear massive or have structure obscured by foliation or development of metamorphic minerals. Pillowed flows (Photo 4.1) occur throughout the formation, but are more common towards the top (Innes 1978a). The Elsie Mountain Formation basalts have been genetically linked to mafic intrusives of the East Bull Lake igneous suite (Vogel, James et al. 1998; Vogel, Vuollo et al. 1998) and appear to represent a plume-derived magma, similar to E-MORB from the Ontong–Java Plateau, with about 20% crustal contamination (Hocking 2003).

Photo 4.1. Well-developed pillow in Elsie Mountain Formation, McKim Township. Hammer is 74 cm long.

Interflow units within the Elsie Mountain Formation include flow-top breccias and minor volcaniclastic units. Mudstone intervals are typically 2 to 3 m thick, although units up to 30 m thick are present in Waters Township (Lloyd 1984). Most of the mudrocks are thinly to thickly laminated. The presence of graded laminated sediments indicates settling in a deep-water prodeltaic setting. The contact of the Elsie Mountain Formation with overlying volcanic and volcaniclastic rocks in the Stobie Formation is transitional, and is placed by Innes (1978a) where interflow sediments make up more than 15% of the section, or where individual flows are less than 20 m thick.

Stobie Formation

The Stobie Formation consists predominantly of felsic to mafic metavolcanic rocks, pyroclastic breccias and pelitic metasedimentary rocks (Debicki 1990). It thickens from about 840 m in the west (Innes 1978a) to at least 1240 m in the east (Debicki 1990). Individual flows can appear massive, with flow-top breccias, or pillowed, with minor interstitial hyaloclastite. Rare beds of mafic pyroclastic material (tuff-breccia and lapilli tuff) are present near the upper part of the formation in units up to 1 m thick.

Siltstone and muddy-sandstone (arkosic wacke) occur as interflow sediments in the middle and upper parts of the formation. Most are thinly to thickly laminated prodeltaic units, and may show signs of syndepositional slumping. Thin sandstone beds may contain ripple lamination. Debicki (1990) records rare 0.5-3.5 m thick units of intraformational breccia (sed-volcani-lithrudite) in the middle and upper parts of the formation in Falconbridge Township, which may be of subaqueous debris flow origin. In the same area she recorded the presence of sandy, matrix supported, slightly bouldery, very-large pebble conglomerate in units 3 to 10 m thick. These sandy conglomerates contain subangular to rounded clasts of granite, quartz, mafic volcanics, arkosic sandstone and mudstone, that fine upwards locally into medium-grained sandstones with rare clasts. Given the deep-water aspect of the laminated mudrocks, these would appear to be sediment gravity flow deposits (possibly grainflows) derived from the northeast.

Copper Cliff Formation

Strata in the Copper Cliff Formation are dominated by quartz-feldspar crystal tuff and felsic lithic tuff and tuff breccia with minor flow-banded rhyolites (Card et al. 1977; Debicki 1990). The Elsie Mountain and Stobie formations are both intruded by granites of the Murray and Creighton plutons. As these do not cut the overlying Copper Cliff Formation, they may represent the magmatic feeder systems for the rhyolites.

The Copper Cliff Formation is between 730 and 300 m thick in McKim and Waters townships, and thins to a zero edge in Garson Township (Figure 4.3; Card et al. 1977; Debicki 1990). Debicki (1990) notes the presence of thin, discontinuous lenses of massive and laminated muddy sandstone, and minor large pebble conglomerates (rhyolite-lith-rudites) in the upper part of the formation, which she considers to have been derived from the contemporaneous volcanics. The only extensive development of mudstone units within the Copper Cliff Formation is in Waters Township, where Lloyd (1984) recorded a sequence about 10 m thick.

Interflow sediments within the basal Huronian volcanic sequence are predominantly of deep-water (sub-wave-base) origin, consistent with deposition in a prodeltaic setting with a rapidly subsiding pull-apart basin (Long and Lloyd 1983; Long 2004). Two prominent planar-laminated sandstone units in Drury Township (Highway quartzites of Barnes and Lalonde 1973) appear to represent subaqueous grainflow deposits that accumulated in a prodeltaic environment.

Matinenda Formation

North, east and west of the Sudbury Structure, on the southern fringe of the Superior Province, the basal strata of the Elliot Lake Group includes planar and trough cross-stratified medium- and coarse-grained sandstones (arkose and subarkose), and minor plane bedded and massive medium- and large-pebble conglomerate (uraniferous, pyritic, quartz-rudite), that accumulated on alluvial fans and in associated shallow ephemeral gravel-bed rivers and mixed gravelly sand-bed braided rivers. The basal conglomerates and underlying soils are poorly exposed in the Sudbury District, but are exposed locally in Denison Township. These fluvial sequences are stratigraphic equivalents of the Matinenda Formation in the Elliot Lake area, which was a major source of uranium ores in the 1960s (Roscoe 1969, 1996). Paleoflow observations in the Matinenda Formation and associated units are complex (Figure 4.5). Major sources for detritus appear to be to the north of the Elliot Lake area and the Cobalt Embayment. The pattern south of the Sudbury structure reflects mixing of these two sources, and may indicate that both the area now underlain by the Sudbury Structure and the Grenville Front were positive areas.

McKim Formation

Above the basal sandstones of the Matinenda Formation and the extrusive rocks of the Copper Cliff Formation is a thick sequence of mudrocks and minor siltstones of deep-water aspect that have been included in the McKim Formation. These are exposed extensively within the city of Greater Sudbury (McKim Township) where the unit is 1500 to 1800 m thick. Card et al. (1977) recognized three main lithotypes within the formation: muddy-sandstones (arkosic and subarkosic wackes); laminated mudstones (argillites); and thin-bedded sandstones (subarkosic arenites). The muddy-sandstone facies forms up to 50% of the formation in Waters and Graham townships, decreasing to 35% in Denison Township. It is typically poorly sorted, fine- to coarse-grained sandstone, with 45-70% mud. Sedimentary structures include ripple marks, ripple drift cross-lamination, parallel lamination and graded bedding. Card et al. (1977) indicate that many of these units have features consistent with deposition as turbidites, and include complete and incomplete Bouma cycles. They indicate a decrease in average set thickness from about 0.6 m in the east, to 0.05-0.1 m in the west, and note that massive graded beds at the base of sets become less abundant. This is consistent with deposition from inertia-dominated homopycnal (buoyant) flows or gravity driven hyperpycnal (gravity) flows in a prodeltaic slope setting, fed from a deltaic source south of Lake Wanapitei (Long and Lloyd 1983). Laminated mudstones make up much of the McKim Formation in the Sudbury District (Photo 4.2). It occurs in thin, composite graded-laminated sets that are typically only a few centimetres thick. Their appearance suggests deposition from buoyant sediment-rich plumes as prodeltaic suspension deposits. Composite sets could represent annual accumulation (varves) or simple flood events of much shorter duration. Sandstone units are dominated by pink to grey, massive and ripple cross-laminated, moderately well to poorly sorted, fine- to medium-grained sandstones (subarkosic arenite and wackes). Collectively they form about 10% of the formation, with the greatest abundance southeast of Sudbury (Card et al. 1977).

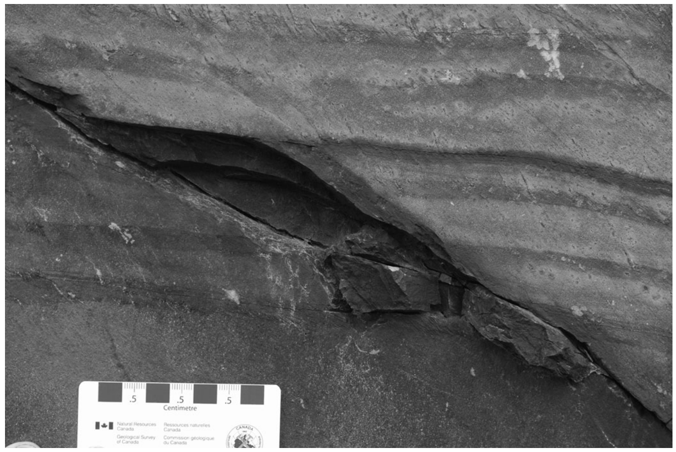


Photo 4.2. Typical exposure of the McKim Formation, north side of Lorne Street, Sudbury (41°28.825'N, 81°00.697'W). Note graded sandstone units (light) are interbedded with thinly laminated mudstones (black).

HOUGH LAKE GROUP

The Hough Lake Group is confined to the Southern Province, and the southern part of the Cobalt Embayment (Figure 4.4). It consists of three main units: the Ramsay Lake Formation; Pecors Formation; and the Mississagi Formation. The basal contact is an erosional discontinuity that can be traced throughout the region and may represent a major sequence boundary.

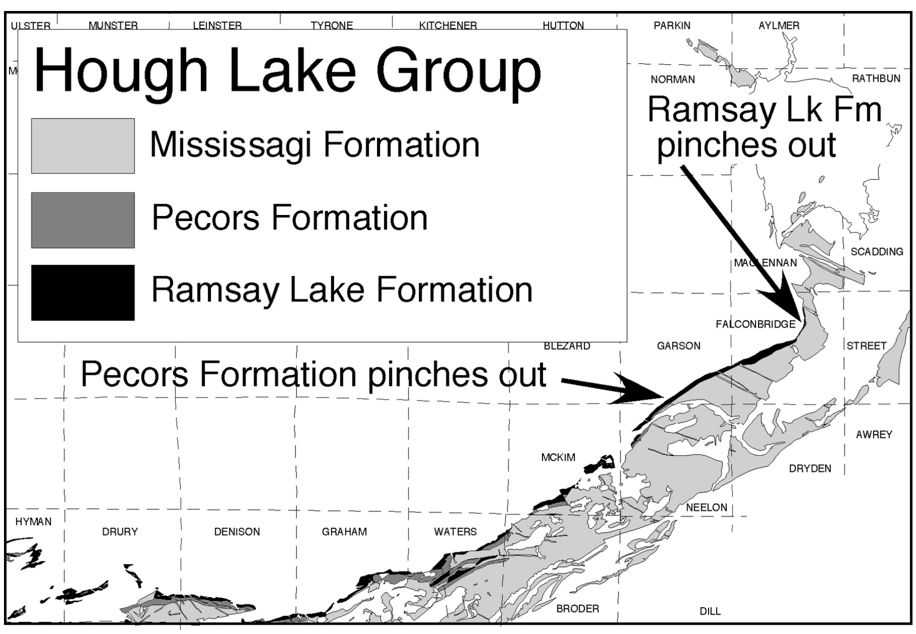


Figure 4.4. Distribution of strata of the Hough Lake Group in the Sudbury area (*after* Ames, Davidson et al. 2005).

Ramsay Lake Formation

The Ramsay Lake Formation is characterized by thick to very thick beds of massive clast-rich sandy diamictite (terminology of Moncrieff 1989, as modified by Hambrey 1994), deposited as an extensive sheet of sub-glacial melt-out till. It occurs extensively along the south side of Ramsey Lake, after which it was named (the difference in spelling reflects a change made after the unit was first defined). The basal contact can be sharp, but may appear transitional due to incorporation of unconsolidated material into the base of the sequence during glacial advance. Clasts within the diamictite are typically rounded to subrounded (Photo 4.3). They include dispersed boulders, cobbles and pebbles, set in a matrix of muddy, medium- to coarse-grained sandstone. Sandstone units are present locally near the base of the formation, and form the dominant lithology in the uppermost third of the formation (Card et al. 1977). Most of the sandstone units appear massive, although ripple cross-lamination is present locally. The formation thickens from 40 m in the north, to about 180 m south of Kelly Lake. It appears to pinch out to the west, in the northern part of Falconbridge Township, due to erosion at the base of the Mississagi Formation (Figure 4.4).

Photo 4.3. Stratified diamictite of sub-glacial melt-out origin, north side of highway 17 near Kelly Lake (46°26.010'N, 81°04.658'W).

Pecors Formation

The Pecors Formation conformably overlies strata of the Ramsay Lake Formation. It is characterized by thinly laminated mudstones (argillites) and muddy-sandstones (wackes) of prodeltaic origin that closely resemble strata in the McKim Formation, from which it can only be distinguished based on location above the Ramsay Lake Formation. The formation is as much as 430-600 m thick in the western part of the study area (Card et al. 1977), and pinches out at the southern boundary of Garson Township, due to erosion beneath the Mississagi Formation (Figure 4.4). Locally the contact between the formations appears sharp and conformable (Photo 4.4).

Photo 4.4. Sharp contact between mudstones of the Pecors Formation (light shade) and planar cross-stratified, medium-grained sandstones of the Mississagi Formation (dark shade), Gibson Road, Waters Township (south of Kelly Lake). Hammer is 74 cm long.

Mississagi Formation

South of the Sudbury Igneous Complex the Mississagi Formation is between 1600 and 3400 m thick. It thins to about 700 m in the east, immediately south of Lake Wanapitei, and has a patchy distribution in the southern Cobalt Embayment due to erosion beneath younger formations, and significant relief at the base of the formation (Long 1987). Regionally it is characterized by medium- to coarse-grained sandstone of arkosic to subarkosic composition, with abundant planar and trough cross-stratification (Photo 4.5). Conglomeratic and argillitic strata are present locally near the base of the formation north of Lake Wanapitei, where the formation can be divided into an upper (sandstone dominated) and lower (conglomerate bearing) member (Long 1976, 1978, 1987). The formation is predominantly fluvial in origin, and was deposited from shallow braided rivers that flowed from a series of tributary basins in the Cobalt Embayment (Figure 4.5). A thick (80 to 120 m) interval of plane laminated and ripple laminated, muddy, fine- and very-fine-grained sandstone (wacke) is present in the upper member in Parkin and Rathbun townships. This resembles the Pecors Formation south of Sudbury, and may reflect local development of lacustrine conditions within the southern part of the Cobalt Embayment due to contemporaneous tectonic subsidence along north-northwest-directed faults (Long 1987). Regional paleocurrent trends can also be used to provide information on basin evolution (Long 1995, 2004; Figure 4.5). Rousell and Long (1998) have used paleocurrent observations to document the initial uplift and collapse of the basin margins, and to negate arguments that the preservation of outliers of Huronian strata north of Sudbury (Figure 4.1) is related to rebound phenomena associated with development of a multi-ring impact structure (cf. Spray and Thompson 1995; Spray 1997).

Photo 4.5. Compound planar cross-bedded medium- to coarse-grained sandstones (subarkose) in a downstream accretionary element of a sandy braided river deposit in the Mississagi Formation, Dryden Township.

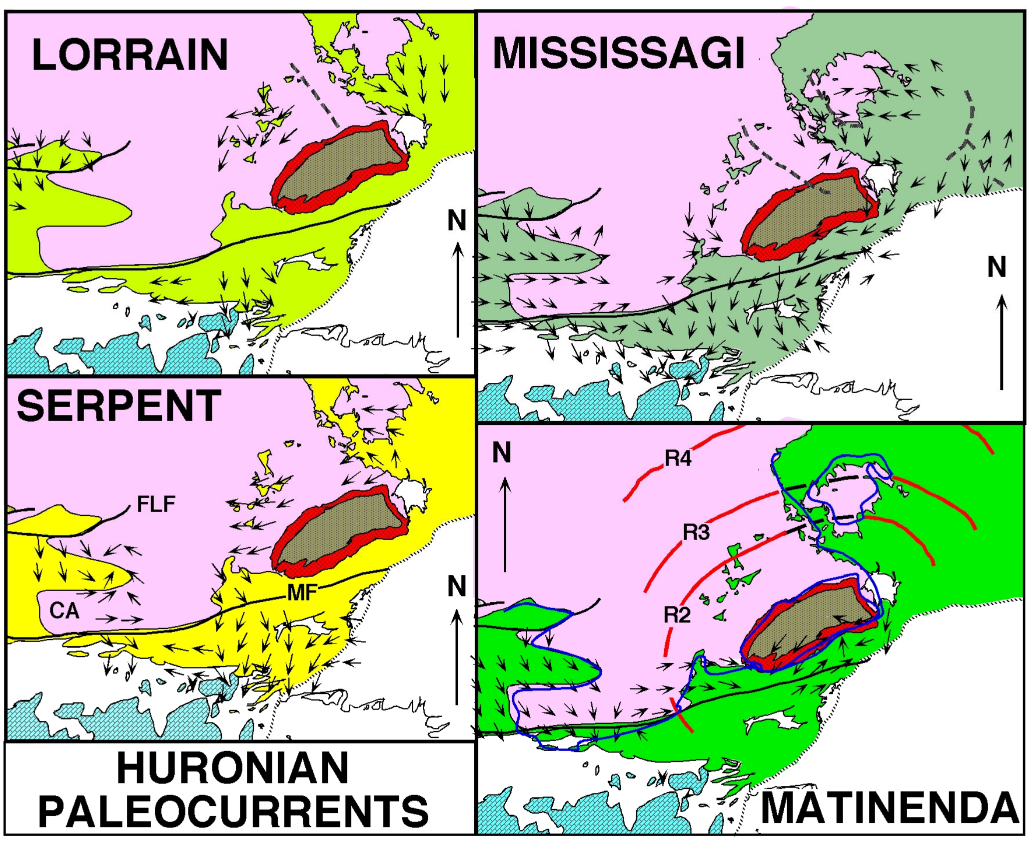


Figure 4.5. Regional paleocurrent trends in Huronian sandstone units (*from* Rousell and Long 1998). R2, R3, and R4 indicate possible location of hypothetical "superfaults" (Spray and Thompson 1995; Spray 1997). FLF = Flack Lake Fault, CA = Chiblow Anticline, MF = Murray Fault Zone. Dashed line indicates probable location of drainage divides.

QUIRKE LAKE GROUP

The Quirke Lake Group consists of three units, known as the Bruce, Espanola and Serpent formations, all of which are exposed in the Sudbury District (Figure 4.6).

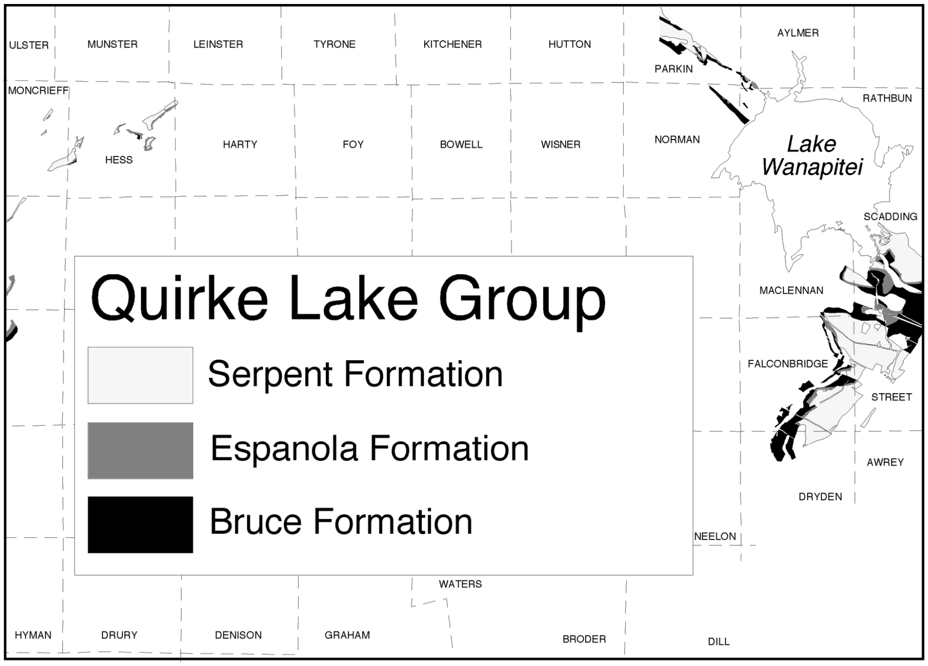


Figure 4.6. Distribution of strata in the Quirke Lake Group (*based on* Ames, Davidson et al. 2005).

Bruce Formation

The Bruce Formation is characterized by massive to weakly laminated, matrix-supported, clast-rich sandy diamictite, with thin interbeds of sandstone and mudstone. Clasts of up to boulder grade are typically well rounded, and are dispersed in a matrix of muddy, medium to coarse sandstone (wacke).

The formation is thinnest in the northwest where it forms a thin, discontinuous layer, 10 to 50 m thick, that lies directly on Archean basement rocks (Card and Innes 1981; Rousell and Long 1998). Northeast of the Sudbury Structure, the formation occurs as a more continuous sheet of diamictite, up to 760 m thick, above cross-stratified sandstones of the Mississagi Formation. South of Lake Wanapitei the formation thickens from about 200 m on the east side of Ashigami Lake in Davis Township (Dressler 1982), to less than 350 m in central Street Township, and at least 1580 m near the border of Falconbridge and Dryden townships (Debicki 1990).

The basal contact with the underlying fluvial strata of the Mississagi Formation may be sharp or may locally appear transitional due to incorporation of sandstone into the base of the formation during glacial advance. Sandstone and mudstone intervals, with local graded bedding and dispersed clasts are most common in the upper part of the formation. Where interbedded with massive diamictites they tend to be laterally discontinuous, suggesting that they were deposited beneath a floating ice-shelf (Young 1981; Photos 4.6 and 4.7).

Photo 4.6. Interbedded (muddy) sandy conglomerate (light), pebbly, sandy mudstone (centre, with 10 cm long pebble), and clast-poor diamictite (dark, at base) in exposure of the Bruce Formation, Hutton Township.

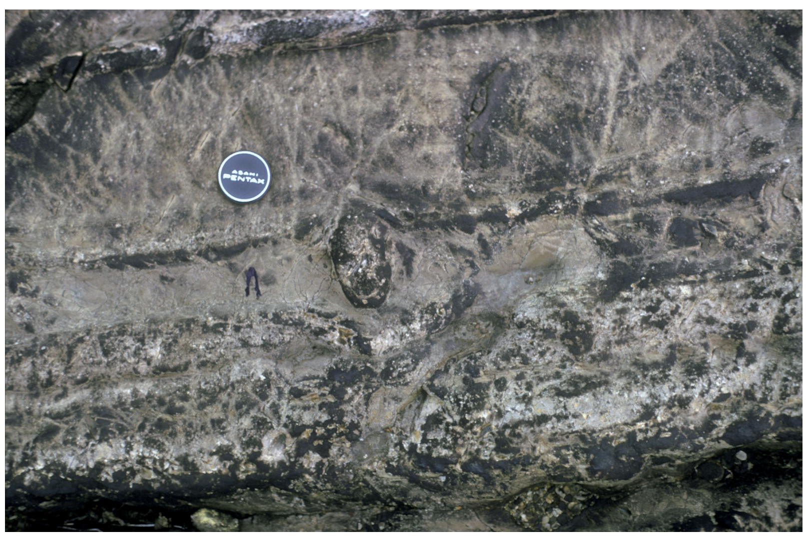


Photo 4.7. Isolated large pebble in graded laminated siltstone interval within the Bruce Formation, Dryden Township.

Espanola Formation

The Espanola Formation conformably overlies matrix-supported conglomerates (diamictites) and sandstones of the Bruce Formation. In most parts of the Sudbury District the Espanola Formation is poorly exposed. In most areas it can be divided into two or three members (Young 1973b): a lower, relatively pure laminated limestone with siltstone interbeds; a middle unit dominated by calcareous siltstone; and an upper unit of interbedded calcareous siltstone and very-fine-grained sandstone (calcareous arkose, lithic arkose and subarkose).

In the vicinity of Geneva Lake the Espanola Formation thickens from 7 m, north of the CP railway line, to a maximum of 90 m in the eastern half of Hess Township (Card and Innes 1981). At this locality (Photo 4.8), the basal 1.2 m consists of intraformational conglomerate (lumpstone) containing clasts of well-laminated dolostone, and minor siliciclastic debris. The overlying 5.8 m consists of relatively pure, flat to wavy laminated dolostone that contains poorly preserved stratiform stromatolites (Long 1976; Rousell and Long 1998). North of Lake Wanapitei the formation is between 60 and 100 m thick (Meyn 1970), and contains thick sequences of highly distorted, well-laminated limestone (calcisiltite) with thin mudstone and siliciclastic siltstone laminae. In Parkin Township this has been exploited as a building stone (floor tiles). The contortion within these beds appears to have been produced by drag-folding (intra-stratal deformation), rather than by syn-sedimentary slumping. South of Lake Wanapitei, Dressler (1982) estimates the formation to be at least 100 m thick, thickening to at least 250 m in northern Aylmer Township and central Falconbridge Township (Debicki 1990).

Photo 4.8. Thin sequence of Espanola Formation, consisting predominantly of thin- to thick-laminated (recrystallized) limestone, in exposure north of the CP railway track, near Bannerman Lake, Moncrieff Township. Underlying Archean quartz monzonite shows no signs of regolith development. Hammer rests on the unconformity and is 74 cm long.

Serpent Formation

The Serpent Formation to the south and west of the Sudbury District is characterized by pink- to white-weathering, medium- and fine-grained sandstone (feldspathic and subfeldspathic arenite), characterized by both large- and small-scale cross-stratification (Long 1976). This is in marked contrast to the lithology of the formation immediately northwest of the Sudbury Igneous Complex, where the formation can be divided into two distinct members. The lower member is characterized in the north by massive and weakly stratified cobble and pebble conglomerates (Photo 4.9), passing laterally into small pebble conglomerates and planar and trough cross-stratified sandstones south of Geneva Lake (Photo 4.10; Long 1976; Rousell and Long 1998). The conglomerates, which include both matrix- and clast-supported varieties, can be distinguished from conglomerates in the overlying Gowganda Formation by a more limited array of clast types. The upper member consists of predominantly massive, well-sorted, medium- and fine-grained sandstone.

Photo 4.9. Matrix- and framework-supported, medium to large pebble conglomerate in the Serpent Formation, north of Bannerman Lake in Moncrieff Township. These are interpreted as debris flow and stream flow deposits that formed on alluvial fans, which prograded south from active faults along the north side of the basin. Scale divisions are 10 cm.

Photo 4.10. Planar and trough cross-stratified, medium- and coarse-grained sandstone in the Serpent Formation, immediately north of the point where the Geneva Lake access road crosses the CP railway line in Hess Township.

The thickness of the formation is strongly dependant on the extent of erosion at the base of the Gowganda Formation. It thickens from a zero edge to as much as 85 m in the Geneva Lake area, and is up to 914 m thick northwest of the Sudbury Basin in Parkin Township (Meyn 1970). South of Lake Wanapitei the formation is locally more than 670 m thick in Street Township, thinning to 350 m in Scadding Township and 380 m in Aylmer Township (Dressler 1982; Debicki 1990). Calcareous sandstone in the lower 240 m of the formation at Gibson Lake in Hutton Township may represent a sandy phase of the Espanola Formation (Long 1976). Paleoflow observations in the Serpent Formation suggest major source areas north of Elliot Lake and in the Cobalt Embayment (Figure 4.5). The area now underlain by the Sudbury Igneous Complex and Sudbury Basin appears to have remained a positive area, which may have concentrated paleoflow to the west into a paleovalley system in the vicinity of the Vernon Syncline (Figures 4.1, 4.5).

COBALT GROUP

The Cobalt Group is divided into two major units: the Gowganda Formation, and the Lorrain Formation. These rest unconformably on (and locally truncate) earlier strata of the Hough Lake and Quirke Lake groups, and can be found directly overlying Archean basement north of the Sudbury Basin (Figure 4.7).

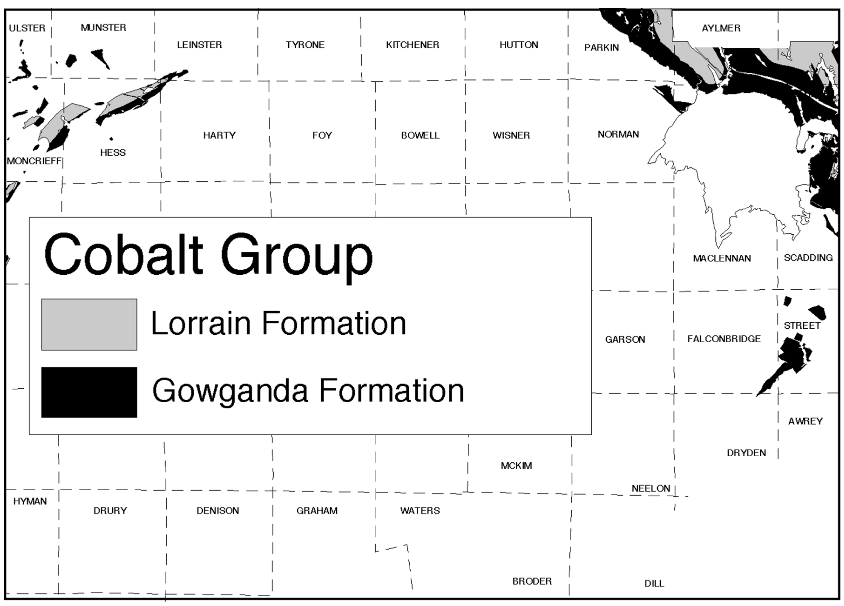


Figure 4.7. Distribution of strata of the Cobalt Group in the Sudbury District (*based on* Ames, Davidson et al. 2005).

Gowganda Formation

The Gowganda Formation is characterized by a heterogeneous sequence of framework- and matrix-supported conglomerate (including diamictites), sandstone, siltstone and mudstone with marked vertical and lateral facies changes (Photos 4.11 and 4.12). Regionally, matrix-supported conglomerates and laminated mudstones with dropstones are more abundant at the base of the sequence. Interbedded sandstone and mudstones are more common in the upper parts of the formation. The basal sequence (Coleman member) was deposited beneath a continental ice sheet, while the upper sequence (Firstbrook member) represents a prodeltaic facies equivalent of the Lorrain Formation (Long and Leslie 1986).

Northwest of the Sudbury Structure the formation is between 150 and 600 m thick and is preserved in a series of paleovalleys, above older Huronian strata, and as isolated outliers, where it typically lies in direct (erosional) contact with the Archean basement (Rousell and Long 1998). Northeast and southeast of the Sudbury Structure the Gowganda Formation forms a more continuous sheet, between 1.3 and 1.8 km thick (Meyn 1970; Long and Leslie 1986).

Photo 4.11. Thin bedded muddy sandstone with layer of pebbles and granules (iceberg-dump deposits) above a clast-poor diamictite in exposure west of Bannerman Lake, Moncrieff Township.

Photo 4.12. Granule-coated phenoplasts of mudstone in graded laminated sandstone units in the basal part of the Gowganda Formation (location as Photo 4.11).

Lorrain Formation

Strata of the Lorrain Formation are preserved locally within paleovalleys northwest of the Sudbury Structure, and in more continuous sheets north of Lake Wanapitei. In the northwest, the formation is represented by 450 to 600 m of planar and trough cross-stratified fine- to coarse-grained sandstone (arkose) with scattered pebble horizons (Card and Innes 1981; Rousell and Long 1998; Long 1976, 2004).

In the southern Cobalt Embayment the formation is at least a kilometre thick, and is characterized by cross-stratified, medium- and coarse-grained arkosic and subarkosic fluvial sandstone with minor pebble layers (Meyn 1970; Debicki 1990). Consistent paleoflow directions in the sandstones indicate that most of the formation is of fluvial origin (Figure 4.5). Bimodal-opposed cross-stratification, indicating a tidal influence, has been recorded in the top 10 m of the formation at Welcome Lake (north of the Sudbury District), but has not been seen elsewhere.

FLACK LAKE GROUP

Deposits of the Flack Lake Group include strata assigned to the Gordon Lake and Bar River formations. These are not present in the Sudbury District, but can be seen to the south in Killarney Provincial Park, to the northeast in the Cobalt District, near Welcome Lake, and to the west, north of Elliot Lake. In these areas the Gordon Lake Formation consists of interbedded ripple-laminated sandstones (quartz arenites) and red laminated mudstones, with minor intraformational conglomerate horizons. The Bar River Formation is dominated by ripple-laminated sandstones (quartz arenites) with bimodal-opposed cross-stratification.

Chapter 5

Nipissing Gabbro

L.S. Jobin-Bevans

Introduction

The Nipissing Gabbro intrudes the supracrustal rocks of the Huronian Supergroup in the Southern Province, as well as the underlying Archean granite-greenstone basement rocks. U-Pb geochronology has yielded crystallization ages of approximately 2200 Ma (2219 Ma, Corfu and Andrews 1986; 2212 Ma, Conrod 1989; 2210 Ma, Noble and Lightfoot 1992).

Morphology

Although many of the Nipissing Gabbro intrusions have been somewhat metamorphosed and deformed, some of the intrusions are thought to have retained their primary morphologies, as reflected by the current outcrop patterns. These patterns include tabular intrusions, open-ring structures, and massive, irregular-shaped bodies, and are interpreted to represent four main morphologies (Jambor 1971; Buchan et al. 1989): 1) undulating sills and dikes; 2) concordant homogeneous sills; 3) cone sheets or ring dikes; and 4) lopolith-like or thick, stock-like bodies. Horst-and-graben structures (block-faulting) also appear to play a major role in determining the stratigraphic level of a gabbro body, particularly in the areas southwest, south and east of Sudbury.

The majority of Nipissing Gabbro intrusions are less than 1000 m thick and occur as roughly horizontal sheets, as undulating sills (basins and arches) or as subvertical dikes (Hriskevich 1968; Jambor 1971; Conrod 1988, 1989). Disseminated to massive Cu-Ni-PGE sulphide mineralization, in these types of intrusions, is concentrated within the basin or limb portions, whereas pods of dominantly massive pyrrhotite occur within the arches (Figure 5.1). Much of the mineralization is associated with an orthopyroxene gabbro unit (Figure 5.2) which is, in general, greater than 100 m in thickness (Lightfoot and Naldrett 1996; Jobin-Bevans et al. 1998, 1999). Arcuate and open-ring exposures of Nipissing Gabbro, described by Buchan et al. (1989) as cone sheets, comprise a third form of intrusion. These forms are distinguished by structural features in surrounding sedimentary rocks that suggest the gabbro intrusions were emplaced as shallow (< 50°), inward-dipping, cone-shaped bodies that are tens of metres to several hundred metres thick (Jambor 1971; Lovell and Caine 1970; Jobin-Bevans et al. 1998).

These types of intrusions contain disseminated and blebby sulphides hosted in orthopyroxene gabbro, occurring within a few hundred metres of the basal contact of the intrusions. The fourth type of intrusion, the lopolithic-like form (i.e., saucer-shaped), is rare and is interpreted to represent deeper "feeder" systems to the stratigraphically higher sill, dike and cone-sheet type of intrusions. These deeper exposures, which are fault bound on a regional scale, are thought to have been exposed through uplift along the bounding fault lines (Dressler 1979; Innes and Colvine 1984; Jobin-Bevans et al. 1998). In the lopolithic-like form, disseminated, semi-massive and massive sulphide mineralization is hosted by orthopyroxene gabbro within tens of metres of the footwall sedimentary rocks, and within topographic irregularities along the footwall contact.

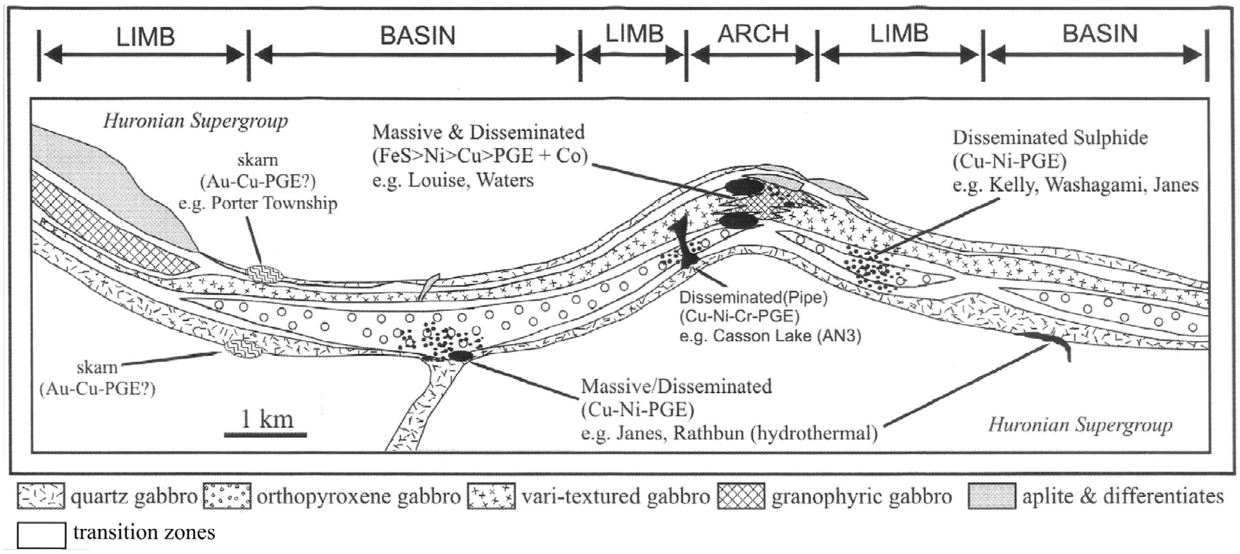


Figure 5.1. Schematic cross-section of a model of an undulating sill of Nipissing Gabbro showing the distribution of lithologies and locations of different types of mineralization (*after* Hriskevich 1968; Lightfoot et al. 1987; Lightfoot and Naldrett 1996).

General Stratigraphy

The general stratigraphy of Nipissing Gabbro, which has been described in previous reports (e.g., Hriskevich 1968; Conrod 1988; Lightfoot and Naldrett 1996), is remarkably consistent within individual intrusions, particularly where the intrusion is well differentiated (Figure 5.2). Small-scale folds are rare in Nipissing Gabbro intrusions and any recognizable folding appears to be related to large-scale, regional events. Consequently, the observed stratigraphy within individual intrusions is interpreted to reflect the original arrangement of layers within magma chambers.

The varying proportions of differentiates and/or hybrid rocks (granophyre, aplite, granite, granodiorite) versus orthopyroxene gabbro and gabbro, in any given intrusion, may reflect the reactivity of the country rock with the magma and the degree of local assimilation and/or contamination. Alternatively, the differing proportion may reflect the relative position of the section and/or its current level of erosion, and therefore surface exposure through the intrusion, if convection played a major role in the development of the intrusion (Conrod 1988; Lightfoot and Naldrett 1989). Previous work by Jambor (1971), Conrod (1988) and Lightfoot and Naldrett (1996) produced a type-stratigraphy for Nipissing Gabbro intrusions. Jobin-Bevans (2004) integrated geological surface mapping, ground geophysical surveys and diamond-drill hole data, allowing for refinement of the stratigraphy into 5 major units (Figure 5.2). Most of the major units have good lateral and vertical continuity but individual layers within these major units can show significant petrologic variations in modal mineralogy and/or thickness.

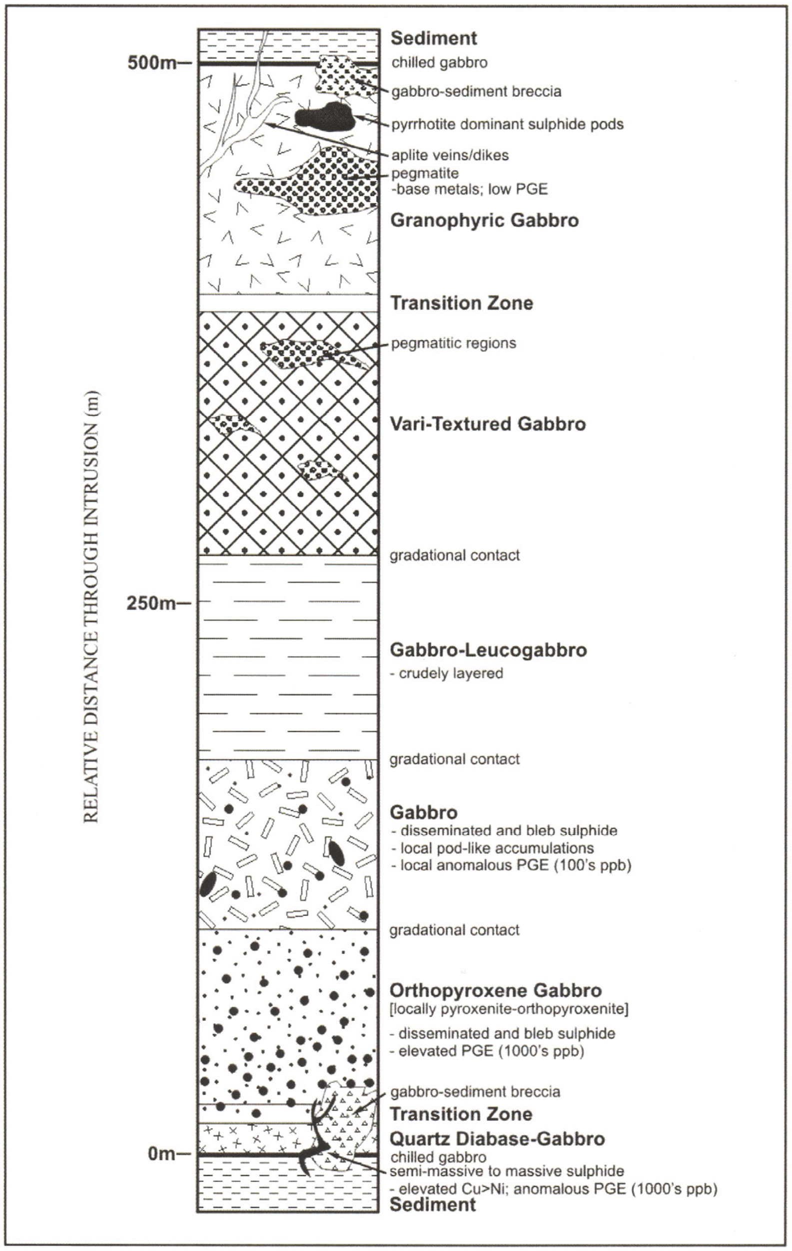


Figure 5.2. Type-section of a well-differentiated Nipissing Gabbro intrusion showing the typical sequence of lithologies and the location of the various types of sulphide mineralization (*after* Lightfoot and Naldrett 1996).

General Geochemistry

Geochemical characteristics of Nipissing Gabbro intrusions have been described by several authors including Jambor (1971), Card and Pattison (1973), Conrod (1989), Rowell and Edgar (1986) and Lightfoot and Naldrett (1996). Rocks from the intrusions are dominantly tholeiitic and sub-alkalic, with evolved rock types and differentiated intrusions trending toward calc-alkalic affinities (Lightfoot and Naldrett 1996).

Data from 100 samples of chilled quartz gabbro (the least differentiated gabbro), collected from 15 different Nipissing Gabbro intrusions, suggest that the parental magma to the Nipissing Gabbro suite was relatively uniform in composition. Characteristic features of the magma include: elevated SiO_2 (50.0-51.5 wt %); moderate MgO (8-9 wt %); strong light rare-earth element (LREE) and large ion lithophile element (LILE) enrichment (La/Sm = 2.5-3.5; Th/Nb = 0.7-0.9); $\varepsilon\ Nd^{CHUR}$ of -2.7 to -5.9 and low $^{143}Nd/^{144}Nd$, indicating enriched mantle or crustal source; and conspicuously negative Nb, Ta, P and Ti anomalies (relative to LILE and LREE), suggesting crustal interaction and/or contamination (Lightfoot and Naldrett 1989, 1996; Lightfoot et al. 1993).

Based on the geochemical characteristics outlined above, and on outcrop patterns, the Nipissing Gabbro supposedly represents the intrusive portion of an eroded continental flood basalt. Magmas apparently cut through Archean basement rocks and sedimentary rocks of the Huronian Supergroup as dikes, then spread laterally through the Huronian rocks as sills (Lightfoot et al. 1986, 1987; Lightfoot and Naldrett 1996).

Mineralization

The region between Sault Ste. Marie and Cobalt contains numerous metallic mineral occurrences (Cu-Ni-Co-PGE-Ag-Au) which are associated with Nipissing Gabbro, the East Bull Lake mafic suite and, to a lesser extent, the Huronian Supergroup. Early work by Card and Pattison (1973), together with current field observations, indicates that mineralization in Nipissing Gabbro varies in type and style across the Southern Province as follows (Figure 5.3):

1) in the northeast (Cobalt area) Co-Ag-(PGE-Ni) sulphides and sulpharsenides occur in quartz-carbonate veins;

2) predominantly contact-related Cu-Ni-Co-(PGE) are most conspicuous in the region immediately southwest of Cobalt (Lake Temagami area);

3) intrusion-hosted Ni-Cu-PGE-(Au) sulphides occur in the regions immediately northeast, south and west-southwest of the City of Greater Sudbury;

4) contact-related and intrusion-hosted Ni-Cu-Co-PGE sulphides are common in the areas southwest of Sudbury (most common but not restricted to the area south of the Murray Fault Zone) and in the Elliot Lake area; and

5) secondary (late) quartz-carbonate vein-associated Cu sulphides are most common in the Nipissing Gabbro between Blind River and Sault Ste. Marie.

Lightfoot et al. (1987) noted that this variation in type and style of mineralization appears unrelated to differences in lithology, degree of metamorphism, or level of intrusion of the Nipissing Gabbro. Sulphide occurrences in Nipissing Gabbro, in the study area, have been described as magmatic (Lightfoot et al. 1986, 1987, 1993) and as hydrothermal (Finn et al. 1982; Rowell and Edgar 1986).

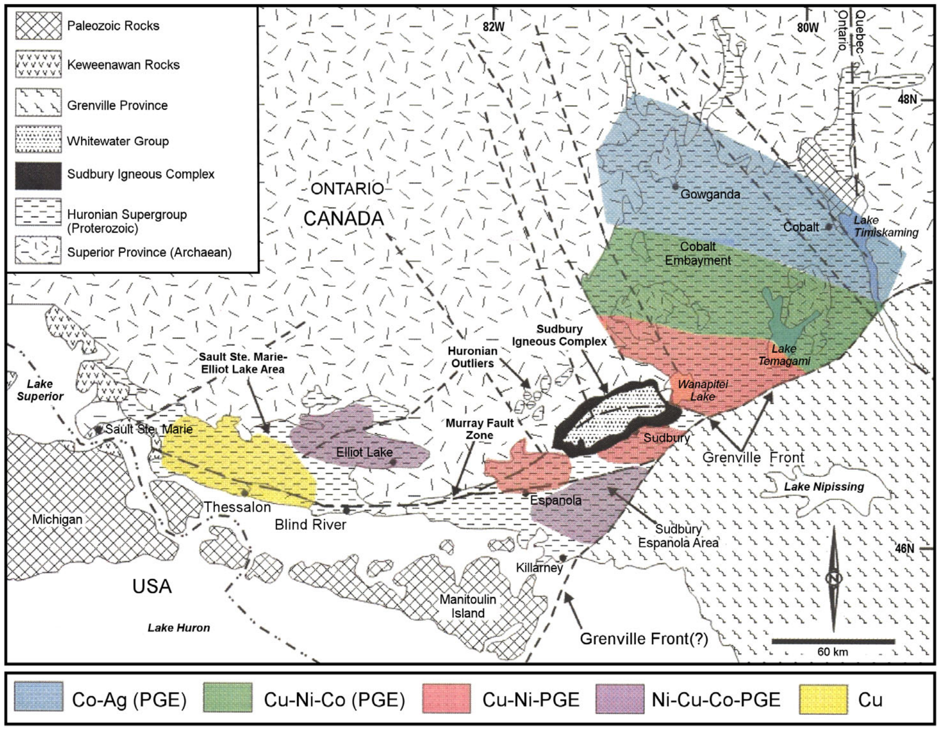


Figure 5.3. Regional distribution of mineralization in Nipissing Gabbro intrusions between Lake Temagami and Sault Ste. Marie (*after* Card and Pattison 1973). See text for details.

PGE-associated sulphide mineralization is dominated by chalcopyrite and pyrrhotite with subordinate pentlandite and pyrite (James, Jobin-Bevans et al. 2002). The mineralization occurs as three main types:

1) disseminated sulphide mineralization with the most consistent and persistent PGE contents (i.e., 500-6000 ppb PGE) and moderate base metal concentrations (i.e., roughly 0.75-1.0% Cu + Ni);

2) contact-associated, disseminated to semi-massive sulphide mineralization with very high PGE (i.e., 8000-123 000 ppb PGE) and base metal contents (i.e., 1.5-3.0% Cu + Ni); and

3) massive sulphide mineralization with typically low PGE (i.e., < 250 ppb PGE) and high base metal content (>3% Cu + Ni).

Pd:Pt ratios in mineralized samples are about 5:1, and in non-mineralized or poorly mineralized samples are about 2.5:1; very high PGE concentrations (i.e., >7000 ppb) tend to have high Pd:Pt ratios (i.e., >10:1); Cu:Ni ratios are commonly 2:1 but may be as high as 40:1 in remobilized sulphide. Background concentrations of PGE, Au, Cu and Ni are estimated to have maximum values of about 40 ppb Pd, 32 ppb Pt, 9 ppb Au, 94 ppm Cu and 376 ppm Ni; these arithmetic averages are based on the analysis of 23 non-mineralized (low-S and <100 ppm Cu) samples (James, Jobin-Bevans et al. 2002).

Much of the known and potentially economic PGE sulphide mineralization occurs within the lower to middle parts of the orthopyroxene gabbro unit (e.g., Lightfoot and Naldrett 1996; James, Jobin-Bevans et al. 2002), and is generally located within the lower one-third of the stratigraphy. This style, indicated by type 1 (above), is best described as stratabound and more precisely as stratiform. For the most part it consists of 1-5% fine- to medium-grained, disseminated and blebby (up to 7 cm in diameter) chalcopyrite, pyrrhotite and pentlandite, with subordinate net-textured sulphide. Blebby sulphides commonly show segregation textures of pyrrhotite and chalcopyrite either as blebs with pyrrhotite cores rimmed by chalcopyrite or as blebs with half pyrrhotite and half chalcopyrite. Lightfoot and Naldrett (1996) reported similar sulphide segregation textures and referred to the blebby sulphide as globules. This style (type 1) of mineralization appears to hold the most promise for large-tonnage, moderate-grade, open-cast mining. Semi-massive (25-80% total sulphide) and massive (>80% total sulphide) sulphide is rare, but where observed, occurs at or near the basal contact of the intrusion, toward the base of the orthopyroxene gabbro unit, or within the basal gabbro-sediment breccia (James, Jobin-Bevans et al. 2002).

Lightfoot et al. (1987), in assuming chilled margin samples represent parental magma compositions, suggested that the uniform composition of the parental magma indicated limited contamination of the magma en route through the crust, or through the assimilation of local country rocks. This implies that magmatic sulphide segregation was not triggered by crustal contamination from local country rocks but rather from large-scale, homogeneous contamination. It is therefore likely that the dominant control on PGE-bearing sulphide mineralization is that of *in situ* normal fractional crystallization within individual bodies, with PGE-rich sulphide precipitation accompanied by orthopyroxene ± olivine crystallization in the lower orthopyroxene gabbro unit.

Emplacement Model for Nipissing Gabbro

Lightfoot et al. (1987, 1993) and Conrod (1988, 1989) have suggested that the calc-alkalic characteristics of Nipissing Gabbro are the result of pre-emplacement enrichment of LILE (Cs, Rb, Ba and Sr) due to interaction with a recycled, Archean crustal component, or continental crustal contamination as the magmas evolved in deep crustal reservoirs (i.e., staging chambers). Lightfoot et al. (1993) surmised that the parental magmas differentiated at depth and precipitated olivine (conspicuously absent from almost all Nipissing bodies), with subsequent emplacement as uniform, low-Mg magmas. The LREE enrichment signature was acquired from continental crust, perhaps during migration from the mantle to the surface (Lightfoot et al. 1993). In considering the homogeneous nature of the parental magma and the large volume of assimilated crustal material (>20%) that would be required to produce the observed REE and trace element signatures, Lightfoot et al. (1993) suggested that the source characteristics of the magmas may have been acquired as a result of subduction events related to the Archean Kenoran Orogeny (2500 Ma). Tomlinson (1996) and Tomlinson et al. (1999) arrived at similar conclusions for magmas of Early Huronian volcanic rocks of the Elliot Lake Group. They suggested that the principal magma source inherited its geochemical signature from metasomatically enriched upper mantle material, which was geochemically modified as a result of subduction-accretionary events associated with the Kenoran Orogeny.

Work by several authors, including Lightfoot et al. (1987), Conrod (1989) and Lightfoot and Naldrett (1989), suggests that petrological variations within the intrusions are largely controlled by the coupling of post-emplacement fractionation and assimilation processes or assimilation-fractional crystallization (AFC). Lightfoot and Naldrett (1989), in applying the AFC model to the Kerns sill (Lake Temagami area), determined that assimilation and fractionation worked in concert to produce a distinct signature. The least fractionated samples are the least contaminated, and the most fractionated samples (characterized by low Mg-number, Ni and Cr and by high incompatible element concentrations) show the largest amount of contamination (characterized by higher Th/Zr, La/Zr and U/Zr and lower $^{143}Nd/^{144}Nd$). Their model, also used by Conrod (1989), suggested that early crystallization is accompanied by little or no contamination (chilled diabase; basal quartz diabase; hypersthene diabase), followed by moderate contamination (vari-textured diabase), and finally by substantial contamination (granophyric diabase) in the upper portions of the sills, accompanied by the assimilation of hanging wall sedimentary rocks and the formation of aplitic rocks.

Chapter 6

Whitewater Group

D.E. Ames, J.A. Stoness and D.H. Rousell

Introduction

Rocks of the Whitewater Group, approximately 2900 m thick, occur only within the Sudbury Basin. However, at the time of deposition the rocks of this group probably extended well beyond the present basin. The group consists of four conformable formations which are, from oldest to youngest, Onaping, Vermilion, Onwatin and Chelmsford formations (Figure 1.3). Coleman (1905) was the first to assign local names to the units of the group: Trout Lake conglomerate, Onaping tuff, Onwatin slate and Chelmsford sandstone. The Trout Lake conglomerate is now considered a basal unit of the Onaping Formation (Gibbins 1997). According to Thomson (1957), W.H. Collins, circa 1912, suggested the name "Whitewater series" after a lake of that name (Figure 1.2).

Fairbairn et al. (1968) obtained a whole-rock Rb/Sr isochron age of 1740 Ma for the Chelmsford Formation. Deposition / emplacement of the group predated intrusion of olivine dikes of the Sudbury Swarm at 1238 Ma (Krogh et al. 1987) and postdated, in whole or in part, the age of the Sudbury Event taken as the isotope age of the Sudbury Igneous Complex (SIC) at 1850 Ma (Krogh et al. 1984). Note that the granophyre, the uppermost unit of the SIC, intrudes the Onaping Formation. According to Gibbins (1997), the deposition of the Onaping Formation occurred over a prolonged period of time. If the formation is the product of meteorite impact, then the Sudbury Event must be somewhat older than 1850 Ma. Many investigators view the SIC as an impact melt sheet (e.g., Dence 1972; Dietz 1972; Grieve et al. 1991; Milkereit, Green et al. 1992; Golightly 1994; Therriault et al. 2002; Mungall et al. 2004; Naldrett 2004). This interpretation is difficult to reconcile if the SIC was emplaced beneath the Onaping Formation. Another possible scenario is the formation of an impact melt sheet which later differentiated into Sublayer, norite and quartz gabbro, followed by the development of the Onaping Formation by hyaloclastic processes and finally, by the intrusion of the granophyre between the quartz gabbro and the base of the Onaping Formation. The Onwatin and Chelmsford formations represent the resumption of "normal" sedimentary processes.

Onaping Formation

INTRODUCTION

The Onaping Formation is named after the Onaping River. Although a type section has not been defined, well exposed and accessible outcrops occur along Highway 144, immediately southwest of the Onaping River. The formation, approximately 1400 m thick, consists of a series of breccias and igneous-textured rocks of contentious origin. The base of the formation is intruded by granophyre of the underlying SIC, whereas the top grades into carbonate and argillaceous rocks of the overlying Vermilion and Onwatin formations. Over the years the Onaping Formation has been interpreted as follows: 1) volcanic (Bonney 1888; Burrows and Rickaby 1930; Thomson 1957; Williams 1957; Stevenson 1972; Muir 1984);

2) impact-induced volcanism (Dietz 1964; Thomson 1969; Muir 1983); 3) fall-back breccia from meteorite impact (French 1968; Dence 1972; Peredery 1972; Peredery and Morrison 1984); or 4) fragmentation of “melt” by repeated interaction with water (Gibbins 1994, 1997). However, detailed mapping over the last two decades indicates that the rocks of the Onaping Formation have a distinct stratigraphic sequence that consists of fall-back breccia, hydroclastic breccia, debris flows and plume-collapse breccias with reworked material in the upper 170 m (Ames 1999). Thus, the formation represents a succession of glass-rich breccias and coeval hypabyssal intrusions that have been hydrothermally altered to a variable degree. Hypabyssal rocks with an igneous texture are classified as quartz diorite, whereas those with a glassy texture, interpreted as shock-melted crust, are called andesite.

Evidence for an impact origin of the Onaping Formation includes shock metamorphic features in quartz, feldspar and zircon (French 1967; Peredery 1972); the presence of diamonds (Masaitis et al. 1999); and an iridium anomaly (Mungall et al. 2004). Peredery (1972) reported shatter cones in clasts of country rocks present in the formation.

STRATIGRAPHY

Historically, the Onaping Formation was classified on the basis of rock colour, that is: Gray, Green and Black members (Peredery 1972; Avermann 1999). The colour relates to the absence of carbon (gray) and the presence of chlorite-actinolite (green) or carbon (black). However, recent investigators (Gibbins 1994, 1997; Ames et al. 1997, 1998, 2002) recognized that rock colour does not reflect depositional units. Accordingly, they have subdivided the formation into three members on the basis of stratigraphy rather than colour. These members are, from bottom to top, Garson, Sandcherry and Dowling (Figure 6.1). See Table 6.1 for a comparison of nomenclature. The basis for the revised nomenclature is illustrated by Figure 6.2, which represents a map of the Onaping Formation in an area northeast of Highway 144. There, the Sandcherry Member is mainly grey, but is locally black, whereas the overlying Dowling Member is largely black but may be locally grey. Thus the contact between carbon-poor and carbon-bearing rocks transects lithologic contacts and is gradational over a distance between 5 to 20 m. The carbonaceous material consists of “organic” carbon or kerogen and resides in the black, semi-opaque matrix.

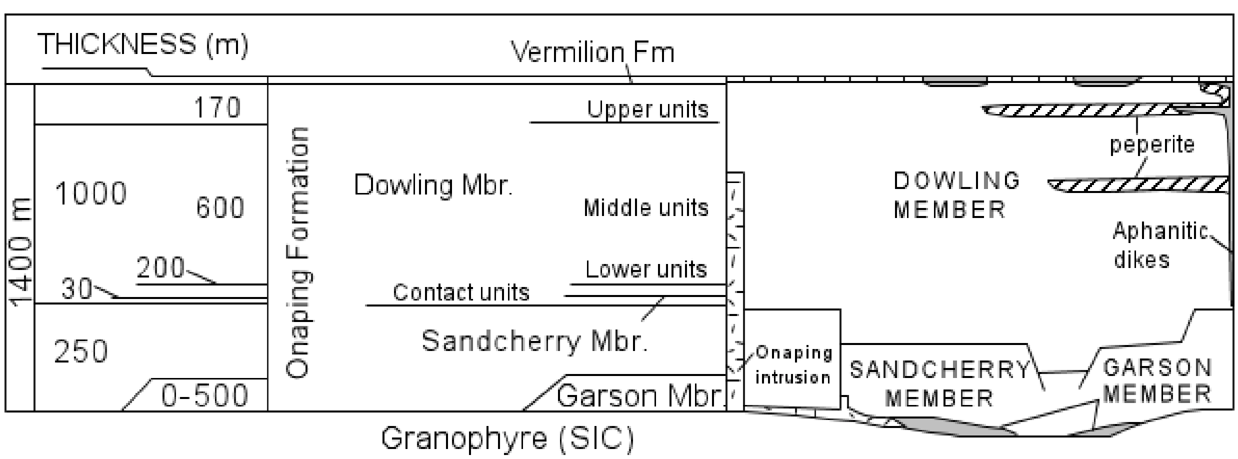


Figure 6.1. Stratigraphic section for the Onaping Formation.

Table 6.1. Comparison of revised with previous stratigraphic nomenclature for the Onaping Formation.

Revised Nomenclature	Previous Workers
ONAPING FORMATION	**ONAPING FORMATION**
Dowling Member	Dominantly Black Member (Peredery 1972; Muir and Peredery 1984) and Green Member (Avermann 1999) and minor Gray Member
Upper unit	Upper Black Member (Avermann and Brockmeyer 1992), part thereof
Middle unit	Upper Black Member (Avermann and Brockmeyer 1992), part thereof
Lower unit	Lower Black Member (Avermann and Brockmeyer 1992)
Contact unit	Green Member (Avermann 1999); chlorite shard horizon (Muir and Peredery 1984)
Sandcherry Member	Gray Member (Peredery 1972; Muir and Peredery 1984; Stevenson 1961, 1963, 1990); also includes minor Black Member
Equant shard unit	
Fluidal fragment unit	Partial melt bodies (Peredery 1972)
Garson Member	
Quartzite breccia unit	Basal Member in the South Range (Muir and Peredery 1984); Quartzite breccia (Stevenson 1961, 1963, 1990)
INTRUSIVE ROCKS	
Aphanitic andesitic dikes	Some melt bodies (Peredery 1972; Muir and Peredery 1984)
Onaping intrusion	Basal intrusion (Gibbins 1994; Ames et al. 1998, 2002; Ames 1999; Ames and Gibson 2004a-e)
Semi-concordant bodies	Basal Member in the North Range (Muir and Peredery 1984)
Discordant bodies	Some melt bodies (Muir and Peredery 1984)

The Garson Member is well exposed in the eastern part of the South Range, particularly in the vicinity of Garson Lake. The member consists of thick-bedded to massive fragmental rocks that are dominated by clasts of quartzite (20 to 85% by volume) derived from rocks of the Huronian Supergroup. The clasts range in size from fragments less than 6 cm in diameter to blocks over 50 m across. The Garson Member also contains up to 15% vitric andesite lapilli and as much as 5% vitric andesite bombs and blocks. Depositional units of lapillistone or coarse tuff breccia may be massive or display reverse graded bedding.

The Sandcherry Member is characterized by less than 60 modal % vitric fragments known as “fluidal fragments” or “blocky equant shards”. Carbon is present locally as isolated pockets and averages 0.3 wt % in the pockets. Units are massive to crudely bedded, contain less than 35 modal % matrix and are block- and bomb-rich. Units that consist of equant shards are interpreted as fall-back breccia. These units are intruded by andesitic shock-melt along northeast- and northwest-striking fractures that form fluidal breccia complexes similar to hydroclastic breccia and peperite (Figure 6.3).

The Dowling Member comprises 75% (approximately 1000 m) of the Onaping Formation. Carbon is pervasive throughout much of the member (0.40 to 0.47 wt %). Units of the member tend to be laterally continuous with strike lengths from 2 km to greater than 10 km and, in general, are of uniform thickness (25 to 300 m). Distinct contact, lower, middle and upper units can be recognized. The lower units are interpreted as a combination of erosive mass flows and debris flows. They were deposited on the Sandcherry Member, which formed an irregular, fractured and tectonically unstable topography (Figure 6.2), perhaps the product of crater collapse. The discontinuous and faulted units at the base of the formation are mantled by volumetrically large pyroclastic-like, iridium-bearing deposits (middle units) derived from the collapsing vapour plume. Note that the middle and upper units contain approximately 0.4 wt % iridium. Rocks of the Dowling Member have a greater matrix content and a different shard morphology than those of the Sandcherry Member. Shards in the Dowling Member (<40% by volume), are altered to actinolite-chlorite and appear wispy and lenticular to plate-like, in contrast to the blocky shards of the Sandcherry Member.

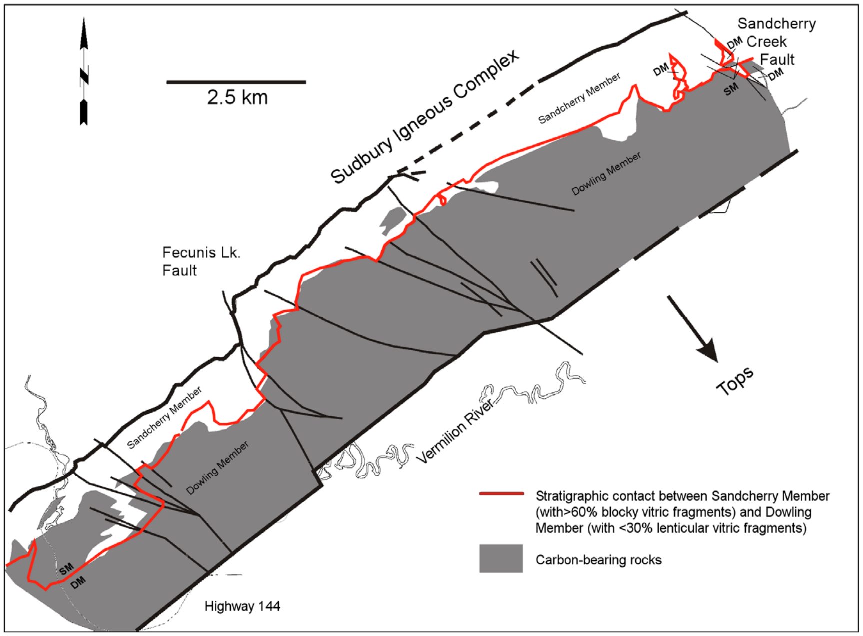


Figure 6.2. Map of the Onaping Formation, in a portion of the North Range, showing the contact between the Sandcherry Member (fall-back and autoclastic breccia) and the overlying Dowling Member (impact plume-collapse breccia). Note the crosscutting distribution of carbon. SM = Sandcherry Member, DM = Dowling Member (*after* Ames 1999).

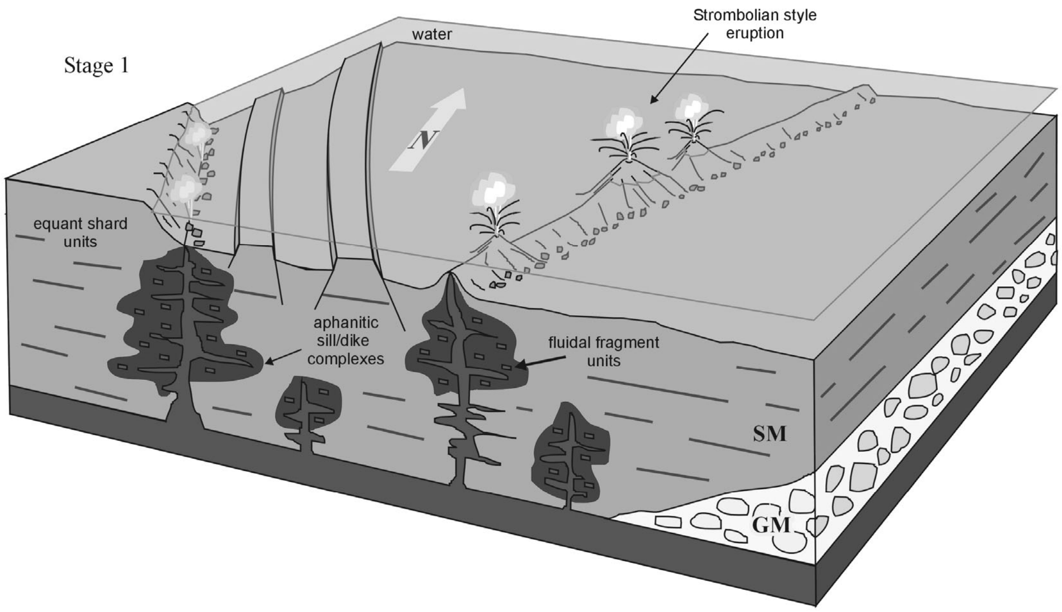


Figure 6.3. Block diagram illustrating a reconstruction of the emplacement of the Sandcherry (SM) and Garson (GM) members of the Onaping Formation (*after* Ames 1999).

The contact unit at the base of the Dowling Member is incipiently welded. The unit is interpreted to have been deposited hot, possibly in a subaerial or shallow subaqueous environment, during a tectonically unstable period in the evolution of the Sudbury impact crater. The contact unit apparently represents the explosive introduction of andesitic impact melt at the surface and, in this regard, is similar to the emplacement of volcanic pyroclastic flows.

The lower units are poorly sorted channel-sheets of considerable thickness, deposited as debris flows, synchronous with faulting, during the rapid collapse of the crater on top of the fault-controlled contact units.

The middle units are interpreted to have been deposited as a series of frequent, thin (1 to 10 m) "pulses" of material, which tends to be better sorted and finer (block- and bomb-poor) than that of the lower units. The middle units were locally enriched in iridium (Ames et al. 2002; Mungall et al. 2004) and formed during the collapse of the vapour plume. Thus, deposition of the laterally continuous and uniform middle units marks a change from a tectonically unstable environment during deposition of the lower units to a tectonically stable environment during deposition of the middle units.

The influx of water into the basin is evident at approximately 300 m above the base of the middle units. Sedimentary structures in the middle and upper units are evidence of major reworking. The upper units represent the final deposition of material. They consist of fine ejecta and iridium-enriched fallout material with subsequent reworking and mixing (Mungall et al. 2004). Upward fining, the shallowing of primary dips and the presence of turbidites all suggest the deepening and subsidence of the Sudbury Basin.

Autobrecciated aphanitic andesite dikes and sills cut units of the lower Sandcherry Member, whereas glassy andesite dikes intrude the Onaping Formation at various stratigraphic levels with vesicularity increasing upward in the section. Vitric blocks and bombs are also andesitic in composition, regardless of the stratigraphic position (Ames et al. 2002). Peperite, formed by the mixing of the dikes (interpreted as melt) and tuff of the upper Dowling Member, apparently occurs only in the southwest corner of the basin, near the Vermilion and Errington mines. The peperite constrains the timing of the emplacement of melt into the uppermost crater-fill material.

HYDROTHERMAL SYSTEM

Heat from the SIC (1850 Ma; Krogh et al. 1984) generated a hydrothermal system (1848 Ma; Ames et al. 1998; Ames, Jonasson et al. 2005) which overprinted the overlying Onaping Formation (Figure 6.4). Regional seawater convection, melt degassing and subsequent fluid/water interaction produced semiconformable alteration zones of albite, chlorite and carbonate with later silica overprinting. In addition, a low temperature (semiconformable ?) potassic alteration affects the top of the Onaping Formation. The carbonate zone, mainly calcite, increases in intensity upward. Note that the same hydrothermal system produced the Zn-Pb-Cu-Ag-Au Vermilion–Errington deposits, as well as widespread Zn-Pb-Cu mineralization in the Vermilion Formation.

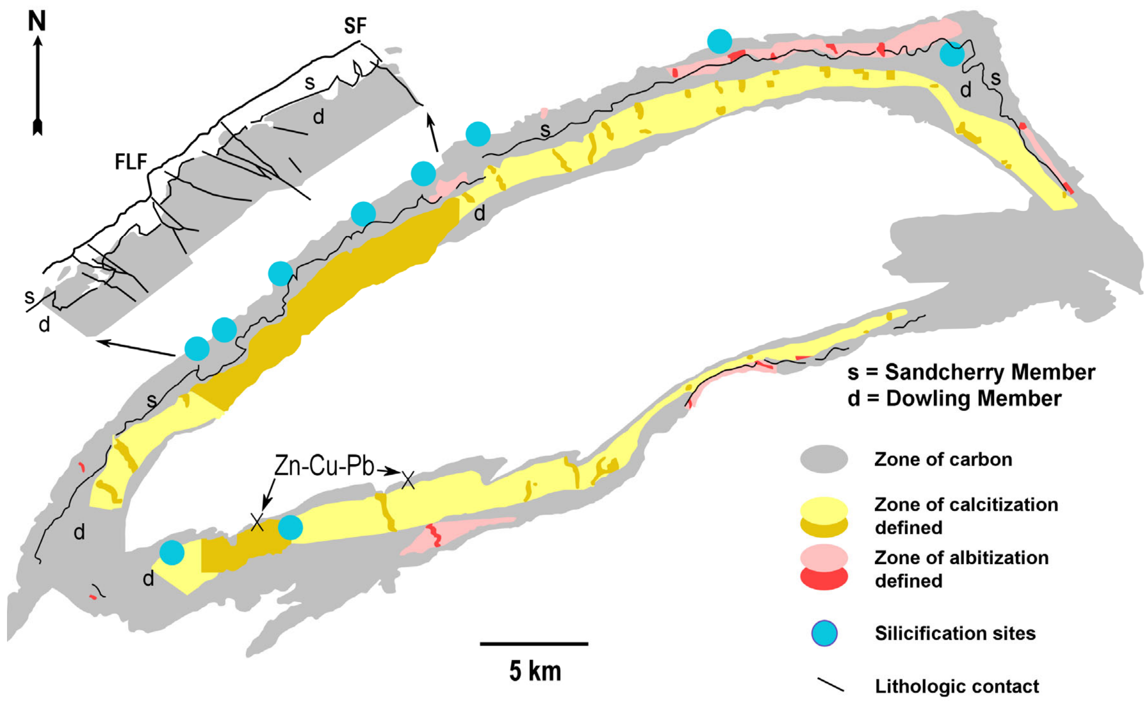


Figure 6.4. Regional distribution of hydrothermal alteration in the Onaping Formation (*from* Ames et al. 1998). FLF = Fecunis Lake Fault, SF = Sandcherry Creek Fault.

Vermilion Formation

Martin (1957) named the Vermilion Formation after a large lake in the southwest corner of the Sudbury Basin (Figure 1.2). The formation conformably overlies the Onaping Formation, is conformably overlain by the Onwatin Formation and hosts Zn-Pb-Cu deposits (see Chapter 13). Later investigators (Dressler 1984a; Rousell 1984b, 1984c; Dressler et al. 1991) followed the suggestion of Arengi (1977) and relegated the unit to a member of the Onwatin Formation because of the lack of evidence for lateral continuity. Extensive exploration drilling by Falconbridge Ltd. in 1991-1992 indicated that the Vermilion unit can be traced basin-wide and, accordingly, should be restored to formation status (Stoness 1994; Gibbins 1997). Exposures of the unit occur only in the southwest part of the basin, in the vicinity of the three former shafts of the Errington Mine. Stoness (1994) divided the Vermilion Formation (average thickness 13.5 m) into three members :1) Lower Carbonate Member; 2) Grey Argillite Member; and 3) Upper Carbonate Member.

The Lower Carbonate Member may be subdivided into a proximal and distal facies with respect to the distance from the Zn-Pb-Cu deposits (Stoness 1994). The proximal facies (3 to 30 m thick) consists of fractured and brecciated local mounds of grey to pink, laminated, colloform carbonate material. Several generations of hydrothermal carbonate breccias were apparently infilled by colloform carbonate in a manner similar to other fresh-water sinter deposits. At the Vermilion and Errington mines the carbonate breccia is locally silicified and is referred to as chert breccia (Paakki 1992). The distal facies comprise dark brown, finely laminated, carbon-rich (approximately 1.24 wt % organic carbon) silty carbonate. The dark brown colour is due to 3 to 10% Mn, while lighter layers lack Mn. The distal facies was apparently deposited by turbidite flow carrying detrital carbonate from the mounds. Thicknesses are 1.6 m in the South Range and 0.5 to 0.9 m in the North Range.

The Grey Argillite Member has an average thickness of 20 m in the South Range and 10 m in the North Range (Stoness 1994). It comprises a series of pyrite-bearing (<1%), laterally continuous, grey siltstone beds, each of which is fining upward. The average thickness of the beds is 10 cm. Local iron-carbonate laminae, 0.5 to 1.0 cm thick, occur within the siltstone beds. The absence of hemipelagic interbeds indicates that these distal turbidites were deposited without an appreciable break between individual turbidite events.

The Upper Carbonate Member is a thin (0.2 to 0.8 m), discontinuous unit comprising two to five sets of concretionary carbonate nodules in a mudstone matrix that is rich in organic carbon (Stoness 1994). The nodules, referred to as ooids by Arengi (1977), range in diameter from less than 0.5 mm at the base of the member to 20 mm at the top of the member. The member was probably deposited at the seawater-sediment interface by rising CO_2-rich hydrothermal fluids. The same hydrothermal activity may have resulted in the emplacement of the Zn-Pb-Cu deposits in the Lower Carbonate Member.

Onwatin Formation

The Onwatin Formation is named after a small lake (actually a widening of the Vermilion River) located at the northeastern end of the Sudbury Basin (Figure 1.2). The formation does not outcrop in the vicinity of the lake. Although the best outcrops are at the southwestern part of the basin, none display more than a few metres of continuous exposure. A type-section has never been defined. The formation has been described by numerous investigators including Coleman (1905), Burrows and Rickaby (1930), Thomson (1957), Sadler (1958), Beales and Losej (1975), Arengi (1977), Rousell (1984b), Rousell et al. (2002) and Stoness (1994). Estimated thickness of the formation ranges from 1100 m (Coleman 1905), to 900 to 1410 m (Arengi 1977), to 600 m (Rousell 1975; Stoness 1994; Gibbins 1997).

The Onwatin Formation consists mainly of carbonaceous and pyritic, massive to laminated argillite and siltstone with minor greywacke. The formation weathers recessively so that much of the flat farmland in the basin is underlain by the Onwatin Formation. The free carbon content ranges from 0.26 to 4.05% (Arengi 1977). Metamorphism has locally concentrated the carbon into anthraxolite veins, a lustrous black material that contains approximately 95% carbon (Burrows and Rickaby 1930). See Chapter 13 for a description of mineralization in the Onwatin Formation.

The Onwatin Formation represents pelagic sediments which were deposited in a deep, restricted basin that probably extended well beyond the present Sudbury Basin. The abundant carbonaceous material and pyrite suggests that bottom waters were stagnant and anoxygenic (Rousell 1984b).

Chelmsford Formation

The Chelmsford Formation, named after the town of the same name (Figure 1.2), is well exposed in a fold belt, elliptical in plan view, located in the middle of the Sudbury Basin. A type-section has never been defined; the best exposure is a 50 m thick section along the Vermilion River. The formation is approximately 900 m thick and represents the remnant of a unit that was originally thicker and more extensive.

The gradational contact between the Chelmsford Formation and the underlying Onwatin Formation marks the transition between argillite and minor greywacke, to greywacke and minor argillite. The Chelmsford Formation is interpreted as a proximal turbidite sequence and has been described by several investigators (Burrows and Rickaby 1930; Thomson 1957; Williams 1957; Cantin and Walker 1972;

Rousell 1972 ,1984b; Rousell et al. 2002). Beds in the formation range in thickness from 0.03 to 5.2 m with an average thickness of 1.23 m (Rousell 1972). These beds contain three of the possible five "Bouma" divisions (Bouma 1962). The basal division (A) is the thickest (average = 1.12 m), the most common (in 99% of all beds) and comprises dark grey, poorly sorted, commonly graded greywacke. Division C (average thickness = 0.3 m) occurs in only 14% of all beds and consists of buff-coloured, fine- to very-fine-grained sandstone with ripples and small-scale cross-bedding. Division D (average thickness = 0.15 m) occurs in 68% of the beds and comprises very-fine-grained sandstone, siltstone and argillite.

The Chelmsford Formation hosts a variety of sedimentary structures including concretions (Photo 6.1) that are richer in iron and carbonate than the surrounding matrix, clasts, channel fillings, current marks, ripples, convolute laminations, load structures, blunt-nosed flute casts and groove marks (Cantin and Walker 1972; Rousell 1972). Also present are slabs of argillite up to three feet long, as well as rare clasts of gabbro and conglomerate. Possible biogenic markings include patterns of branching arms and fine lines on bedding planes, and narrow vertical pipes that are locally looped (worm burrows?). Paleocurrents flowed to the southwest, parallel to the long axis of the Sudbury Basin (Cantin and Walker 1972; Rousell 1972), perhaps suggesting deposition in an elongate trough. However, the formation may represent the remnant of a more extensive unit which was deposited by paleocurrents that travelled down a southwest-dipping regional paleoslope. This paleoslope may represent the continuation of a regional paleoslope that began during the deposition of sediments of the Huronian Supergroup (2250-2219 Ma, Bennett et al. 1991) since paleocurrents in the Supergroup northwest and southeast of the basin also indicate a southwesterly flow (Long 1978; Rousell and Long 1998).

Photo 6.1. Concretions in the Chelmsford Formation, Highway 144.

Chapter 7

Breccias in the Footwall

J.S. Fedorowich, J.P. Golightly and D.H. Rousell

Introduction

Breccias that occur in the Footwall of the Sudbury Igneous Complex (SIC) include Footwall Breccia (FB) and Sudbury Breccia (SB). These expressions may appear confusing to those unfamiliar with Sudbury geology. However, the two types are distinct in terms of appearance and setting. The South Range Breccia belt (SRBB) or Frood belt, consists mainly of SB but is treated separately because of its size (45 km long) and economic importance, as it hosts large mineral deposits such as the Frood–Stobie Mine.

Footwall Breccia

INTRODUCTION

FB is one of the most puzzling rock types in the Sudbury area (Dressler 1984b). Also known in the past as "late granite breccia" and "leucocratic breccia", it has been described by numerous investigators (Langford 1960; Souch and Podolsky 1969; Greenman 1970; Pattison 1979; Muir 1981, 1983; Lakomy 1990; Dressler et al. 1991; McCormick et al. 2002). FB is an important host for Ni-Cu-PGE mineralization, particularly in the North Range. FB occurs in four settings: discontinuous sheets; "megabreccia"; offset dikes; and as intrusions in felsic norite.

DISTRIBUTION

FB is exposed along the north, east and west margins of the SIC. The bulk of FB occurs as discontinuous lenses and sheets, up to 150 m thick, between the Sublayer, the lowermost unit of the SIC, and the underlying Footwall. FB is thickest in embayment structures, which represent complex depressions at the top of the Footwall. The most extensive and thickest exposures of FB are in embayment structures in the Onaping–Levack area of the North Range, between the Sandcherry Creek Fault and the Fecunis Lake Fault (Figure 7.1; see also Coats and Snajdr (1984, Chart B) for detailed subsurface maps of FB distribution in the area). Note that FB hosts a significant portion of the Ni-Cu-PGE mineralization in the Onaping–Levack area. However, not all embayments or thick sections of FB are mineralized. The contact between FB and Sublayer is generally sharp but is gradational where the Sublayer has incorporated numerous Footwall and FB clasts. A sharp, intrusive contact between a norite apophysis cutting a FB mass indicates a pre-norite age for the breccia at this locality (Dressler 1984b).

Locally in the North Range, dikes of FB, up to 1 m wide, extend into the Footwall as much as 150 m, so that they enclose autochthonous Footwall "fragments" which may be tens of metres across. This zone of injection breccia, gradational between the main FB body and the Footwall, was termed "megabreccia" by Pattison (1979). FB also occurs in North Range Offset dikes such as Whistle and Foy. The contact, with inclusion-bearing quartz diorite in the dikes, is gradational. In the North Range, small plug-like bodies of remobilized FB intrude the felsic norite of the SIC.

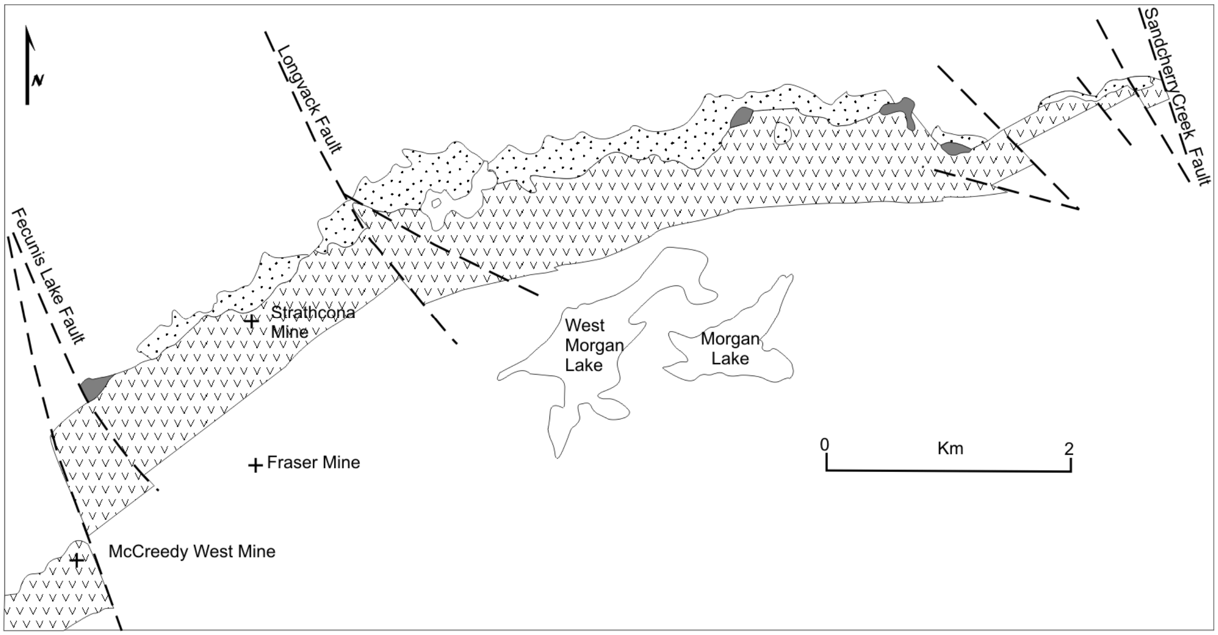


Figure 7.1. Map showing the distribution of Footwall Breccia (dot pattern) in the North Range between the Sandcherry Creek Fault and the Fecunis Lake Fault. Solid pattern is the Contact Sublayer and the V's pattern is the basal unit (mafic norite, felsic norite) of the Main Mass, Sudbury Igneous Complex (*after* Ames, Davidson et al. 2005).

DESCRIPTION

FB (Photo 7.1) is a heterolithic breccia characterized by a variety of angular to subrounded fragments of diverse sizes. Most fragments are local and include gabbro, diabase, granodiorite, mafic gneiss, mafic volcanic rocks and Huronian arenites (Dressler 1984b). The matrix is fine- to medium-grained and is commonly light-coloured, varying from pinkish-white to dark grey. It ranges in texture from sub-igneous near the SIC contact to metamorphic in the Footwall dikes. Pattison (1979) referred to the matrix as "mosaic-granoblastic metamorphic". McCormick et al. (2002) ascribe textural changes in the matrix, as the SIC contact is approached, to contact metamorphism. These changes include: coarsening of quartz and feldspar; feldspar changes from polygons and tablets to igneous laths; and quartz changes from polygons to interstitial oikocrysts. Granophyre veins and dikes, in FB near the SIC contact, suggest local incipient melting of the matrix of FB (Dressler 1984b; Lakomy 1990).

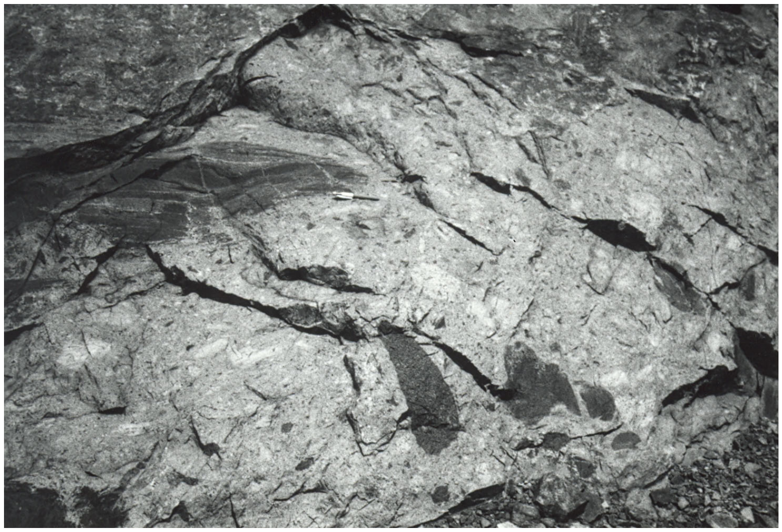


Photo 7.1. Footwall Breccia (Strathcona Mine area); pen for scale.

Plagioclase is the dominant mineral in FB together with quartz, K-feldspar, amphibole, chlorite, biotite, epidote, pyroxene, sericite/muscovite, apatite, ilmenite and magnetite. Colour variations are largely a function of the mafic mineral content.

Ni-Cu-PGE mineralization in FB is of three general types: blebby to disseminated, veinlet, or lenses and veins. The main sulphide minerals are pyrrhotite, chalcopyrite, pentlandite and pyrite. In the Strathcona embayment (plunge 21° south-southwest), Ni-Cu-PGE mineralization is surrounded by chlorine and alkali halos that extend at least 700 m from the axial trace of the embayment (McCormick et al. 2002). The halos are interpreted as the product of the interaction of residual magmatic fluids with the host rocks. The fluids alter the host-rock mineralogy and precipitate chlorine-bearing minerals such as apatite, phlogopite, amphibole, scapolite and ferropyrosmalite.

ORIGIN

Clasts of SB in FB and thermal overprinting of FB by the SIC suggest the following (overlapping?) sequence of formation: SB, FB and SIC. Lakomy (1990) interpreted the origin of the breccias, in terms of the meteorite impact model, as follows. SB formed in rocks surrounding the transient crater by frictional melting due to movement between large blocks, whereas FB represents a part of the uplifted crater floor directly beneath the impact melt sheet (SIC). FB occurs mainly in the North and East ranges where the exposed crustal level may be as much as 5 km higher than that in the South Range (Naldrett et al. 1970). This intimates that FB formed along the upper part of the crater wall (Dressler 1984b). According to Dressler et al. (1991), FB represents a parautochthonous mass of crushed and shock-metamorphosed rocks originally similar to SB (i.e., pseudotachylite). The breccia formed along the crater walls prior to the emplacement of the Onaping Formation and the SIC. Before the intrusion of the SIC, FB was in contact with the breccias at the base of the Onaping Formation. In conclusion, SB, in the footwall rocks of the North Range, consists of pseudotachylite that was interpreted to have formed by frictional melting along major faults (Spray and Thompson 1995). However, it seems unlikely that frictional melting could produce a FB body 150 m thick.

Sudbury Breccia

GENERAL FEATURES

SB is widely distributed in the footwall rocks that surround the SIC and is of economic importance because it hosts several Ni-Cu-PGE deposits. In general, the breccia consists of subrounded fragments set in a fine-grained to aphanitic matrix which may be fragmental, pseudotachylitic or recrystallized. SB has been described by numerous investigators over the years including Fairbairn and Robson 1942; Yates 1948; Zurbrigg 1957; Speers 1957; Card 1978b; Dressler et al. 1991; Müller-Mohr 1992a, 1992b; Thompson and Spray 1994; Thompson 1996; and Rousell et al. 2003.

SB occurs as dikes (Photo 7.2) or irregular-shaped bodies that range in size from millimetre-scale veins to that of the South Range Breccia Belt (see below). Contacts between the breccia and the country rock are usually sharp. Details of fragment properties such as orientation, aspect ratios, roundness and fractal dimensions are given in Rousell et al. (2003). SB bodies tend to follow zones of structural weakness such as lithologic boundaries, faults, joints and foliations. Dressler (1984b) found no obvious relationship between the orientation of SB dikes and the geometry of the outer surface of the SIC. However, in the Strathcona Mine (North Range), SB bodies are apparently oriented subparallel to the outer margin of the SIC (Fedorowich et al. 1999).

Figure 7.2 shows the regional distribution of SB. The most obvious feature of the distribution pattern is a zone of relatively intense brecciation adjacent to the outer margin of the SIC. The zone is approximately 15 km wide in the Footwall of the South Range and from 5 to 10 km wide in the Footwall of the East and North ranges. Fedorowich et al. (1999) attempted to quantify the distribution of SB in the vicinity of the Strathcona Mine (Figure 7.1). In a 7 km^2 area, 4.7% of the total outcrop area (22.8%) consists of SB, with local concentrations in subareas varying from 2 to 8%. Based largely on exposures along Highway 144 in the North Range Footwall, Thompson and Spray (1994), Spray and Thompson (1995) and Thompson (1996) proposed four zones of enhanced SB development which begin at 0, 25, 38 and 78 km northwest of the outer margin of the SIC and are 10, 5, 3.5 and 2 km wide, respectively. However, other investigators suggest that more field work is required in order to demonstrate that the exposures are, in fact, laterally continuous (Rousell et al. 2003).

The Ni-Cu-PGE deposits at Sudbury may be classified into four types based on the structural setting (Naldrett 1984b; Morrison et al. 1994; Golightly and Rousell 2000): 1) SIC–Footwall contact; 2) Footwall; 3) offset dike; and 4) sheared. SB is the main host in the Footwall deposits, which include McCreedy West and McCreedy East (North Range), Victor (East Range) and Lindsley (South Range). Quartz diorite is the principal host for the ore in the offset dikes but SB also locally hosts mineralization. Examples from radial offset dikes include the Milnet Mine in the Whistle offset (Grant and Bite 1984) and several deposits in the Copper Cliff offset (Cochrane 1984). The McConnel dike near Kirkwood Mine (Grant and Bite 1984) and the Food–Stobie deposits (Grant and Bite 1984) are examples from concentric offsets. (See Figure 1.2 for the location of deposits.)

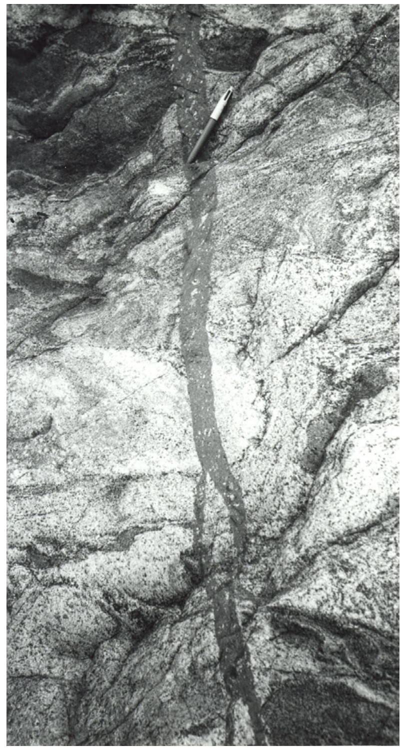


Photo 7.2. Pseudotachylite Sudbury Breccia dike with sharp, straight contacts (Hardy Mine area).

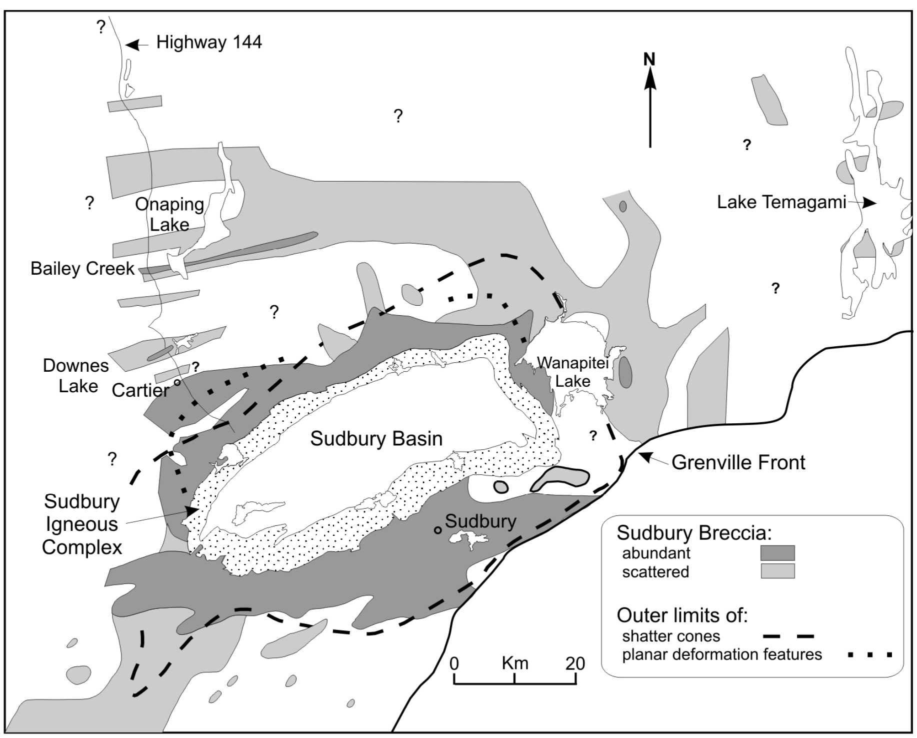


Figure 7.2. Map indicating the regional distribution of Sudbury Breccia and the extent of shatter cones (Guy-Bray 1966) and planar deformation features (Dence 1972). ? = lack of data. (*After* Rousell et al. 2003.)

CLASSIFICATION

Several classifications of SB, based on a variety of criteria, have been proposed (e.g., Fairbairn and Robson 1942; Mitchell and Mutch 1957; Speers 1957; Card 1978b; Müller-Mohr 1992a, 1992b; Dressler 1984b; Peredery and Morrison 1984; Thompson and Spray 1994). A simple three-fold classification, based on the nature of the matrix, is as follows (Rousell et al. 2003): 1) clastic; 2) pseudotachylite; and 3) microcrystalline.

Clastic SB consists of rounded fragments set in a clastic matrix. The matrix is characterized by delicate flow surfaces that are visible because of slight differences in composition between the layers. The flow surfaces tend to be parallel to the dike walls and mould themselves around clasts (Photo 7.3). Some flow-surface patterns resemble steeply plunging folds and others sheath folds (Photo 7.4). Clastic SB is the dominant type in rocks of the Huronian Supergroup and in Nipissing gabbro of the Southern Province. Clastic SB occurs in the Huronian rocks at Lake Temagami, located 80 km east of the base of the SIC and the most distal locality of SB reported to date. There are rare occurrences of flow surfaces in pseudotachylite SB in the Superior Province and local flow surfaces are present in SB in the Creighton and Murray plutons. Tectonometamorphism has affected clastic SB, to a varying degree, as follows: fragments are locally tectonically elongated (e.g., Frood belt); crystalloblastic micaceous minerals define a schistosity; and late chlorite porphyroblasts cut across the foliation. Dressler et al. (1991) report the presence of garnet and staurolite in the Frood belt.

Photo 7.3. Clastic Sudbury Breccia showing flow surfaces that are parallel to dike walls and which mould around clasts (Kelly Lake). Coin for scale.

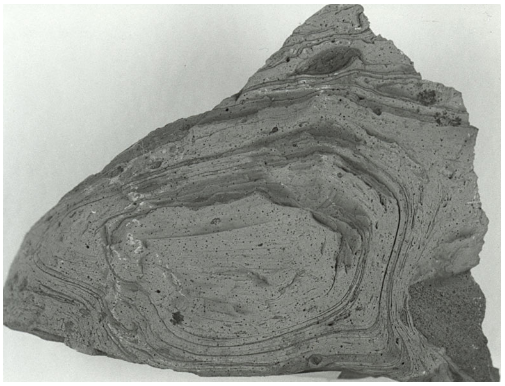


Photo 7.4. Clastic Sudbury Breccia with surface disposed as a sheath fold (Lake Temagami); width of specimen is 13 cm.

Pseudotachylite SB is characterized by a black, aphanitic matrix and occurs in the granitic and gneissic rocks of the Superior Province northwest of the SIC (Photo 7.2) and in the granitic rocks of the Creighton and Murray plutons in the footwall of the South Range. Pseudotachylite SB resembles pseudotachylite from the Vredefort Structure, South Africa, which is the type-area for this rock (Shand 1916; Killick and Reimold 1990; Reimold 1995). Pseudotachylite SB bodies vary in size from millimetre-scale veins to bodies tens of metres across. The dark colour of the matrix appears to be due, at least in part, to finely disseminated opaque material. Shades of grey and red occur locally. Clasts are mainly subrounded and are generally derived from the host rocks. Clasts with a foliation normal to that in the adjacent host rock, indicate rotation of clasts by a flowing matrix. Based on back-scattered electron images on a scanning electron microscope, Thompson and Spray (1994) interpret the texture of the matrix as cryptocrystalline micro-igneous with a grain size about 10 µm.

In the North Range, footwall rocks adjacent to the SIC have undergone contact metamorphism in a zone as much as 1.2 km wide (Dressler 1984b). In the South Range, an overprint of regional metamorphism has obliterated the effects of contact metamorphism. Rousell et al. (2003) describe the effects of contact metamorphism on the SB matrix and classify the breccia as “microcrystalline SB”. In an inner zone, 60 to 140 m wide, twinned plagioclase grains (75 modal %) have an average diameter of 0.1 mm with an upper size of 0.5 mm (Photo 7.5). Clinopyroxene grains (20 modal %) are as much as 0.3 mm in diameter with an average diameter of 0.05 mm. Biotite, up to 1 mm in diameter, is a minor mineral. The grain size of the plagioclase gradually decreases outward from the SIC base. In hand specimen, the matrix of microcrystalline SB and pseudotachylite SB look similar.

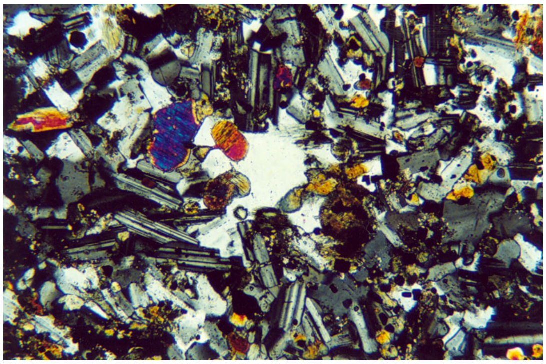


Photo 7.5. Photomicrograph (crossed polars) of the matrix of microcrystalline Sudbury Breccia from the Levack Mine area. Minerals include plagioclase crystals, approximately 0.2 cm long, together with pyroxene and quartz. Width of photo is 1.2 mm. (Thin section C821404 courtesy of Inco Ltd.)

ORIGIN

The formation of clastic SB may be considered in terms of two processes: rock fragmentation followed by material flowage. Presumably, the shock wave associated with meteorite impact suddenly deposited a considerable amount of heat in the compressed rocks. Possibly the heat caused the vaporization of pore water in the relatively wet supracrustal rocks of the Southern Province. Vaporization, in turn, significantly raised the pore pressure, resulting in an explosive dilation which shattered the rocks (Melosh 1989). The flow surfaces, together with the rounded fragments, intimates fluidization either by gas flow (Reynolds 1954), acoustic vibration (Melosh 1989) or by some combination of both. The flow surfaces in clastic SB are interpreted to have formed by lamellar convection flow (Rousell et al. 2003). However, lamellar flow was probably preceded by turbulent flow. Clastic SB apparently developed within extension fractures. Rousell et al. (2003, p.167-168) describe the process as follows:

> "The walls of some bodies of clastic SB are locally penetrated by fractures which are filled by SB matrix. New fragments are in the process of being detached from the walls. Some large clasts, cut by matrix-filled fractures, are in the process of splitting. Long, matrix-filled fractures, branching off from the main bodies, have straight walls suggesting that the breccia material was not actually derived from them. Instead, breccia matrix was forcefully injected into the offshoots from "breccia producing" or "feeder" bodies. Truncation of the flow-surfaces at the branch lines indicates that the offshoot dikes are younger than the feeder bodies. Apparently, fracture propagation, wall and clast fragmentation, flowage and clast-rounding were on-going processes of unknown duration".

According to some investigators (Thompson and Spray 1994, 1996; Butler 1994; Thompson 1996) the matrix of pseudotachylite SB developed as a result of high-speed slip along faults oriented concentric to the outer margin of the North Range SIC. The faulting occurred during the collapse phase of the cratering process. Field work and an examination of radar images have led others to question the existence of the supposed faults (Rousell et al. 1997; Lowman 1997; Rousell and Long 1998; Cowan et al. 1999; Long 2004). Pseudotachylite SB probably formed by a combination of commutation and melting with the latter possibly the result of microcrystalline slip, along numerous fractures, in the relatively dry rocks of the Superior Province.

South Range Breccia Belt

INTRODUCTION

The South Range Breccia Belt (SRBB) is a concentric offset dike that occurs in the Footwall of the South Range. The SRBB represents a long, narrow, arcuate belt that extends about 45 km, from Kirkwood Mine in the east to Vermilion Mine in the west (Figure 7.3). The belt varies from roughly 10 to 500 m in width and dips steeply to the southeast. It lies within rocks of the Huronian Supergroup and is subconcordant to the strike of lithologic units of the group. The SRBB is cut by the Copper Cliff Offset dike with the Kirkwood and Vermilion Offset dikes at either end of the belt (Scott and Spray 1999, 2000).

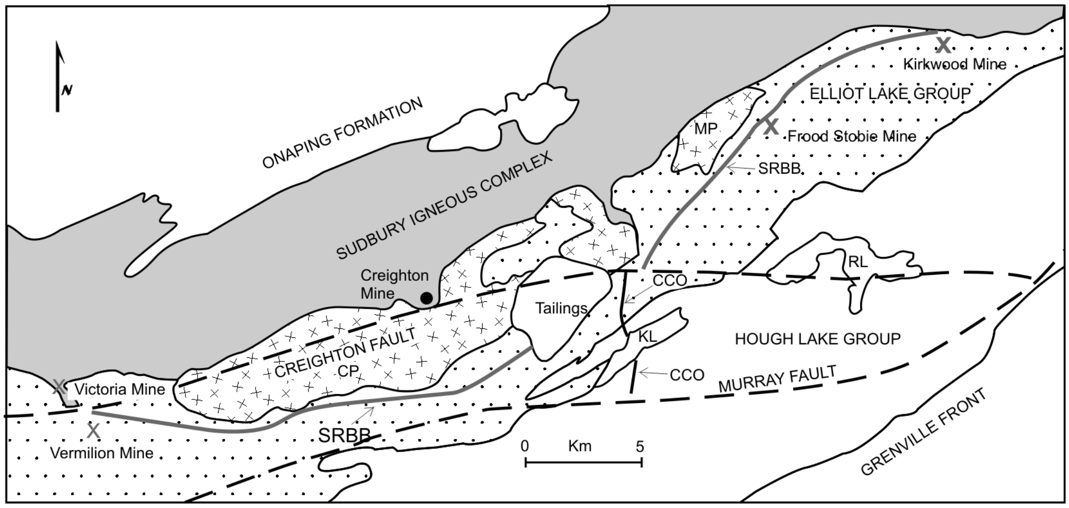


Figure 7.3. Map showing the setting of the South Range Breccia Belt (SRBB) (*after* Scott and Spray 2000). CCO = Copper Cliff Offset, CP = Creighton Pluton, MP = Murray Pluton, KL =Kelly Lake, RL = Ramsey Lake.

The SRBB contains relatively little quartz diorite when compared to typical offset dikes. Nevertheless, there are several quartz diorite bodies in the belt that range in size from 1 to 100 m in length. The quartz diorite body that hosts the Frood–Stobie Mine, the largest Ni-Cu-PGE deposit in the Sudbury area, is approximately 2.9 km in length (Zurbrigg 1957).

The belt consists largely of SB. The breccia matrix is very-fine-grained, consists mainly of quartz, biotite and ilmenite and, locally, exhibits a well-developed flow structure, interpreted as pseudotachylite by Scott and Spray (1999, 2000). Flow structures are a characteristic feature in SB bodies within supracrustal rocks of the South Range Footwall. Rousell et al. (2003) classify these bodies as clastic Sudbury Breccia. The SRBB is the most extensive body of SB in the area and further differs from other SB bodies as it hosts melt bodies and contains a more polymictic-fragment lithology (Table 7.1).

Table 7.1. Fragment types and matrix statistics from the South Range Breccia Belt. Data are from 16 measured cross-sections between the Frood–Stobie Mine and the Copper Cliff Offset (Müller-Mohr 1992b).

Fragment Type	No. of intervals	Total length of intervals (cm)	Percentage length of grand total lengths	Average interval length (cm)
Matrix	486	9884	27.24	20
Metasediment	243	12 340	34.02	51
Basalt	137	8222	22.66	60
Amphibole	53	2663	7.34	50
Unidentified	18	1223	3.37	68
Metasediment / Basalt	7	440	1.21	63
Granite	7	819	2.26	117
Metavolcanic	6	315	0.87	53
Quartz clast	3	45	0.12	15
Arkose	2	37	0.10	19
Dacite	2	55	0.15	28
Other sediments	3	235	0.64	78
Grand totals	967	36 278	100%	

ORIGIN

Early workers offered a novel explanation as to the origin of the quartz diorite and Frood–Stobie deposits but were rather general as to the formation of the SRBB as a whole (Zurbrigg 1957; Hawley 1965). In brief, ore-forming fluids, consisting of quartz diorite and sulphide liquids which were segregated at the base of the formerly overlying SIC, together with exotic mafic fragments (Photo 7.6), were injected downward into a steeply dipping extensional fracture zone (Figure 7.4). This gave rise to upside-down zoning. Later erosion removed the overlying SIC and detached the SRBB from the SIC.

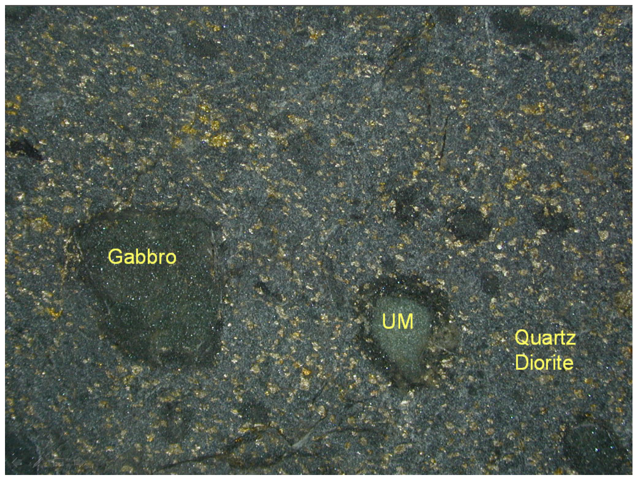


Photo 7.6. Gabbro and ultramafic (UM) inclusions, as well as disseminated sulphide minerals, in quartz diorite from the Stobie Mine. The gabbro inclusion is 20 cm across.

Later investigators interpreted the origin of the SRBB in terms of the astrobleme theory. Scott and Spray (1999) applied the terrace-collapse model (Figure 7.5) of Spray and Thompson (1995) for the formation of the SRBB and the emplacement of the Frood–Stobie deposit. Initially, the inner wall of the crater collapsed along a listric normal fault with the initial generation of pseudotachylite melt (Figure 7.5a). A high-speed displacement of 2.5 km is required to generate a pseudotachylite width of 250 m. Later, the hanging wall overrode pooled sulphides which were back-injected into the pseudotachylites (Figure 7.5b). Finally, compression caused the friction-generated melt to flow over the top of the hanging wall while pseudotachylite dikes were injected into extension fractures (Figure 7.5c). Note that in both of the above models the SRBB was interpreted to have formed in a zone of crustal extension. Later deformation rotated the South Range so as to dip steeply northwest with local overturning to the southeast. Presumably, the SRBB would be rotated in a similar manner resulting in the "superfault" (Figure 7.5c) having a steep dip to the southeast.

ECONOMIC IMPORTANCE

While the exact significance of the SRBB remains debatable, its economic importance as the host of the Frood–Stobie Ni-Cu-PGE deposit is unequalled in the Sudbury area and even world-wide. The only other mine of historical interest is the small, but PGE-rich, Vermilion Mine located near the western end of the belt. The Victoria and Kirkwood mines may be related to the western and eastern terminus, respectively, of the belt. FNX Mining Company recently (2005) discovered the Segway mineral occurrence near the Kirkwood Mine.

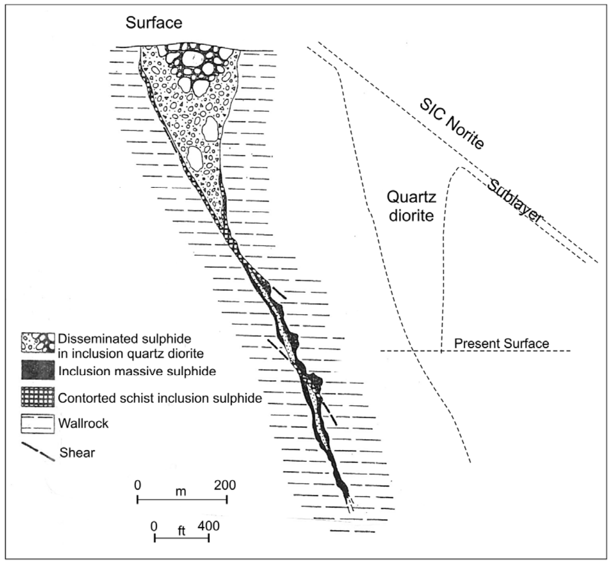


Figure 7.4. Generalized section through the Frood ore body looking southwest (*from* Souch and Podolsky 1969). The sketch on the right shows the possible up-dip extension of the orebody to reach a now-eroded extension of the Sudbury Igneous Complex. See text for explanation.

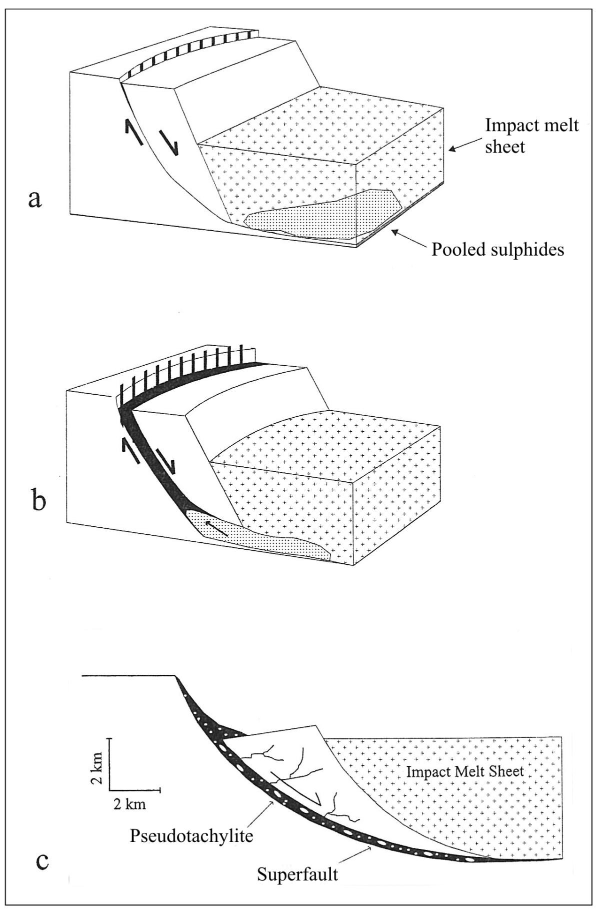


Figure 7.5. Formation of the South Range Breccia Belt by crater collapse along a listric normal fault. a) initial failure; b) generation of pseudotachylite and overriding pooled sulphides; c) frictional melt flows over top of hanging wall block and injection of pseudotachylite dikes. (*After* Scott and Spray 1999.)

Chapter 8

Sudbury Igneous Complex

E.F. Pattison

Introduction

The Sudbury Igneous Complex (SIC) formed during the creation of the Sudbury Structure at 1850 Ma (Krogh et al. 1984). The SIC comprises three major components, namely Main Mass, Contact Sublayer and Offset Sublayer (Figure 8.1). The rocks of the Main Mass are differentiated into three units which are, from bottom to top, quartz-bearing norites, quartz gabbro (formerly called transition zone) and granophyre. The Contact Sublayer comprises a group of relatively small gabbronorite bodies. They are localized along the basal contact of the SIC, between Main Mass norite and the various units that comprise the footwall of the SIC. The Offset Sublayer consists of a number of dikes that either radiate outward from the base of the SIC (Radial Offsets) or are sub-parallel to the basal contact of the SIC (Concentric Offsets). The Contact and Offset sublayer units are of major economic importance since they host much of the Ni-Cu-PGE mineralization associated with the SIC. Figure 8.2 shows a typical section through the SIC as well as the underlying and overlying rocks.

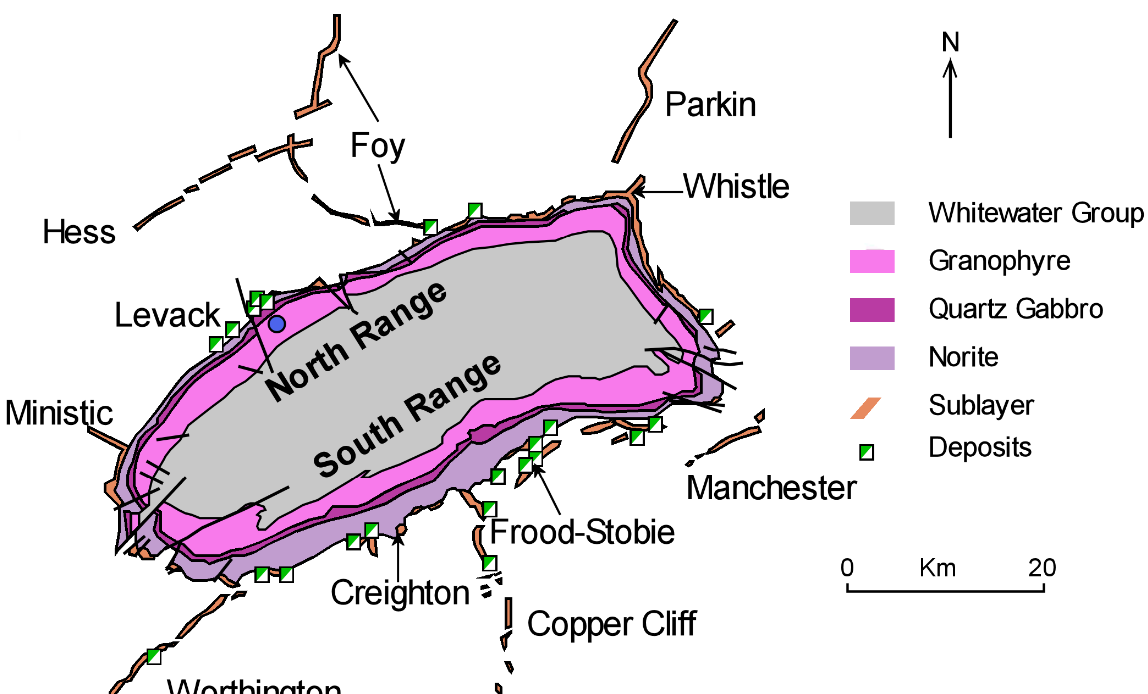


Figure 8.1. Geological sketch of the Sudbury Igneous Complex showing the distribution of major units of the Main Mass (Granophyre, Quartz Gabbro and Norite) and the distribution of the Contact and Offset sublayers.

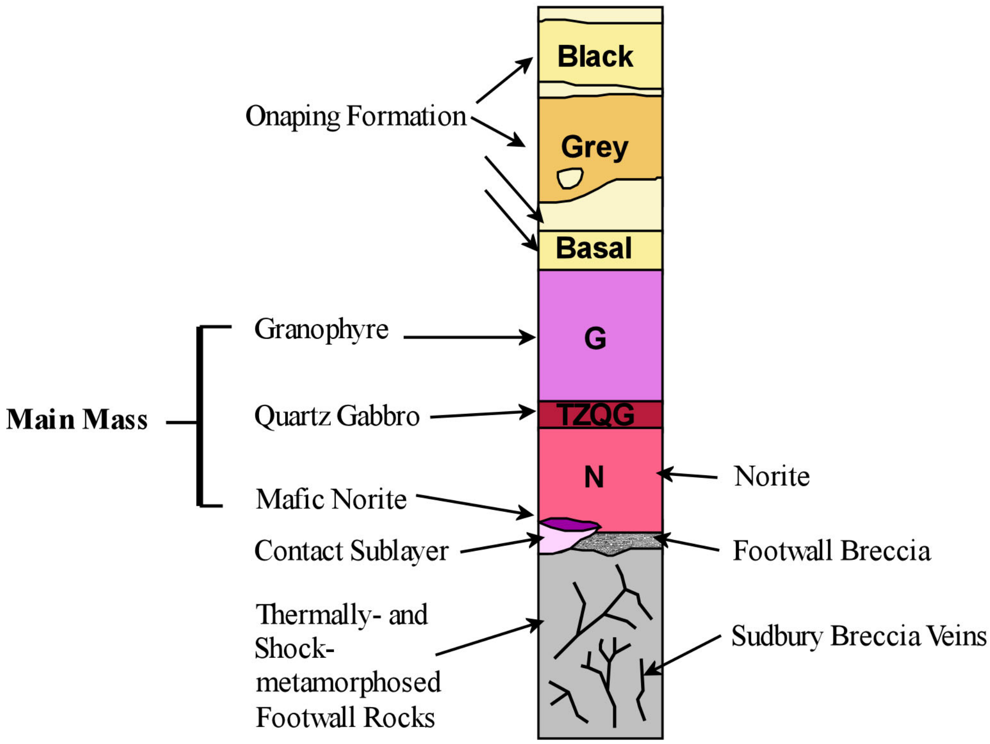


Figure 8.2. Typical stratigraphic section through the Sudbury Igneous Complex (*after* Grieve 1994).

To these traditional subdivisions of the SIC might be added the various small intrusions and glass components of the Onaping Formation (see Chapter 6) and, possibly, the glassy and pseudotachylite matrices of some of the Sudbury Breccia bodies which surround the SIC as components of the Sudbury Structure (see Chapter 7). The rationale for including these rocks as part of the SIC lies in their coeval formation at 1850 Ma.

Location, Size and Form

The SIC was emplaced along the unconformable contact between the Archean Superior Province and the Paleoproterozoic Southern Province of the Canadian Shield. The Sudbury Structure has a possible diameter of as much as 200 km as defined by the occurrence of Sudbury Breccia. However, the outer limits are ill-defined and debatable. The SIC occupies a relatively small area in the core of the structure, transects the contact between the Superior Province and Southern Province and intrudes rocks of both provinces. The SIC has traditionally been geographically divided into the North and South ranges, with the eastern part of the North Range called the East Range. Footwall rocks of the North and East ranges consist mainly of Archean gneisses and felsic intrusions, whereas those of the South Range comprise Paleoproterozoic mafic volcanic and sedimentary rocks and felsic intrusions. Granophyre, the upper unit of the Main Mass, intrudes the Onaping Formation, which represents the basal unit of the Whitewater Group. Southeast-dipping reverse faults at the southwest corner of the SIC approximate the contact between the North and South ranges, whereas southwest-dipping reverse faults at the southeast corner of the SIC approximate the contact between the East and South ranges.

The Main Mass and Contact Sublayer of the preserved portion of the SIC, after 1850 Ma of erosion and deformation, outcrop as a northeast- to east-northeast-trending elliptical ring with long and short axes of approximately 60 and 27 km, respectively (Figure 8.1). Maximum outcrop widths of the SIC are approximately 4 km in the North Range and 7 km in the South Range, corresponding to a true thickness of about 2 km and 5 km, respectively. The thickness figure for the South Range is minimal as the south-dipping reverse faults caused tectonic shortening.

Over the years the SIC has been described as a sill (Coleman 1905), a ring dike (Phemister 1926), a laccolith or funnel-shaped intrusion (Wilson 1956; Naldrett et al. 1970), or an impact-generated melt (e.g., Dickin et al. 1999). The present three-dimensional form of the SIC is well constrained to a depth of 2 to 3 km by data accumulated from over one hundred years of mining and exploration. These data suggest that the basal contact of the SIC generally dips inward; about 35°S in the North Range and 55°N in unfaulted parts of the South Range. However, major portions of the South Range have been tectonically tilted to the vertical with local overturning. Much of the East Range has also been steepened, such that the average dip is approximately 70°W. The basin-like configuration of the SIC is supported by seismic reflection profiles obtained by means of a major Lithoprobe program (Milkereit et al. 1992). Figure 8.3 displays two possible interpretations of the seismic data. The interpretation of Wu et al. (1994) suggests that the North Range SIC dips continuously to the southeast to a position below the South Range. On the other hand, the interpretation of Card and Jackson (1995) represents what might be termed a more "traditional" view, whereby the SIC closes at a relatively shallow depth in a synclinal fashion, reminiscent of the folded-sill concept.

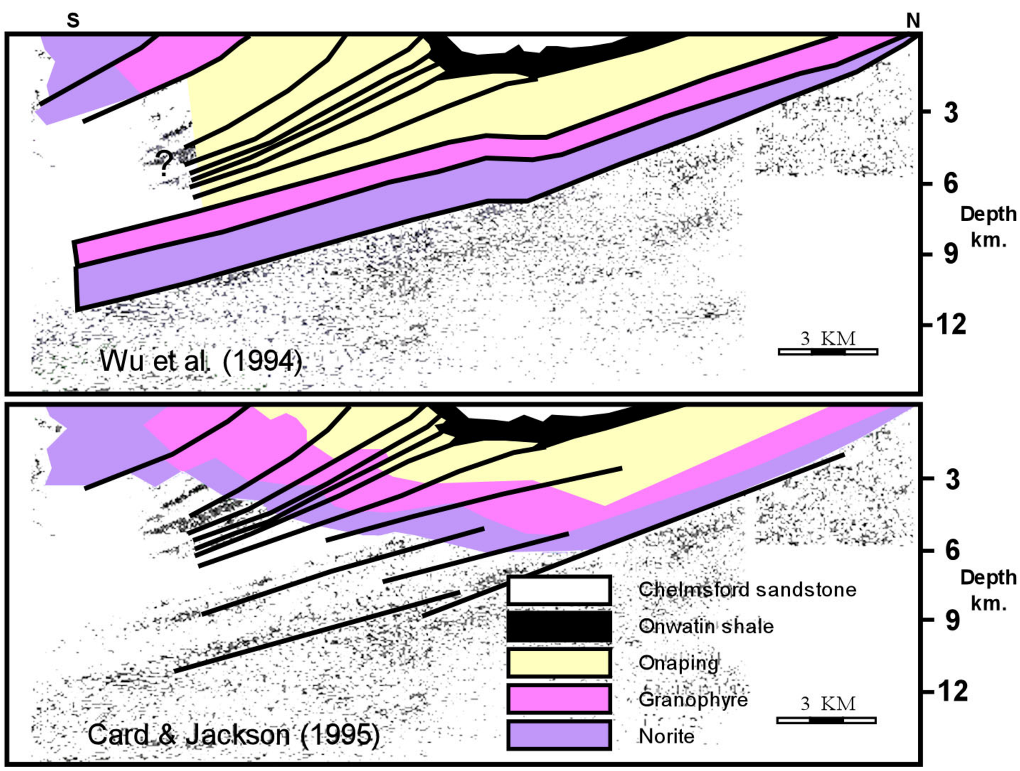


Figure 8.3. Two disparate interpretations of the cross-sectional form of the SIC based on the results of the 1991-92 Lithoprobe seismic transect through the Sudbury Structure. Heavy lines are possible faults.

Offset dikes extend for as much as 30 km into the footwall and are usually less than 100 m wide. The Foy offset (Figure 8.1) is an exception as it is greater than 200 m wide near the junction with the Main Mass. The dikes thin with increasing distance from the Main Mass and tend to be only a few metres wide at their distal extremities.

Origin

The SIC has long been recognized as unusual, if it were derived by fractionation, because of the silica-rich nature of the bulk composition and the excessive amount of granophyre. Several models for the origin of the SIC have been proposed over the past 100 years. See Giblin (1984) and Naldrett (2004) for historic reviews.

Most early concepts invoked the intrusion of an endogenic, presumably mantle-derived magma or magmas, along an unconformity between the overlying Whitewater Group and the underlying Archean or Paleoproterozoic footwall rocks. The magma was emplaced either as a single pulse, which then differentiated into mafic and felsic layers, or as separate pulses of mafic and felsic magma. The high silica content was attributed to contamination by assimilation of siliceous country rocks, while the high proportion of granophyre formed one of the main arguments for a two-magma origin of the SIC. The model of Naldrett et al. (1970), reminiscent of the structure of the Skaergaard intrusion, appeared to resolve the one- versus two-magma dilemma. The SIC was interpreted as a funnel-shaped body in which an upper, saucer-shaped cap of granophyre overlies mafic layers located deeper in the funnel.

A revolution in geologic thought was initiated by Dietz (1964) with the original suggestion that the Sudbury Structure represented the roots of an ancient meteorite impact crater or “astrobleme”. Dietz (1972) visualized a major impact, excavation of a crater, deposition of the Onaping Formation as a fall-back breccia or “suevite” and intrusion of the SIC as a impact-generated, but probably mantle-derived magma. This was the first proposal for an exogenic origin for the Sudbury Structure; many investigators still consider the model to be remarkably valid.

The impact model was further supported by new radiogenic isotope studies of the SIC and surrounding rocks. Figure 8.4 represents a neodymium-strontium isotope-evolution diagram with data from various components of the Main Mass and Sublayer. Also shown is a mixing line between Bulk Earth and upper crust indicating the calculated percentage contamination in the various silica phases, and a line between data for upper crust and Levack granulite (commonly known as the Levack Gneiss Complex). The best fit, through the line connecting Levack granulite and upper crust, supports the concept of massive crustal assimilation during the formation of the SIC (Naldrett et al. 1986).

More recent studies (Golightly 1994; Grieve 1994) suggest that the SIC might be entirely impact melt with no contribution from the mantle. According to Grieve (1994), an impact crater with the apparent size of the Sudbury Structure might produce approximately 35 000 cubic km of melt, which is more than enough to account for the volume of the SIC and any other associated melt bodies such as those in the Onaping Formation.

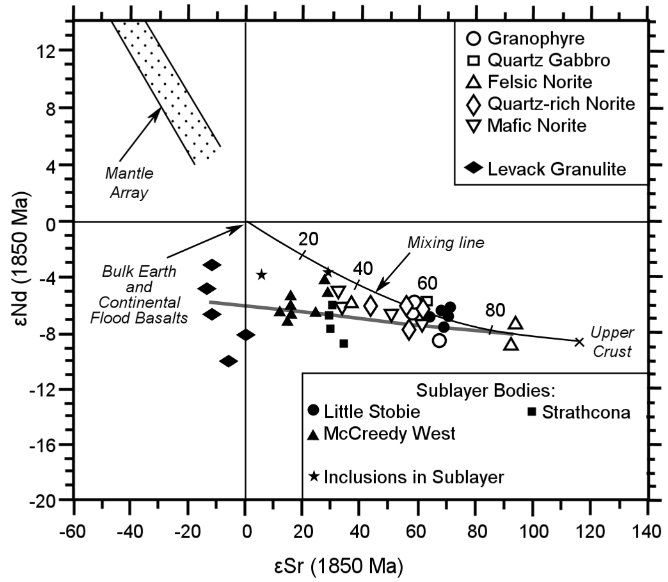


Figure 8.4. Neodymium-strontium isotopic evolution diagram showing data for the various components of the Main Mass and Sublayer of the SIC (*after* Naldrett et al. 1986). Note the mixing line between Bulk Earth and upper crust showing calculated percentage contamination in the various SIC phases and a line between data for Levack Gneiss and upper crust. Both lines are well away from the mantle array and support the concept of massive crustal assimilation during the formation of the SIC.

Main Mass

The SIC may be described as a layered complex but this is true only in a general sense. Although the SIC represents a differentiated igneous body with concentric rings of norite, gabbro and granophyre, it lacks many features of true layered complexes. These features include graded bedding, modal layering, cross-bedding and scour-and-fill structures (Naldrett and Hewins 1984). The only visual evidence of convection processes in the SIC is an igneous lamination, in South Range norite, due to a parallel alignment of lathy plagioclase crystals.

The changes that occur within the SIC are subtle and cannot be determined by visual inspection alone. However, the changes are revealed by detailed lithochemical, petrographic and mineral chemistry studies along transects through the SIC.

LITHOCHEMISTRY

Collins (1934) was one of the first to demonstrate the chemical changes that occur across the SIC. Figure 8.5 illustrates the variations in SiO_2, MgO, TiO_2 and P_2O_5 in a traverse across the SIC in the North Range. Variations across the South Range are similar. The granophyre and the units below the quartz gabbro exhibit gradual variations in the content of the oxides, whereas the intervening quartz gabbro shows abrupt changes. Of particular interest are the sharp peaks for TiO_2 and P_2O_5 due to the appearance of titaniferous magnetite and apatite as cumulus phases. These minerals provide important clues regarding crystallization processes within the SIC at this stage in its evolution. Most investigators agree with Collins (1934) that the data support a process of progressive differentiation within a single crystallizing magma. Others, including Knight (1917, 1923) and Phemister (1926) argued for the intrusion of two separate magmas, one basic and the other felsic, separated by a hybrid mixing zone.

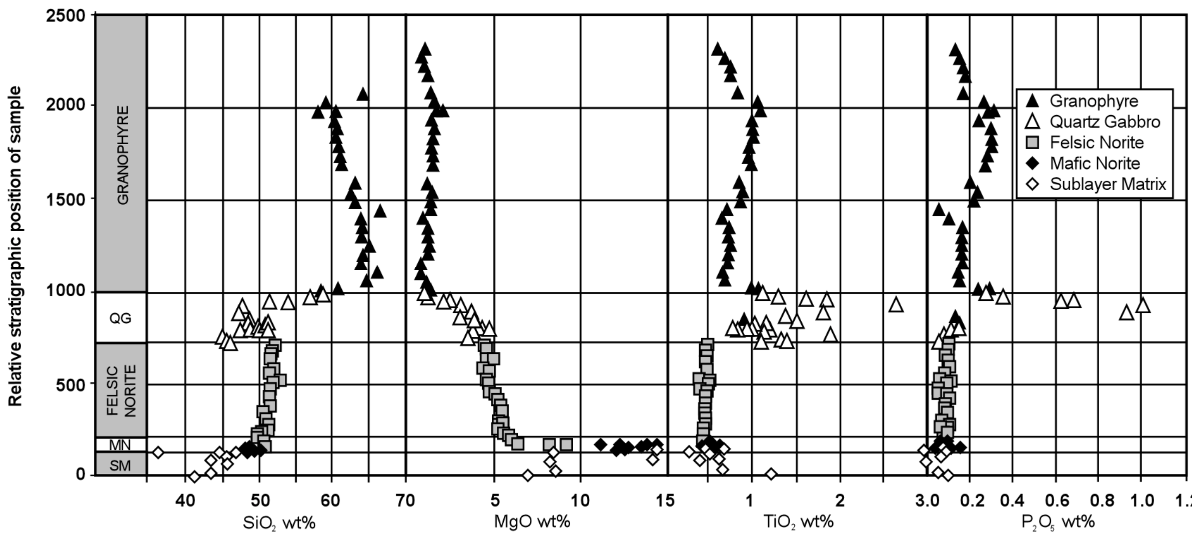


Figure 8.5. A typical lithochemical section through the SIC showing variations in SiO_2, MgO, TiO_2 and P_2O_5 (*after* Lightfoot, Doherty et al. 1997; Lightfoot et al. 1997a, 1997b). Note the relatively smooth variations through the norite and granophyre units in contrast to the rather abrupt variations through the quartz gabbro unit. QG = Quartz Gabbro, MN = Mafic Norite, SM = Sublayer Matrix.

PETROGRAPHY

Figure 8.6 illustrates the general petrography of the South and North ranges and is based on detailed modal analysis of closely spaced samples (Naldrett and Hewins 1984). Correlations of units between the two ranges are also indicated.

Subdivisions of the South Range, from bottom to top, are as follows.

1) Quartz-rich norite: is a fine-grained rock with as much as 15 modal % quartz that may have an opalescent blue colour.

2) South Range norite: is characteristically medium-grained, with a quartz content of 5%, except near the top where the content increases abruptly.

3) Quartz gabbro: has abundant augite, no hypersthene, with over 60% quartz at the top. The unit is locally enriched in magnetite and apatite.

4) Granophyre (formerly called micropegmatite): consists of approximately one part tabular to prismatic plagioclase and three parts a granophyric intergrowth of quartz, plagioclase and K-feldspar. Near the upper contact a plagioclase-rich phase resembles the quartz gabbro (Peredery and Naldrett 1975).

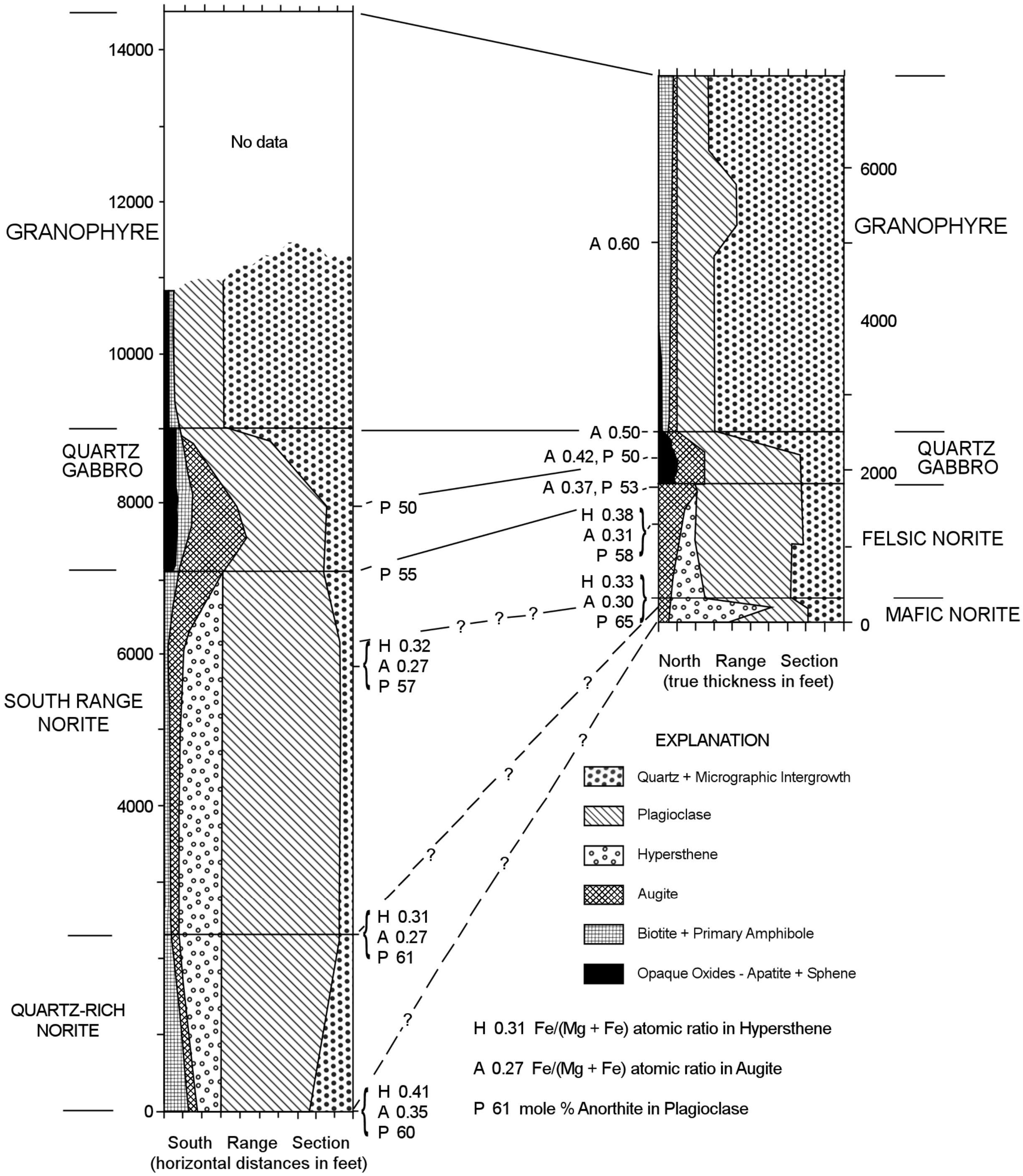


Figure 8.6. Diagram showing modal variations through the Main Mass of the Sudbury Igneous Complex on the South Range (left section) and the North Range (right section) (*from* Naldrett and Hewins 1984). Note that horizontal distances (left section) and true thicknesses (right section) are in feet. Compare these modal data with the lithochemical data set out in Figure 8.5. Also shown are some mineral chemical data for plagioclase (P), augite (A) and orthopyroxene (H). More detailed mineral chemical data are presented in Figure 8.7.

Subdivisions of the North Range, from bottom to top, are as follows.

1) Mafic norite: is unique and is characterized by abundant cumulus hypersthene which is poikilitically enclosed by large plates of plagioclase. The unit occurs sporadically and typically overlays embayment structures.

2) Felsic norite: a medium-grained rock, it is the main norite unit in the North Range.

3) Quartz gabbro: is similar to that on the South Range.

4) Granophyre: is also similar to that on the South Range.

The Main Mass can also be subdivided into units which are based on the appearance / disappearance of certain cumulus minerals. The following assemblages are present, from the base up.

1) Orthopyroxene only: This assemblage is limited to mafic norite of the North Range where cumulus hypersthene is enclosed by intercumulus plagioclase.

2) Orthopyroxene-plagioclase cumulates: occur in the lower half of South Range norite and felsic norite. Augite occurs as an intercumulus phase.

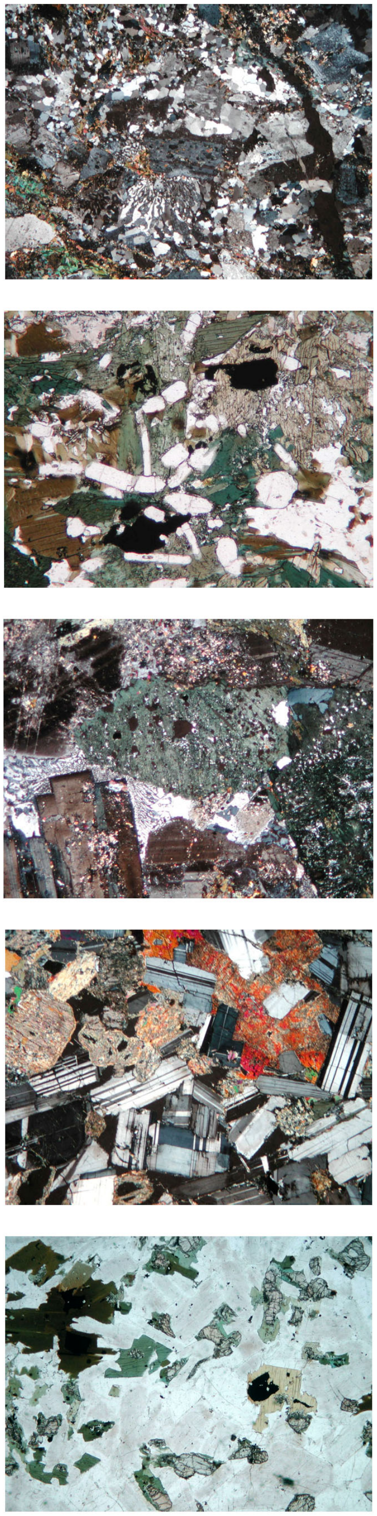
3) Orthopyroxene-plagioclase-clinopyroxene cumulative assemblage: occurs in the upper half of South Range norite and felsic norite. The upper contact of the norite units is defined by the last appearance of cumulus hypersthene. This contact is not readily observable in the field as the amount of hypersthene at the top of the norite units is small.

4) Plagioclase-clinopyroxene cumulates: define the quartz gabbro member.

5) Plagioclase-clinopyroxene-magnetite-apatite cumulates: define the oxide-apatite member of the quartz gabbro unit.

6) Plagioclase only cumulus mineral: characterizes the granophyre unit.

The sequence of cumulus minerals is compatible with crystallization from a single differentiating magma and conforms with the lithochemical variations presented above. Abrupt changes in the chemistry of the rocks correlate with the entry and disappearance of cumulus minerals. Photomicrographs of typical examples of the major units from the South and North ranges are shown in Photos 8.1 and 8.2, respectively.



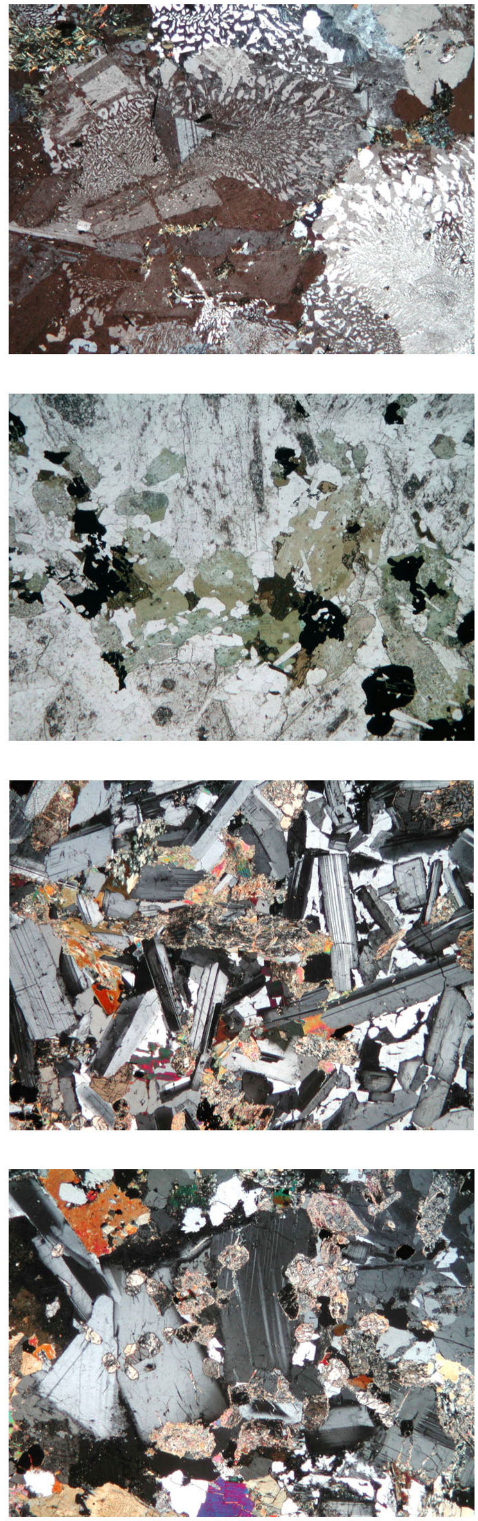


Photo 8.1. Photomicrographs of various units of the South Range SIC in stratigraphic order. From top to bottom these are: granophyre, oxide-apatite rich quartz gabbro, quartz gabbro, South Range norite, and quartz-rich norite. Each image is approximately 4 mm wide.

Photo 8.2. Photomicrographs of various units of the North Range SIC in stratigraphic order. From top to bottom these are: granophyre, oxide-apatite rich quartz gabbro, felsic norite, and mafic norite. Each image is approximately 4 mm wide.

MINERAL CHEMISTRY

Figure 8.6 shows compositional data for plagioclase, augite and hypersthene (Naldrett and Hewins 1984) while Figure 8.7 indicates variations in the composition of augite and hypersthene with stratigraphic position (Naldrett et al. 1970). From the base upward the data, in general, show the normal trends within a fractionating magma system, i.e., a decrease in the An content of plagioclase and an increase in the Fe content in augite and hypersthene. Note that pyroxenes from the two lower units (quartz-rich norite and mafic norite) show a reversal of this trend whereby the Fe content of the pyroxenes increases toward the footwall. The reason for this trend is uncertain. Perhaps it is due to the assimilation of footwall crustal rocks by the SIC melt body or to the *in situ* crystallization of a non-cumulus rock (Naldrett et al. 1970). Pattison (1979) documented similar trends within the Contact and Offset sublayers whereby the Fe content of pyroxenes increases toward footwall contacts.

CONCLUSIONS

Available evidence suggests that the Main Mass formed by a process of fractional crystallization of a crustally derived melt body. The body was likely produced as a melt pool within an impact crater, perhaps as large as 250 km in diameter. Because this melt body may have been superheated well above liquidus temperatures (1700° to 2000°C), the melt had the capacity to assimilate considerable volumes of crustal material. The aberrant, silica-rich composition of this evolving melt is a consequence of its derivation from, and subsequent contamination by, continental crust.

Contact Sublayer

Based on studies at several classic mining sites, Souch et al. (1969) originally defined the Sublayer as "a distinctive inclusion-bearing facies of the Nickel Irruptive". This definition included the Contact Sublayer, quartz diorite dikes of the Offset Sublayer and Footwall Breccia. Modern usage excludes Footwall Breccia from the Sublayer.

DISTRIBUTION, OCCURRENCE AND GENERAL CHARACTERISTICS

Contact Sublayer comprises a variety of non-cumulative, igneous-textured gabbronoritic bodies. They occur as discontinuous lenses and irregular sheet-like bodies, at the contact between the Main Mass of the SIC and the underlying footwall rocks, on all three ranges (Figures 8.1 and 8.2). The thickness, which locally exceeds 700 m, is controlled by the topography of the footwall contact.

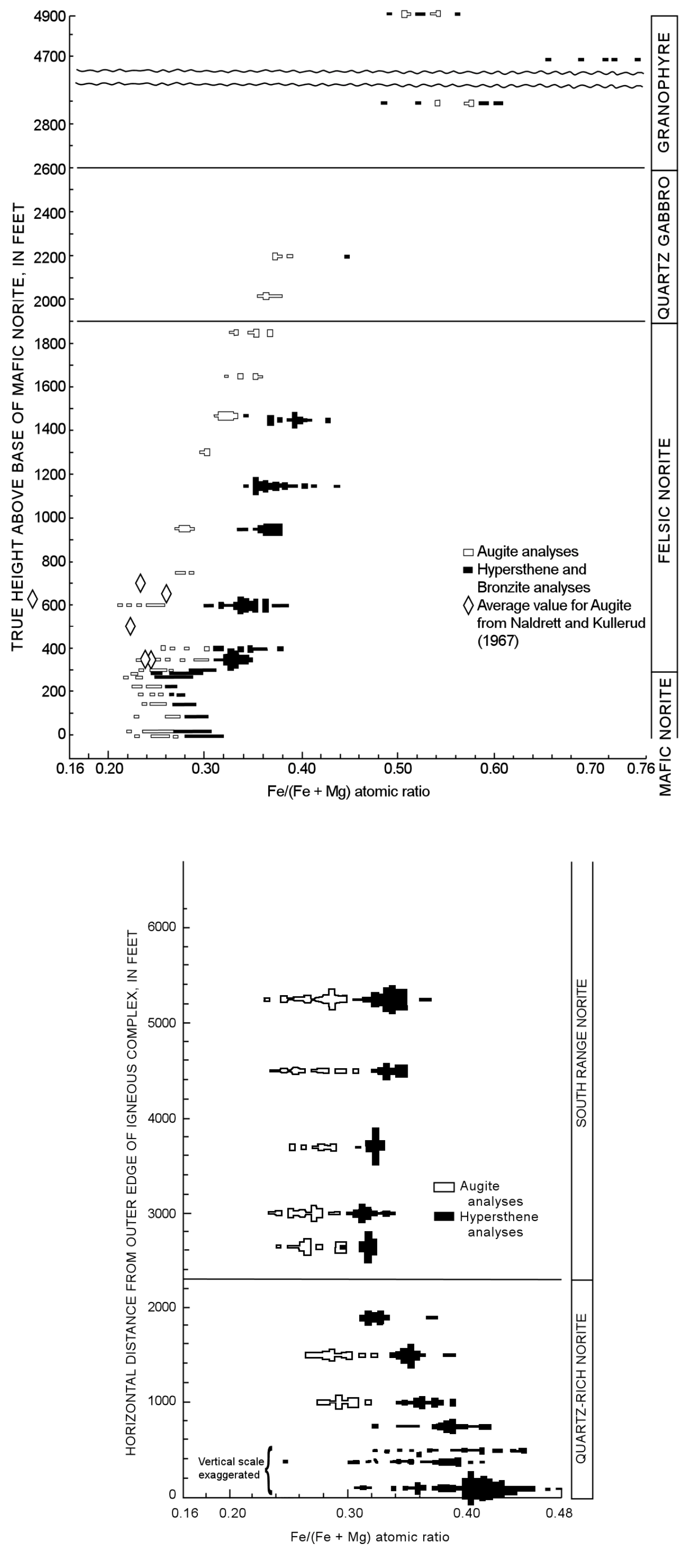


Figure 8.7. Diagrams showing the variations in pyroxene compositions through the Main Mass of the SIC on the North Range (top) and South Range (bottom) (*after* Naldrett et al. 1970).

Major bodies of Contact Sublayer are emplaced within kilometre-scale radial depressions or "troughs", along which are developed smaller secondary embayments or "terraces" (Morrison 1984). Figure 8.8 illustrates an idealized cross-section through a North Range embayment structure. Bodies of Contact Sublayer extend into the footwall and are overlain, by a relatively planar contact, with mafic and felsic norite of the Main Mass. Morrison (1984) interpreted the terraces as slump features, similar to those visible in large lunar and Martian impact craters. Apparently, they formed during the initial stage of crater collapse. However, the terraces and troughs may represent mechanical thermal-erosion channels formed by the melting and assimilation of footwall rocks by a superheated impact melt sheet (C.M. Lesher, Department of Earth Sciences, Laurentian University, personal communication, 2006).

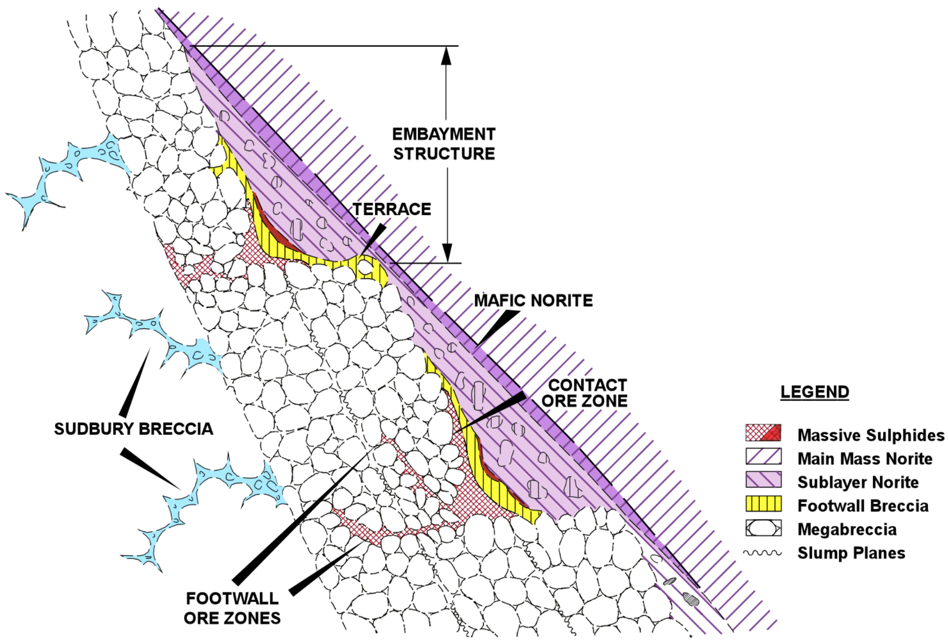


Figure 8.8. Schematic vertical section (not to scale) through part of an idealized embayment structure on the North Range of the Sudbury Basin, showing the development of two terraces, the formation of thicker bodies of Contact Sublayer within the terraces and the relationship with brecciated footwall and the overlying Main Mass of the SIC (*after* Morrison 1984). Note the development of mafic norite above the sublayer-filled terraces.

On the South Range, Contact Sublayer overlies brecciated Paleoproterozoic supracrustal rocks and coeval mafic and felsic plutons of the Southern Province. On the North Range, Contact Sublayer overlies brecciated Archean migmatites and granitoids; in mineralized environments, a zone of Footwall Breccia is frequently present between the Sublayer and footwall.

The contact between Contact Sublayer and the overlying Main Mass ranges from sharp to gradational. The relative age of the two units is controversial as it is based on the reported presence of inclusions of one unit in the other. Possibly the units are coeval, with the contradictory field relationships due to post-emplacement readjustments.

The Contact Sublayer is a mixture, in varying proportions, of three components: igneous silicate matrix, magmatic Cu-Ni-Fe sulphides and inclusions of silicate rock. These components can be present in any combination, including massive sulphide material with no silicate component to unmineralized silicate rock.

Inclusions of silicate rock, an almost ubiquitous feature in Contact Sublayer, range in size from single mineral xenocrysts to bodies several tens of metres in diameter. Three main types of inclusions are commonly present. These are as follows.

1) Fragments of identifiable country rocks, including those from mafic and ultramafic bodies in the footwall of the North (Moore et al. 1993) and South ranges.

2) Cognate xenoliths, related to the SIC. These include poikilitic melanorites and olivine melanorites. They are similar to rocks of the basal mafic norite of the Main Mass.

3) Exotic inclusions, ranging from anorthosite to dunite, which have no known source (Scribbins et al. 1984).

PETROGRAPHY

The Contact Sublayer is a highly variable rock in terms of both texture and mineral composition. It is a fine- to medium-grained gabbronoritic rock composed mainly of orthopyroxene, clinopyroxene, and plagioclase. Olivine is sporadically present, along with variable amounts of quartz, granophyre and K-feldspar as interstitial components. Magmatic sulphide minerals vary from nil to 100 modal %. The texture is commonly intersertal to subophitic. Porphyritic phases, characterized by large prismatic orthopyroxene grains, occur locally. Photomicrographs of some typical textures are shown in Photo 8.3.

LITHOCHEMISTRY

Because of the sulphide- and inclusion-rich nature of the Sublayer, it is difficult to characterize the lithochemical affinities of the silicate matrix component. In general, Contact Sublayer is of tholeiitic composition, which corresponds closely to the gabbronoritic petrography. The best way to compare different volumes of Sublayer rock is to standardize them against a common entity. The usual practice at Sudbury is to compare them with the average composition of felsic norite from the North Range (Lightfoot et al. 2001).

Figure 8.9 presents examples of extended-element, felsic-norite-normalized, spidergrams for Contact Sublayer from several locations in the North and South ranges. Contact Sublayer rocks from different locations display widely variable lithochemical characteristics, which implies considerable local control over the composition of the Contact Sublayer. This control may be best explained by the assimilation of various amounts of local footwall material. The amount and type of crustal contamination might be inferred from the intensity of Th-U and Zr-Hf peaks on the spidergrams. The more negative the peaks, the less the amount of crustal contamination.

Photo 8.3. Photomicrographs of Contact Sublayer, showing differing compositional and textural variants. From top to bottom these are: olivine-rich xenolithic norite, xenolithic norite, orthopyroxene-rich sublayer norite, and porphyritic sublayer norite. Each image is approximately 4 mm wide.

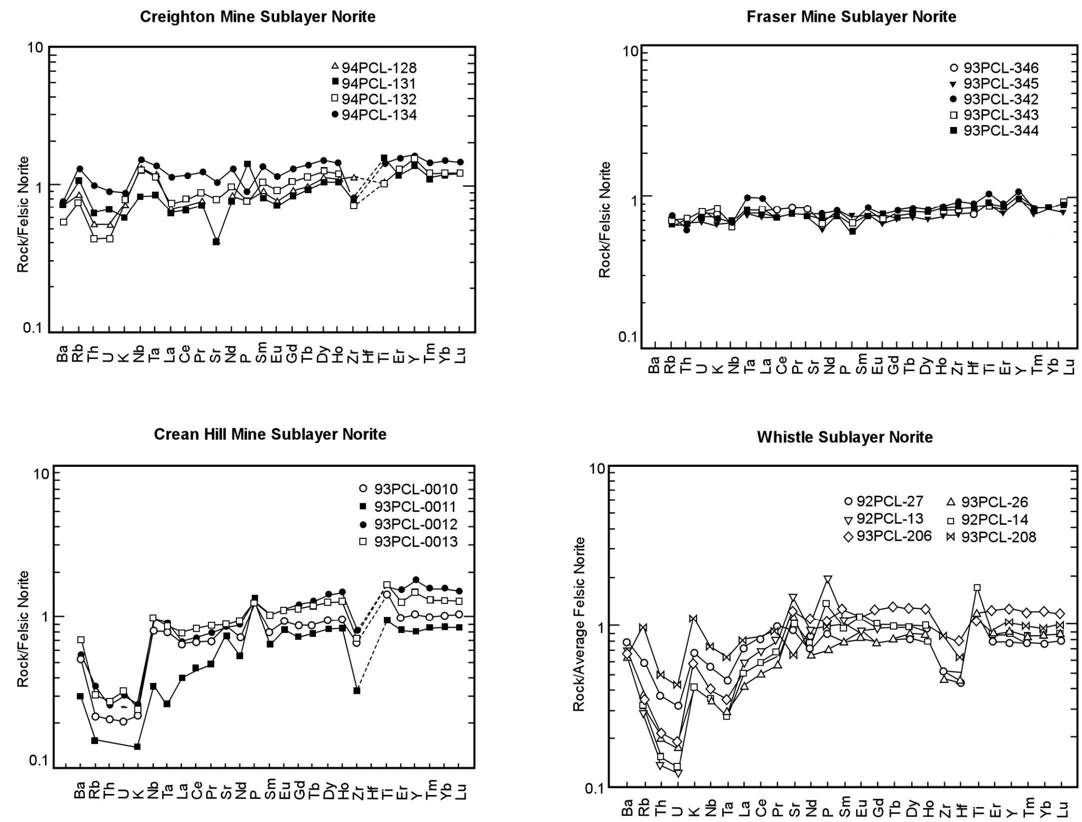


Figure 8.9. Felsic-norite-normalized, extended-element spidergrams for Contact Sublayer from various locations on the North and South ranges of the SIC, showing the similarities in composition within a single embayment contrasted to the differences between different embayments (*after* Lightfoot, Doherty et al. 1997; Lightfoot et al. 1997b).

Offset Sublayer

DISTRIBUTION, OCCURRENCE AND GENERAL CHARACTERISTICS

Offset Sublayer, commonly referred to as Quartz Diorite or QD, comprises a variety of non-cumulate, sometimes quench-textured, igneous rocks that form a number of relatively thin, dike-like intrusions in the footwall of the North and South ranges (Figure 8.1). Three groups of offset dikes were recognized by Grant and Bite (1984).

1) Radial offsets: radiate from the base of the Main Mass of the SIC and include Copper Cliff, Worthington, Ministic, Foy, Whistle–Parkin, and the recently discovered Trill and Pele offsets.

2) Concentric offsets: are oriented sub-parallel to the base of the SIC and include Hess and Manchester.

3) Discontinuous or breccia-hosted offsets: consist of isolated lenses, of various sizes, within zones of Sudbury Breccia. Examples of this group include the MacLennan and Frood–Stobie. The latter was formerly classified as a concentric offset. Small metre- to decimetre-scale QD bodies of this type are commonly referred to as melt pods.

Radial dikes intersect the SIC at various angles and extend as much as 30 km into the footwall. Distances from the base of the SIC include 12 to 15 km for the Hess offset, 5 km for the Manchester offset and 1 to 2 km for discontinuous or breccia-hosted offsets. Offset dikes vary in thickness from several hundred metres near the SIC contact to a few metres at the distal extremities.

There is no obvious structural control as to the distribution of radial offsets. They usually emanate from embayment-like depressions, of various sizes, at the base of the SIC. Embayments are commonly filled with Contact Sublayer. However, the nature of the change from Contact to Offset Sublayer is unclear. In the South Range footwall, the radial dikes cross-cut all lithologic units of the Southern Province as well as fold hinges whose formation is attributed to the Penokean Orogeny. Apparently, dike emplacement post-dated the main deformational phase of the orogeny. The offset dikes are displaced by faults (late Penokean?) and cut by undated mafic dikes (trap and quartz diabase) and by olivine diabase dikes of the Sudbury swarm (1238 Ma; Krogh et al. 1987). Similar relationships are evident in the North Range footwall. Concentric offsets are essentially parallel to local geologic trends but are otherwise similar to the radial offsets. According to Grant and Bite (1984), both radial and concentric offsets were emplaced in enhanced, but discontinuous, zones of Sudbury Breccia. In contrast, the discontinuous offsets are found within major, continuous zones of Sudbury Breccia that have been interpreted as possible crater collapse structures (Scott and Spray 2000). A zircon-baddeleyite U-Pb crystallization age of 1852 + 4/-3 Ma has been reported for the Foy offset (Ostermann et al. 1996). Within error, this is identical to the age of the Main Mass of the SIC of 1850 Ma (Krogh et al. 1984).

INTERNAL ARCHITECTURE OF OFFSETS

Grant and Bite (1984) present a comprehensive review of the characteristics of many offset dikes. Later detailed work on individual dikes include Worthington (Lightfoot and Farrow 2002), Foy (Tuchscherer and Spray 2002), Whistle–Parkin (Murphy and Spray 2002), Copper Cliff (Mourre 2000) and Hess (Wood and Spray 1998). Major characteristics of radial and concentric offsets can be summarized as follows.

1) Most offsets appear to be composite intrusions consisting of two phases: a central core of inclusion- and sulphide-enriched quartz diorite encased in a marginal sheath of essentially inclusion- and sulphide-free quartz diorite. The nature of the contact relationships between the two phases are often obscure. However, Lightfoot and Farrow (2002) describe, from Worthington, xenoliths of the inclusion-free phase in the inclusion- and sulphide-bearing quartz diorite. Flow differentiation is a possible alternative mechanism for producing the apparent composite nature of the offsets. Heavy components in a moving magma, such as sulphide droplets, tend to move toward the centre of the dike (Bhattacharji and Smith 1964).

2) The proximal portion of some radial offsets are partially occupied by Contact Sublayer rocks (e.g., Foy, Pattison 1979; Copper Cliff, Capes 2001) and/or Footwall Breccia (Foy, Pattison 1979; Whistle–Parkin, Murphy and Spray 2002). The relationships between these rocks and contact quartz diorite is uncertain.

3) Although Grant and Bite (1984) suggest that the contacts with country rocks show little or no chilling, many examples of chilled, even quenched, quartz diorite are known. These include: small dikelets of aphanitic, devitrified glass with local variolitic texture; skeleton pyroxenes; branching dendritic pyroxenes; and clusters of elongate pyroxene crystals that form a spherulitic texture. These features indicate rapid crystallization from a rapidly cooled, possibly super-heated melt.

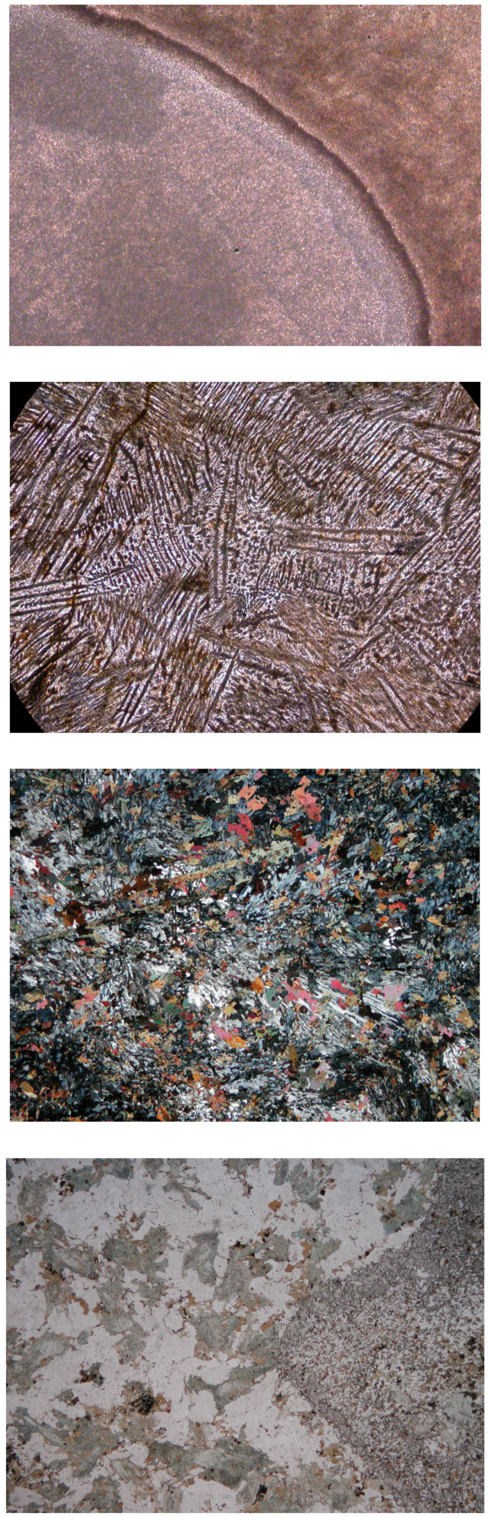


Photo 8.4. Photomicrographs of Offset Sublayer showing different textural variants. From top to bottom these are: quenched aphanitic variety from the Trill Offset with well-developed variolitic texture (the light-coloured sphere is about 2 cm in diameter); quenched variety from Trill showing well-developed skeletal pyroxene; chilled variety from Copper Cliff showing elongate prisms of altered pyroxene and sheath-like plagioclase; and typical inclusion-bearing amphibole-biotite Quartz Diorite from Copper Cliff. All images, except the uppermost, are approximately 4 mm wide.

4) Some offset dikes show features such as bends, discontinuities and gaps.

5) Offset Sublayer represents a mixture, in varying proportions, of inclusions of magmatic Cu-Ni-Fe sulphides and silicate rock in an igneous silicate matrix.

6) Within the central core of inclusion- and sulphide-enriched quartz diorite, zones of ore occur in the form of steeply plunging pipes. The Copper Cliff Offset is the most intensely mineralized offset (Figure 8.10). There the sulphides occur as disseminated centimetre-scale blebs and as semi-massive to massive accumulations of pyrrhotite, pentlandite and chalcopyrite.

7) Silicate inclusions in quartz diorite are similar to those found in Contact Sublayer and comprise country rock, possible cognate inclusions related to the SIC and exotic inclusions of various types.

PETROGRAPHY OF OFFSET SUBLAYER

In the footwall of the North Range, Offset Sublayer is relatively unaltered. It consists of lathy, intermediate plagioclase, acicular to prismatic ortho- and clinopyroxene, minor primary biotite and amphibole and highly variable amounts of interstitial quartz and granophyre. The normal texture varies from equigranular to varieties characterized by very elongate prismatic orthopyroxene crystals, up to several centimetres in length, which locally occur as spherulitic clusters. In the distal portion of some offsets and in small offsets (e.g., Trill) are textures attributed to rapid quenching. These include clusters of dendritic pyroxene and, rarely, aphanitic material, possibly originally glassy, within thin quartz diorite dikes. Variolitic textures within the aphanitic dikes are reminiscent of those found in volcanic rocks. Offset Sublayer, in the footwall of the South Range, is altered. Fresh amphibole is rare as it is replaced by aggregates of hydrothermal biotite and amphibole. Locally, it can be recognized that the secondary minerals represent pseudomorphs after original pyroxene but, in general, there is no evidence of original mineralogy, and the rocks are referred to as amphibole-biotite quartz diorite. See Photo 8.4 for textural variations in the Offset Sublayer.

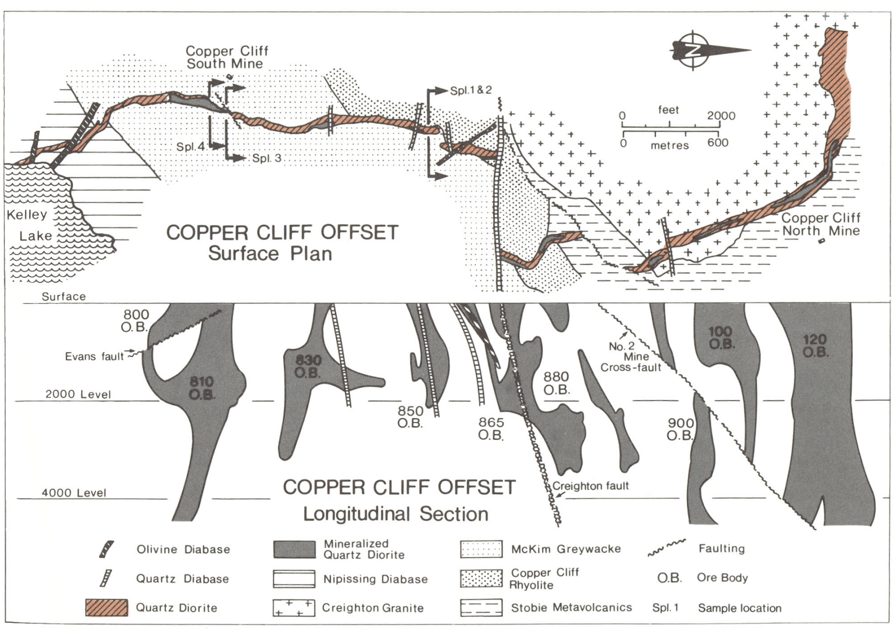


Figure 8.10. Plan and longitudinal section (looking west) showing the distribution of orebodies within the Copper Cliff Offset (*from* Cochrane 1984). Note the vertical to steeply plunging attitude of the individual bodies.

LITHOCHEMISTRY OF OFFSET SUBLAYER

Despite textural and mineralogical variations, the presence of varying amounts of sulphide and silicate rock inclusions and the evidence for extensive assimilation of a variety of footwall lithologies, the fundamental chemical compositions of the various offsets are remarkably similar. Figure 8.11 sets out four felsic-norite-normalized, extended-element spidergrams from several offsets from the North and South range footwalls. Note the restricted range in composition from a single offset and the similar pattern for all the offsets.

These data suggest that the offset dikes were derived from a homogeneous magma. Variations in the abundances of some elements between different offsets and along the lengths of individual offsets can be explained by local differences in adjacent wall rocks.

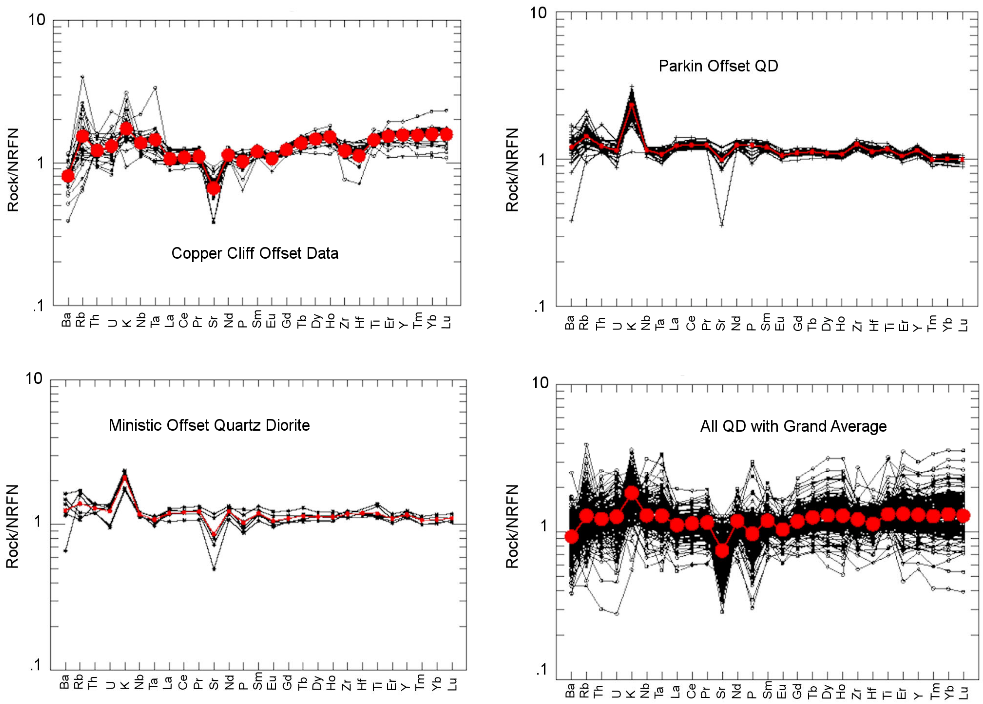


Figure 8.11. Felsic-norite-normalized, extended-element spidergrams from various offsets on the North and South ranges of the SIC (*from* Lightfoot et al. 1997a). Note the restricted range in composition within a single offset as well as the generally similar patterns displayed by all the offsets as shown in the bottom spidergram. Compare this to the data for the Contact Sublayer shown in Figure 8.9. Red dots and lines are average values. QD = Quartz Diorite.

Chapter 9

Structural Geology

D.H. Rousell

Introduction

The Sudbury area has undergone two episodes of ductile deformation which were spread over a protracted period of time. Brittle deformation is manifested by several fault groups, some of which have been reactivated by new stress regimes. In this chapter, emphasis is placed on the structure of the Sudbury Basin and the Sudbury Igneous Complex (SIC). The structure of the Southern Province, south of the SIC, is also briefly described. See Chapter 2 for a discussion of the structural geology of Archean rocks (Superior Province) in the footwall of the North and East ranges. Models for the present disposition of the SIC are included in Chapter 8. The major structural elements of the Sudbury area are set out in Figure 9.1.

Folds

Figure 9.2 plots the trace of the axial surfaces and the plunge of the hinges of map-scale folds in the Chelmsford Formation and several small-scale folds in the Onwatin Formation. The map also shows the traces of northeast-striking faults, three of which displace axial surface traces in a dextral strike-slip sense. There is a tendency of folds in the southwest corner of the belt to plunge to the northeast and those in the northeast to plunge southwest, but many folds are doubly plunging (Photo 9.1). Note the rather rare northwest-trending axial surface traces. The rose diagram of the strike of the axial surfaces of map-scale folds (Figure 9.2, inset A) indicates that the strike of the major class-interval is 063°.

Figure 9.2, inset B, is a stereoplot of fold hinges and poles to bedding planes. The plunge of map-scale folds is shallow, between 10° and 15°, in a northeast or southwest direction. Small-scale fold hinges plunge as much as 60°. The normal to the pole circle represents the approximate mean attitude of fold hinges which is, in terms of plunge and trend, 00° 064°. The close coincidence between the strike of the axial surfaces and the mean trend of the hinges suggests that the folds are upright, or nearly so. The distribution of poles to bedding planes indicates an inter-limb angle of approximately 100°, which suggests that the folds may be classified as open (Park 1997).

A map in the area of the Vermilion and Errington mines (Figure 9.3) plots the plunge of fold hinges and the trace and dip (where known) of fold axial surfaces. Some folds are doubly plunging while others plunge to the northeast or to the east-southeast. Fold axial surfaces strike to the northeast. Many folds are inclined (overturned?) to the northwest. The stereogram (inset of Figure 9.3) indicates that folds plunge as much as 37° east-northeast and up to 55° east-southeast. The mean orientation of the hinges of these easterly plunging folds (N = 11), in terms of plunge and trend, is 30° NE 072°. The mean orientation of fold axial surfaces is 059° 64° SE, indicating that the folds are inclined, perhaps overturned, to the northwest.

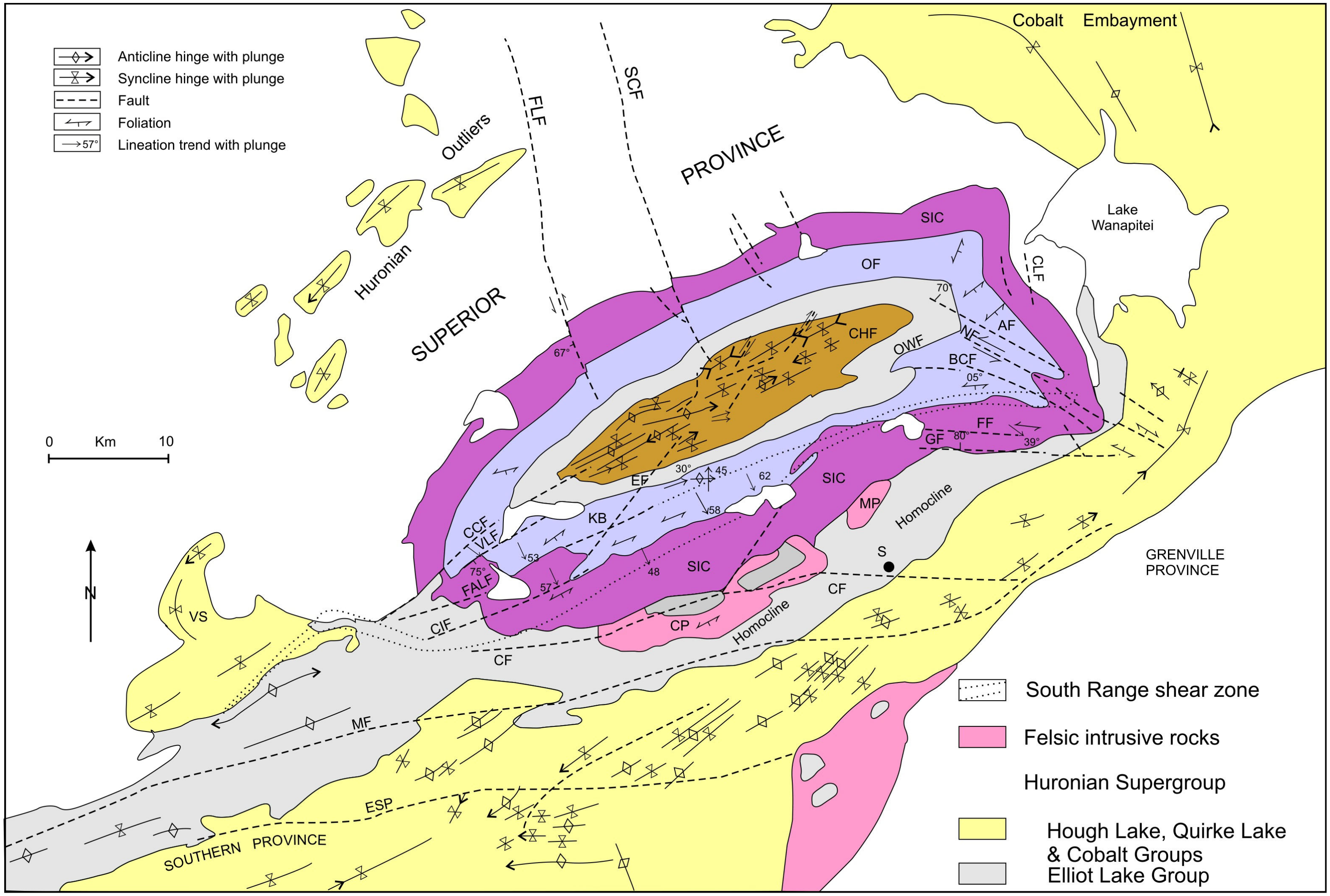


Figure 9.1. Map showing the major structural elements in the Sudbury area. AF = Airport Fault, BCF = Bailey Corners Fault, CCF = Cameron Creek Fault, CF = Creighton Fault, CHF = Chelmsford Formation, CIF = Chicago Fault, CLF = Capre Lake Fault, CP = Creighton pluton, EF = Errington Fault, ESP = Espanola Fault, FALF = Fairbank Lake Fault, FF = Falconbridge Fault, FLF = Fecunis Lake Fault, GF = Garson Fault, KB = kink bands, MF = Murray Fault, MP = Murray pluton, NF = Norduna Fault, OF = Onaping Formation, OWF = Onwatin Formation, SCF = Sandcherry Creek Fault, S = Sudbury, SIC = Sudbury Igneous Complex, VLF = Vermilion Lake Fault, VS = Vernon Syncline.

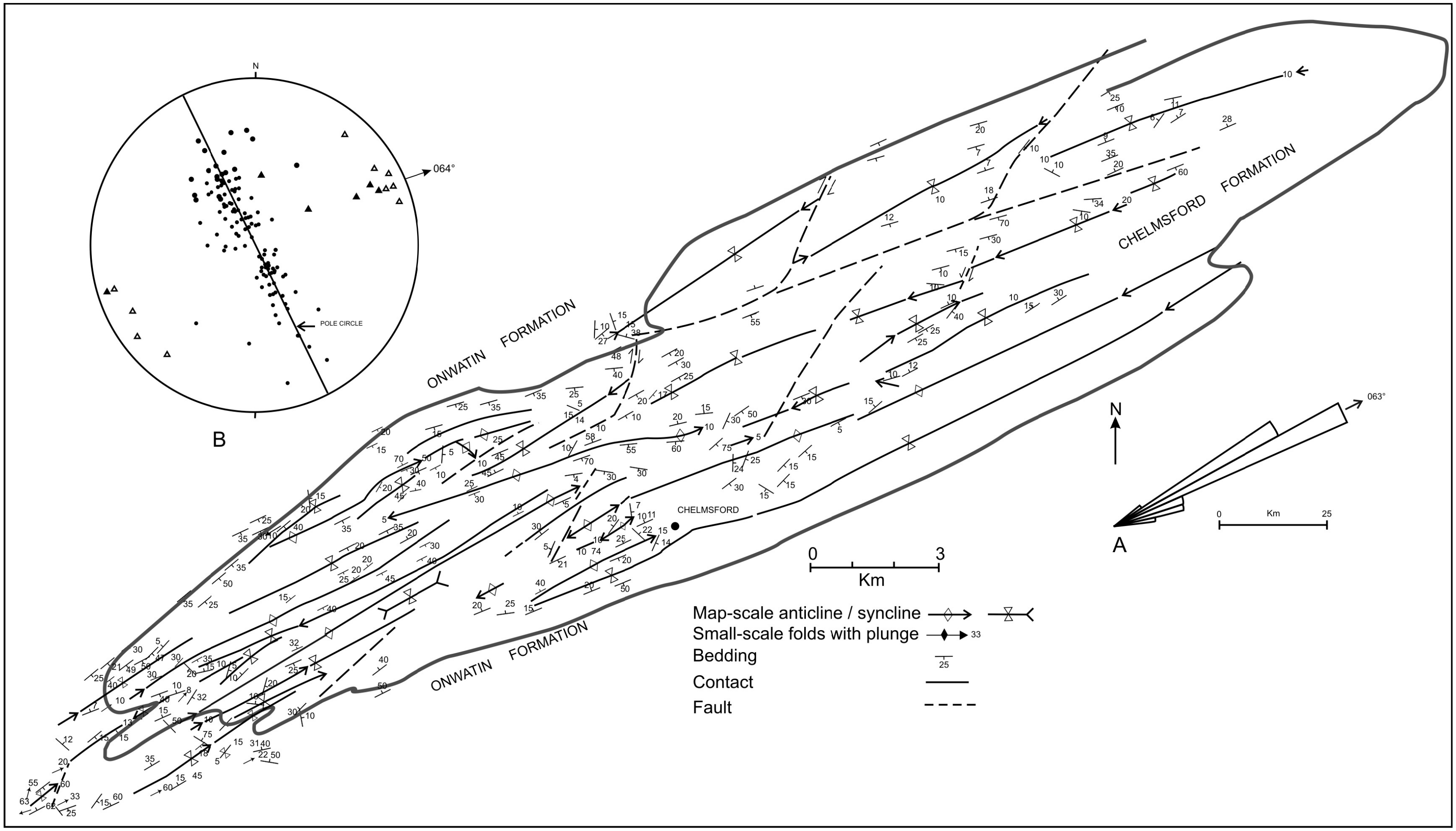


Figure 9.2. Map indicating the traces of the axial surfaces, the plunge of the hinges of folds, and the attitude of bedding surfaces in the Chelmsford and Onwatin formations (*from* Card 1967, unpublished map of K.D. Card and data of the author). **Inset A** is a rose diagram of the strike of the traces of the axial surfaces of the folds. The scale refers to the cumulative length of the traces within each 5° class interval. **Inset B** represents a lower hemisphere equal-area plot of: 1) poles to bedding planes (solid dots, 1 per 0.8 km^2), N = 115, with the normal to the pole circle trending 064°; 2) map-scale fold hinges (open triangles) in the Chelmsford Formation (N = 7) and the Onwatin Formation (N = 1); and 3) small-scale fold hinges (closed triangles) in the Onwatin Formation (N = 6).

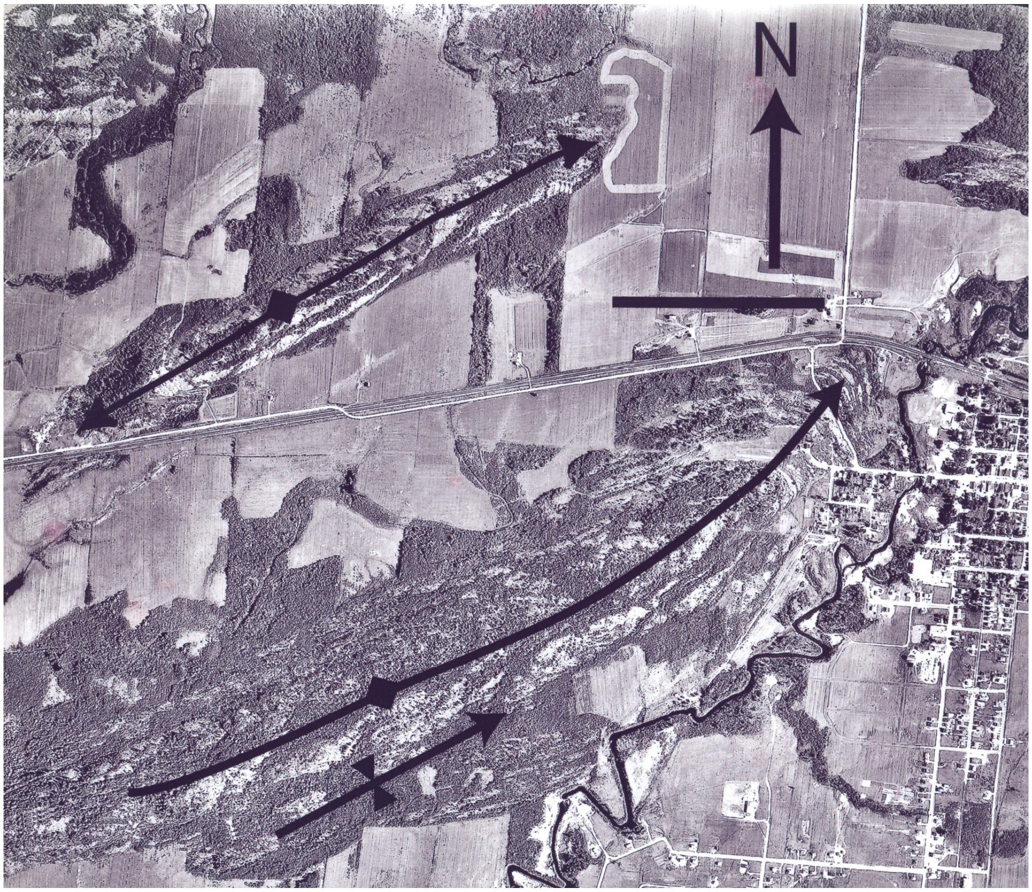


Photo 9.1. Aerial photo of folded rocks of the Chelmsford Formation. The large anticline, located west of the town of Chelmsford, plunges 15° NE. Northwest of the town, a smaller anticline doubly plunges 10°NE and 10°SW. Scale bar is approximately 0.57 km. (Air photo courtesy of the Ministry of Natural Resources.)

Photo 9.2 shows an open fold from the South Fold Nose (see Figure 9.3 for location). The attitude of the hinge is 54°E 100°. Structural data (Stoness 1994) from the area of the South Fold Nose are plotted on the stereogram of Figure 9.4. The attitude of the mean of the fold hinges is 53°E 109°. The cleavage in the area is approximately axial planar so that the mean cleavage suggests a mean axial plane of 084° 82°S. The lack of exposure of the north limb of the inferred map-scale fold precludes obtaining a pole circle for poles to bedding planes.

A geological map of the 500 foot level (152 m) and a cross section, in the area of the Errington No.2 shaft, is set out in Figure 9.5 (Martin 1957; Thomson 1957). The map shows a number of folds which, in general, plunge away from an arcuate arch that trends in a northerly direction. The section indicates steeply dipping, nearly isoclinal folds that are overturned to the north-northwest and displaced by steeply dipping reverse faults.

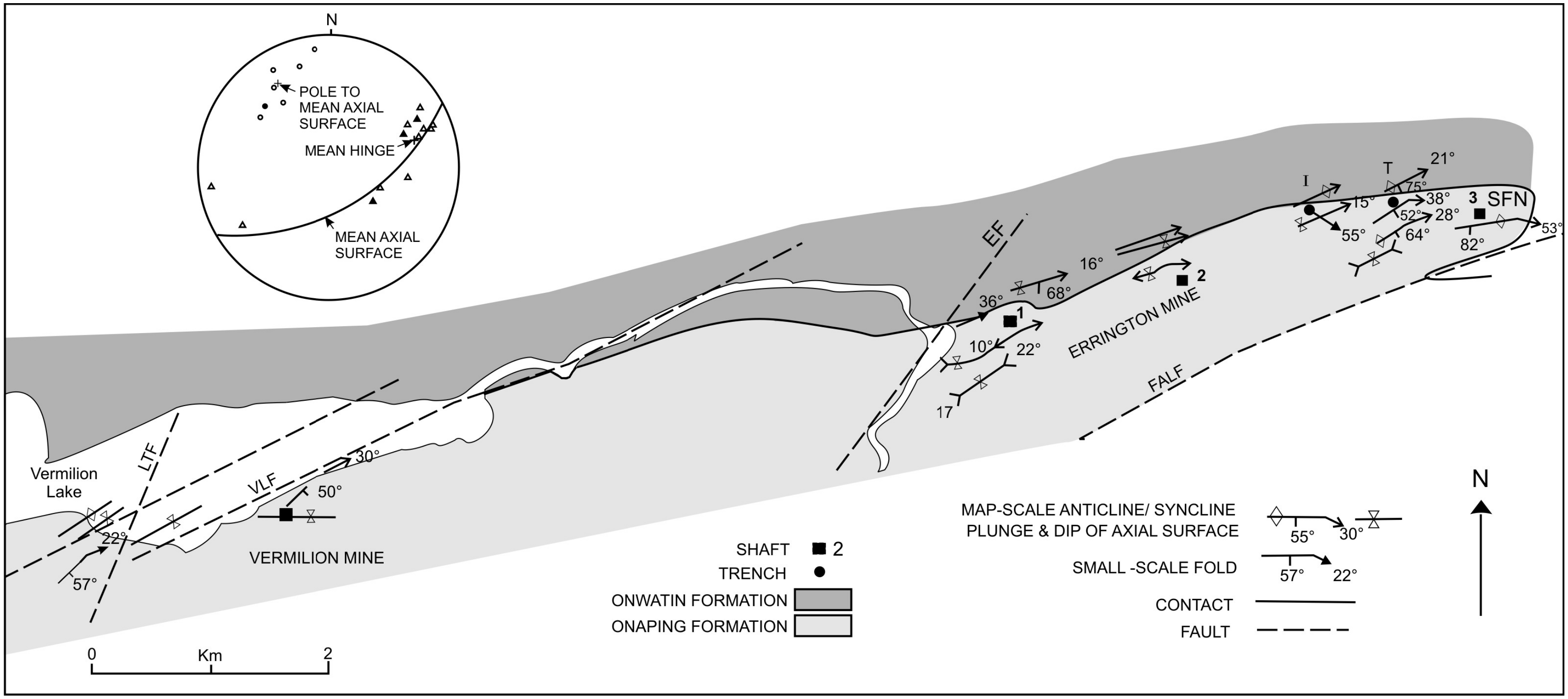


Figure 9.3. Map of the Vermilion and Errington mines area showing the trace and dip (where available) of the axial surfaces and the plunge of the hinges of folds. The trace of major faults is also shown (Martin 1957; Thomson 1957; Dressler 1984a; Paakki 1992; Stoness 1994; Gray 1995; and data of the author). EF = Errington Fault, FALF = Fairbank Lake Fault, I = Irwin trench, LTF = late thrust fault, SFN = South Fold Nose, T = Tounigy trench, VLF = Vermilion Lake Fault. 1, 2 and 3 = shafts of the Errington Mine. **Inset** is a lower hemisphere equal-area plot of fold data from the Vermilion and Errington mines area. Map-scale fold hinges (N = 10) are open triangles, small-scale fold hinges (N = 3) are closed triangles. The poles to the axial surfaces of map-scale folds (N = 6) are open circles and the pole to a small-scale axial surface is a closed circle. The mean orientation of the hinges (cross) of easterly plunging folds (N = 11) is 30° NE 072°. The mean orientation of fold axial surfaces is 059° 64° SE. Means were determined by the statistical method of Ramsay (1967).

Photo 9.2. Looking east at an open fold from the South Fold Nose. The fold, in the Grey Argillite Member of the Vermilion Formation (Stoness 1994), plunges away from the viewer. The attitude of the hinge is 54°E 100°. Note the ptygmatic folds in the dark, centimetre-scale, competent sandstone layers. (See Figure 9.3 for the location of the South Fold Nose.)

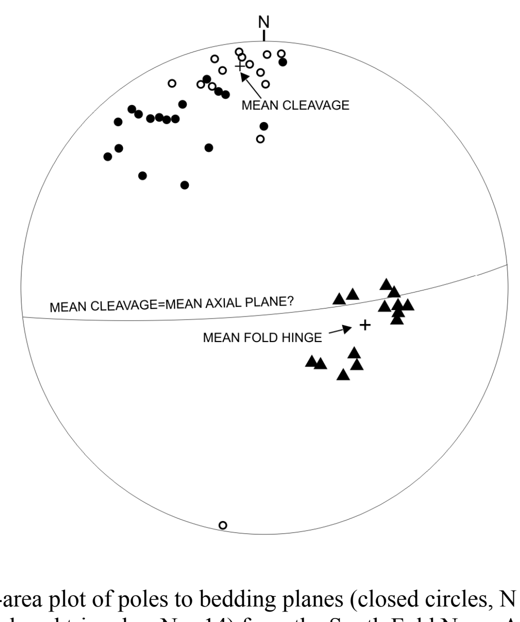


Figure 9.4. Lower hemisphere equal-area plot of poles to bedding planes (closed circles, N = 18), cleavage planes (open circles, N = 14) and small-scale fold hinges (closed triangles, N = 14) from the South Fold Nose. A mean fold hinge of 53°E 109° and a mean pole to cleavage planes were determined by the statistical method of Ramsay (1967). Assuming the cleavage is axial planar, then the mean fold axial plane is 084° 82°S. (*Data from* Stoness 1994.)

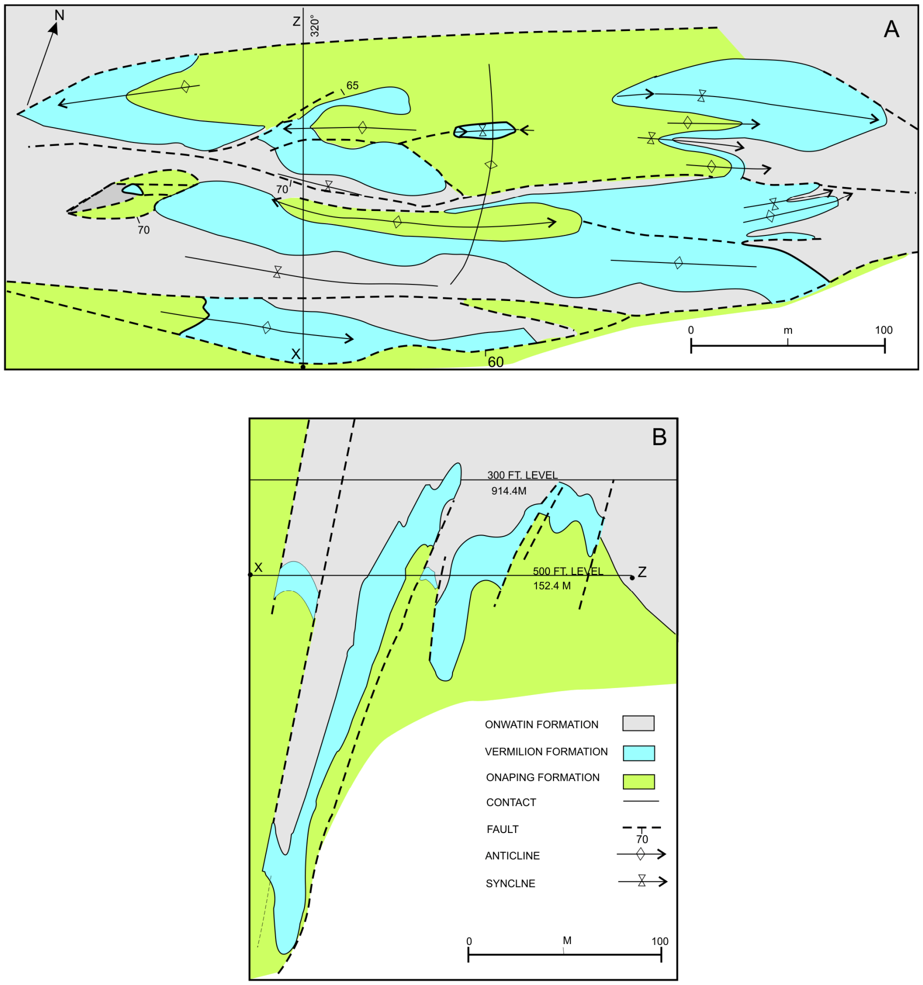


Figure 9.5. A. Geological map of the 500 foot (152 m) level of the Errington Mine. **B.** Vertical section, looking west, through line X-Z (*after* Martin 1957; Thomson 1957). The No.2 shaft is located, on the map, approximately 300 m east of point X.

Foliation

The map of Figure 9.6 plots foliation planes in rocks of the Sudbury Basin including rough cleavage in the Chelmsford Formation, slaty cleavage in the Onwatin Formation and schistosity in the Onaping Formation and SIC (Rousell 1975; Shanks and Schwerdtner 1991; Cowan 1996; and unpublished data of the author). Note the prominence of northeast strikes and southeast dips. The SIC is unfoliated in the North Range and locally unfoliated in the South Range. The Onaping Formation is locally unfoliated in the North Range and, where present, tends to be weaker than that further south. The stereogram of poles to foliation planes (inset Figure 9.6) shows a strong cluster of poles in the northwest quadrant. The mean value, excluding four poles due to a limitation in the statistical method, is 060° 67°SE. Two excluded poles in the southeast quadrant represent cleavage in the Chelmsford Formation with dips of about 31° NW. The two excluded poles in the northeast quadrant are plots of foliations from the relatively large South Lobe, which strike east-southeast and dip southwest. Foliation in the smaller North Lobe strikes approximately 033°.

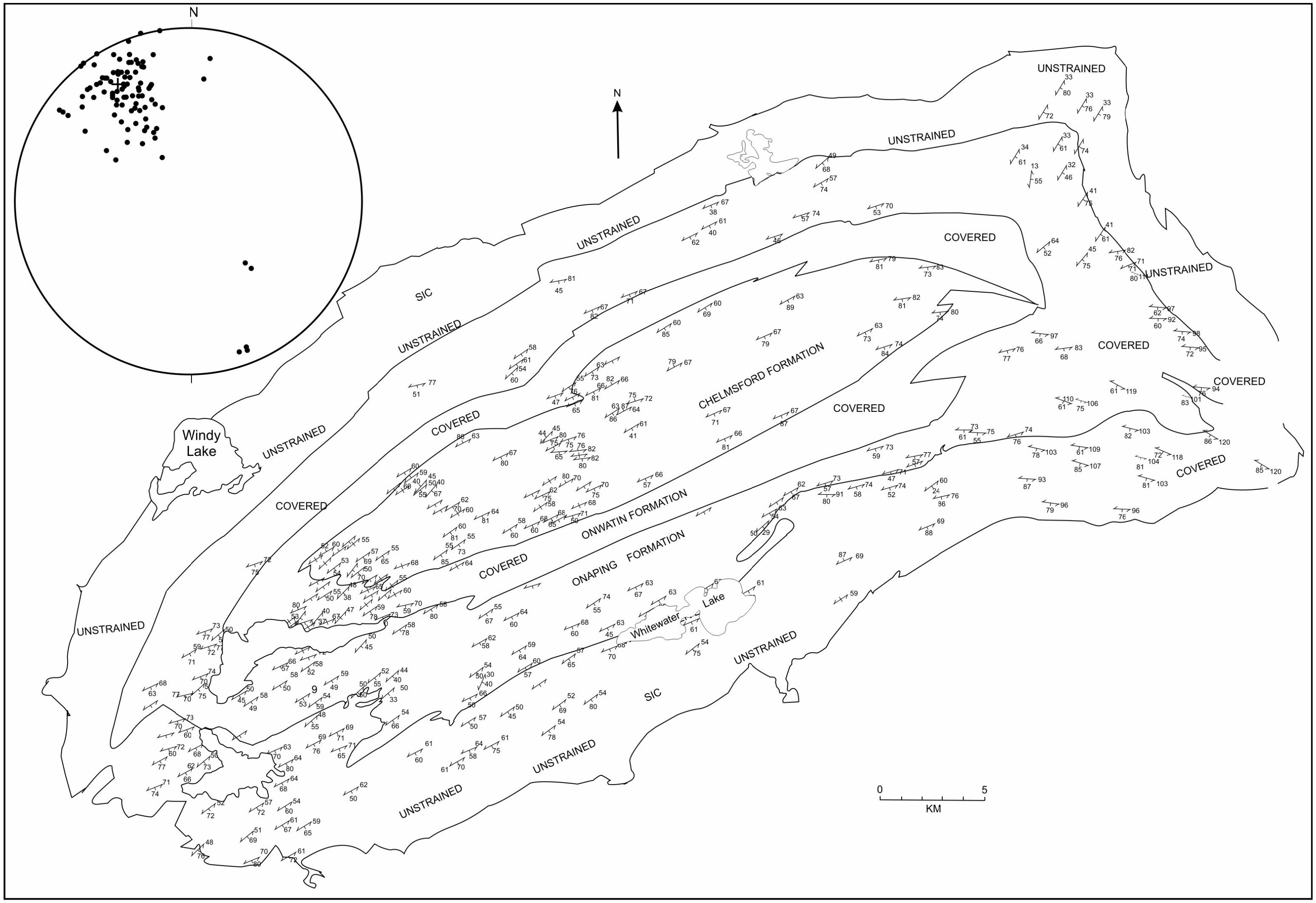


Figure 9.6. Map showing the attitude of foliation planes in rocks of the Sudbury Basin, including rough cleavage in the Chelmsford Formation; slaty cleavage in the Onwatin Formation; and schistosity in the Onaping Formation and Sudbury Igneous Complex (*from* Rousell 1975; Shanks and Schwerdtner 1992; Cowan 1996; and data of the author). **Inset** is a lower hemisphere, equal-area plot of poles to foliation planes, one per 1.2 km^2 (N = 112). The mean value is 060° 67° SE (marked by cross), excluding the two poles in the northeast quadrant (from the South lobe) and two poles to the relatively shallow-dipping foliations in the southeast quadrant (from the Chelmsford Formation). The exclusion of these poles is due to the limitations of the statistical method (Herget 1977).

Lineation

Stretching lineations are plotted on the map and stereogram of Figure 9.7. Excluding three plots, the average attitude, in terms of plunge and trend, is 59° SE 154°. Note the lack of lineations in most of the North Range and in the southeast portion of the South Range. The trend of the lineations is rather constrained, to the south-southeast, with the exception of easterly trends in the Northeast Lobe. Plunges, on the other hand, range between 32° and 82°. In general, lineations tend to plunge down-dip, except east of Whitson Lake and in the southern part of the East Range, where they pitch southeast.

Kink Bands

Rotation of the tectonic foliation in the Onaping Formation produced kink bands that are present at several locations along the entire length of the South Range. Kink bands are particularly well developed along the Gordon Lake Road (see Figure 9.1), where they are described in detail by Rousell (1980).

The kink bands are narrow with an average width of 0.8 cm and a maximum width of 4 cm. The average distance between kink bands is approximately 20 cm, but locally they are so closely spaced as to be penetrative (Photo 9.3). The external foliation strikes to the northeast and dips moderately to the southeast. The internal foliation, within the kink bands, has been rotated to the vertical and beyond so as to dip steeply to the northwest. The kink band boundaries, or kink planes, dip at shallow angles to the northwest and the hinge lines trend to the east-northeast and are nearly horizontal. The average orientation of these elements are plotted on the stereogram of Figure 9.8. Microfractures and former open spaces occur at kink band boundaries and are filled by quartz and calcite.

The angle between the external foliation and kink planes (α) is approximately 60° in slates and phyllites formed by the fixed-boundary mode of propagation. If slip between foliation planes is the sole operating mechanism, then rotation proceeded until the angle between the foliation plane inside the kink band and the kink plane (β) equals α. The kink band is then locked and no further rotation can take place. In the Onaping Formation kink bands, α is unusually large (average of 84°), exceeds that of β (average of 73°), and 86% of the kink bands are over-rotated (Figure 9.8).

The mechanism of formation of the Onaping kink bands is by slip along closely spaced foliation planes and along the margin of fragments, together with solution and redeposition, and possible ductile flow of the matrix. The large α angle is a function of planar anisotropy as it is more widely spaced and less well developed than in slates and phyllites. The amount of shortening achieved by kinking is trivial and is approximately 0.7% parallel to the external foliation. The axis of maximum principal stress that formed the kink bands is normal to the long axis of the Sudbury Basin and plunges between 15° and 50° southeast. The kink bands probably formed during a late stage of the Penokean Orogeny.

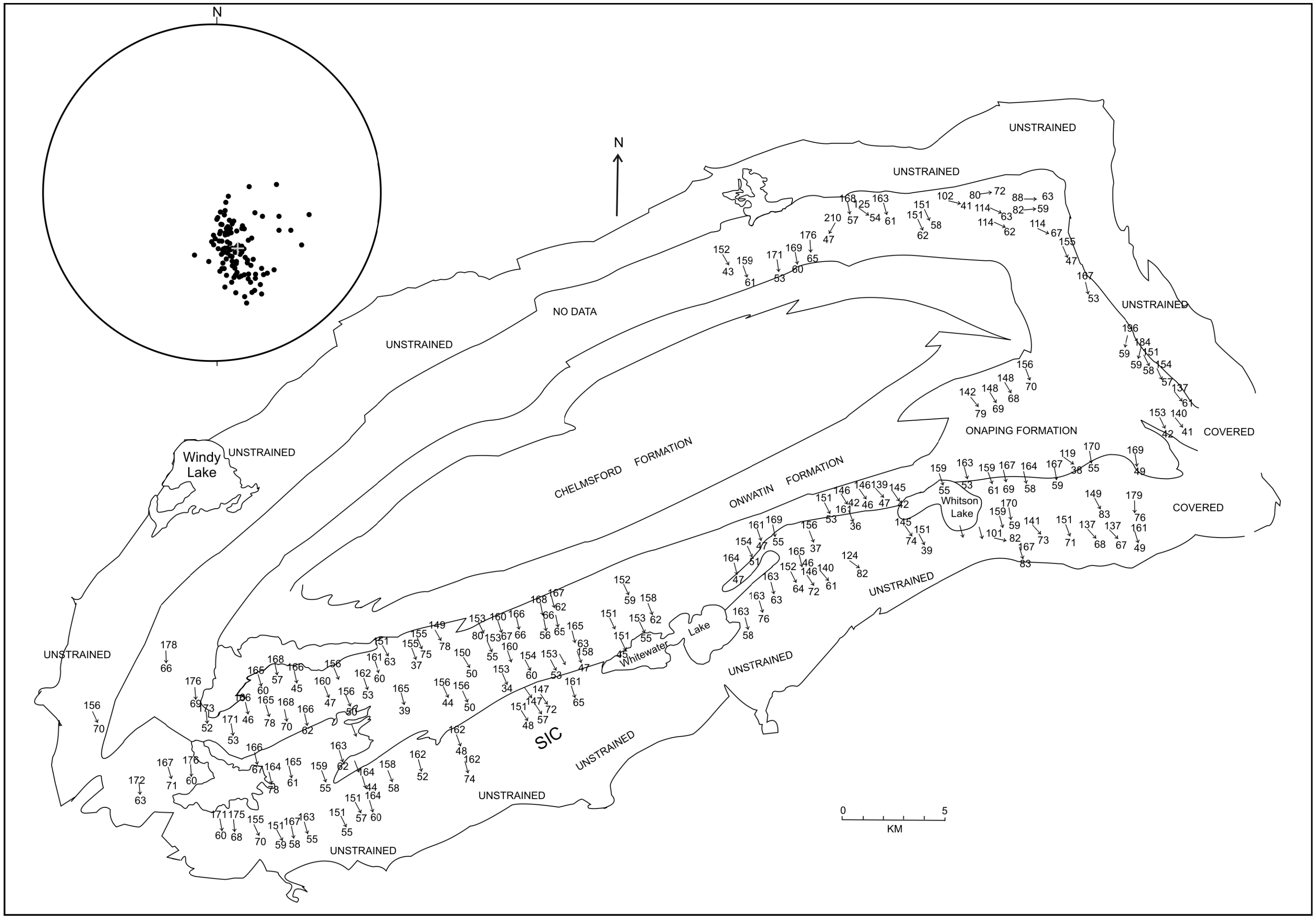


Figure 9.7. Map showing the attitude of stretching lineations in the Sudbury Basin (*from* Rousell 1975; Shanks and Schwerdtner 1991; Cowan 1996). **Inset** is a lower hemisphere, equal-area plot of stretching lineations, one per 1.6 km^2 (N = 109). The mean value is 59° SE 154° SE (marked by cross), excluding two plots in the northeast quadrant and one in the southwest quadrant. The exclusion of these poles is due to the limitations of the statistical method (Herget 1977).

Photo 9.3. Closely spaced kink bands in the Onaping Formation as exposed on the Gordon Lake Road. Ruler is 16 cm long. For location of the kink bands, see Figure 9.1.

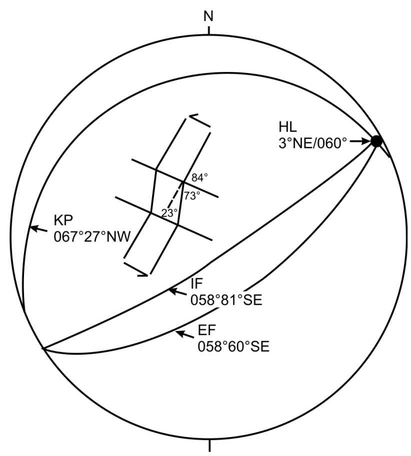


Figure 9.8. Lower hemisphere equal-area plot of the average orientations of elements of kink bands in the Onaping Formation (*from* Rousell 1984a). The kink bands are located on the Gordon Lake Road. EF = external foliation, IF = internal foliation, KP = kink plane, HL = hinge line. **Inset** is a profile view of the average kink band.

Sudbury Igneous Complex

The original disposition of the SIC is an unresolved problem of fundamental importance. Advocates of the meteorite impact model insist that the present shape of the SIC was due to the folding of a more-or-less flat impact melt sheet into a doubly plunging syncline. However, structural evidence suggests that the SIC was emplaced in approximately the present configuration (lopolith? ring dike?) with the South and East ranges steepened by later deformation. The original shape is evaluated in terms of the following: 1) ductile deformation in the North Range; 2) comparison with structures in the Southern Province; 3) significance of differences in dip of base and top of the SIC; and 4) attitude of fabric in the SIC.

The Onaping Formation in the South Range has undergone relatively intense ductile deformation, as shown by a strong tectonic foliation and stretching lineation and doubly plunging isoclinal folds that are locally overturned to the northwest. In contrast, the foliation in the North Range is local and weak and folds are lacking. In short, deformation intensity decreases from southeast to northwest across the basin. If the SIC achieved the present position by a rigid body rotation of a horizontal sheet, then the Onaping rocks in the North Range would be compressed and the SIC, at the North Lobe hinge, strongly strained. Field evidence, for large strains at the North Lobe and elsewhere in the SIC of the North and East ranges, is lacking (Rousell 1984a; Muir 1984; Cowan et al. 1999). The above evidence suggests the North Range was emplaced in approximately its present position.

Northeast-trending structural basins occur in the Southern Province (Card 1978b), although the largest (12 x 5 km) is considerably smaller than the Sudbury Basin. Thus, the present disposition of the SIC is compatible with structural elements in the Southern Province.

The weighted near-surface dips of the base of the SIC are 42° SE in the North Range and 65° NW in the South Range, excluding a 56°S dip at Garson (Souch and Podolsky 1969). In the East Range, dips at the base are 70° to 75° W, with a dip at Nickel Rim of 82° E (Figure 9.9). In the Lindsley Mine, the base of the SIC becomes overturned at a depth of 1200 m. For the North Range, the average dip at the top of the SIC is 25° SE. A melt sheet would probably have a more-or-less flat top and a curved, convex-upward base. If folded, the base would dip steeper than the top, as in the North Range. Note that Wilson's (1956) lopolith model for the SIC also has steeper dips at the base than at the top. On the other hand a ring dike presumably has approximately the same dips on the top and bottom.

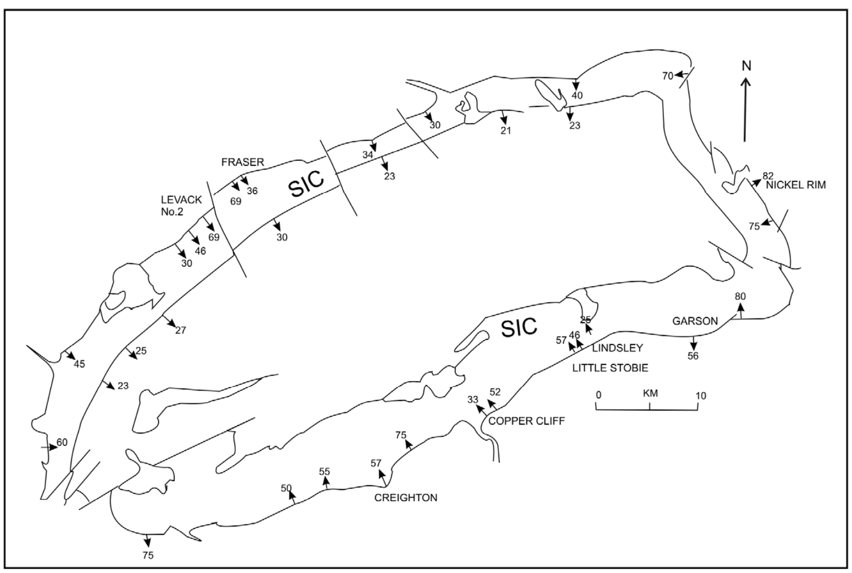


Figure 9.9. Map with near-surface dips at the base, within and at the top of the Sudbury Igneous Complex (SIC) (*from* Mamen 1955; Hawley 1965; Peredery 1972; Naldrett and Hewins 1984; Rousell 1984a; Cowan and Schwerdtner 1994; Riller and Schwerdtner 1997; Lisa Gibson, Inco Ltd., personal communication, 2004).

Early field observations (Wilson 1956; Hawley 1962; Naldrett et al. 1972) reported the dip of crystal layering in the norite to be flatter than the dip of the basal contact. Hawley (1962) interpreted the layering as a type formed by crystal settling and concluded that the SIC was emplaced in its present form. In the South Range, igneous laminations in the norite dip less steeply than the adjacent footwall, suggesting that the SIC resembles a funnel. In the North Range, phase layering is parallel to the footwall contact, indicating that there the SIC is sill-like (Naldrett et al. 1970). In order to reconcile these observations, Naldrett and Hewins (1984) modelled the SIC as a funnel-shaped body at depth (uplifted South Range) which passed upward and outward into a sill (North Range).

Detailed petrofabric studies of the SIC, using modern techniques, are summarized by Cowan et al. (1999). Foliations and lineations, in gabbro and norite, were determined based on the orientation of tabular and acicular plagioclase crystals, respectively. Of the nine oriented specimens, two were from the same site in the North Range, four specimens from the South Range are less than four km apart, while the three specimens from the East Range are widely distributed. The foliation in all nine specimens dips inward, toward the basin centre. The foliation dip of eight of the specimens is less than the dip of the nearby base of the SIC by the following: North Range, 5°; South Range, 30°, 30°, 20°, 10°; East Range, 60°, 25°, 15°. Lineations, from seven of the eight lineated specimens, plunge down-dip toward the basin centre. Cowan et al. (1999) interpreted their data as follows: 1) the foliation was not the result of the noncylindrical folding of a semi-consolidated melt sheet, as this mechanism would produce S-planes which would dip steeper than the basal contact; 2) a ponded melt sheet would develop a foliated mineral fabric devoid of lineations; and 3) the shallow dip of the foliation, relative to the base, and the down-plunging lineation are best explained by gravitational settling of magmatic plagioclase on inclined magma-chamber walls.

By means of the anisotropy of magnetic susceptibility, supplemented by digital image analysis, Cowan et al. (1999) identified a plagioclase lineation in the granophyre at eight sites in the North Lobe. The lineation is orthogonal, or nearly so, to the base of the SIC at all sites. The orthogonal crystal growth is attributed to a temperature gradient and preferred crystal nucleation in a magma chamber. Apparently, even minor tectonic shortening will destroy the orthogonality. Cowan et al. (1999) conclude that the mineral fabric and the low overall strain in the North Lobe preclude a fold origin for the lobe.

South Range Shear Zone

A “sheared zone” approximately 64 km long and from 1 to 5 km wide (Figure 9.1), recognized by Burrows and Rickaby (1930) and named South Range Shear Zone (SRSZ) by Shanks and Schwerdtner (1991), represents a northeast-trending, southeast-dipping zone of reverse shear. According to the former authors, rocks of the Onaping Formation in the shear zone have a pronounced cataclastic structure. Thomson (1957) noted that shearing was particularly intense near faults, and interpreted the shearing pattern to be due to northwest-directed thrusting and folding prior to the emplacement of the SIC. Rousell (1975) recognized a penetrative tectonic foliation and lineation, as defined by flattened and elongate fragments in Onaping rocks of the South and East ranges (Photo 9.4) and ascribed the structure to flattening rather than shearing.

Based on structural studies in the central and southwest parts of the South Range, Shanks and Schwerdtner (1991) defined the SRSZ as a band of rock containing both an L-S fabric and a shear-plane fabric. Cowan (1996) studied the SRSZ in the eastern part of the South Range. The northern boundary of the SRSZ is defined by the Fairbank Lake Fault (Ames, Davidson et al. 2005). North of the fault, the rocks contain an L-S fabric, lack a shear-plane fabric, and are considered to lie within a field of folding strain. Stevenson (1961) recorded folds in large quartzite fragments, at the base of the Onaping Formation, which lie within the SRSZ. However, the timing of the folding is not clear, as the quartzite may have been folded prior to brecciation and incorporation into the Onaping Formation. Shanks and Schwerdtner (1991) calculated a minimum net displacement across the shear zone of 8 km, with a corresponding horizontal shortening of nearly 6 km, assuming heterogeneous simple shear in a northwest-southeast vertical plane. The suggested displacement may be excessive, as cross-sections, modelling the 8 km displacement, place the top of the granophyre approximately 4 km northwest of the actual location

(Rousell 2006, illustrated on accompanying poster). Possibly the SRSZ comprises a family of shears. Shanks and Schwerdtner (1991) noted that in the competent norite in the southwest rim, the zone consists of a series of ductile shear zones and variations in strain gradients in northwest-southeast vertical sections are incompatible with heterogeneous simple shear.

Photo 9.4. Tectonically elongated quartzite fragments and near-vertical foliation in the basal Onaping breccia (3 km southwest of the Sudbury Airport).

The Lindsley Mine is located on the South Range (Figure 1.1). The ore zones occur at the contact between the base of the SIC and the Footwall. The contact dips 45° NW from the surface to a depth of 1200 m, below which the contact reverses to a dip of 45° SE (Binney et al. 1994). The ore zones below 1200 m are deformed by a southeast-dipping zone of reverse shear with an apparent radiometric age of approximately 1658 Ma (see Chapter 10). The attitude, slip direction and metamorphic grade of the Lindsley shear zone are similar to that of the regional South Range Shear Zone (Bailey et al. 2004).

Fractures

FAULTS

A fault "set" is a collection of faults with the same orientation, direction and sense of slip. They presumably formed in the same stress regime (Angelier 1994). Faults in the Sudbury area cannot be unequivocally classified in terms of sets because of uncertainties as to the direction and sense of slip. Accordingly, they are classified in terms of five major groups whereby faults in a given group have approximately the same strike and geographic location (Rousell et al. 2002). The traces of the major faults in the Sudbury area are plotted on Figure 9.1. The average strike of the groups are shown in Figure 9.10. The groups are Murray, Vermilion Lake, Errington, Fecunis Lake and Norduna.

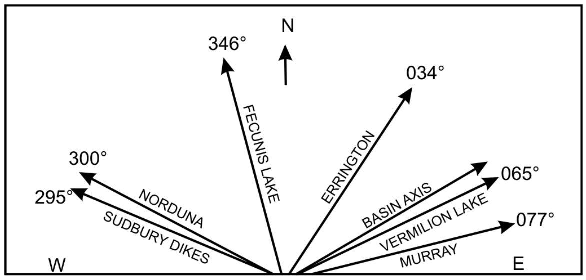


Figure 9.10. Diagram showing the average strike of the five major fault groups of the Sudbury area, as well as the average strike of dikes of the Sudbury swarm. The direction of the long axis of the Sudbury Basin is also indicated. See Figure 9.1 for the location of the faults.

The faults of the Murray Fault Group are the Murray Fault (dip 70°S), the Creighton Fault (dip 70° to 90°N), the Espanola Fault, the Garson Fault and the Falconbridge Fault. Supposedly, the Murray and Creighton faults have had a long history of activity including: normal growth faulting (south-side down) associated with deposition of the Huronian Supergroup; reverse faulting (south-side up) during the Penokean Orogeny; and Neoproterozoic dextral oblique-slip, south-side up (Card 1978a; Sims et al. 1980; Cochrane 1991). The steep north dip of the Creighton Fault (Cochrane 1983) intimates it was neither a growth fault nor a north-verging reverse fault. The dip of the Garson Fault is 70°N to vertical on the surface and 70°S at the 3600 foot level of the Garson Mine (Cochrane 1991). The latest movement on the Garson Fault is normal-sinistral oblique-slip (pitch of net slip is 8°E) as indicated on the 3600 foot level of the Garson Mine. The sinistral slip is opposite the late dextral slip on the other faults of the Murray Fault Group. Ore deposits of the Falconbridge and East mines occur within fault zones that are located at the contact between the base of the SIC and the footwall (see Falconbridge Fault in Figure 9.1). The pattern is complex and the reader is referred to Lochhead (1955) and Owen and Coats (1984) for details.

Faults of the Vermilion Lake Group displace the base of the SIC at the southwest corner of the Sudbury Basin. The four major faults of the group are, from north to south, Cameron Creek Fault, Vermilion Lake Fault, Fairbank Lake Fault (Thomson 1957) and Chicago Fault. Note that in a later compilation (Dressler 1984c), the Vermilion Lake Fault is shown as a splay of the Cameron Creek Fault with the Vermilion Lake Fault of Thomson (1957) left unnamed. A similar format is followed by the most recent compilation (Ames, Davidson et al. 2005) except that the Cameron Creek Fault is also left unnamed. The faults of the group are steeply dipping reverse faults, south-side up. Although equated with the Murray Fault Group by some investigators (Card and Hutchinson 1972), they are classified separately because of their location. The proposed traces of the Vermilion Lake Group, in the central part of the basin, are along drift-covered valleys so that the connection with the Norduna Group (Thomson 1957) is problematic (Rousell 1975, 1984a). The faults in the Vermilion and Errington mines are considered to occur within the Vermilion Lake Fault zone (Thomson 1957).

An attempt is made to determine the displacement on the top of the SIC by the Fairbank Lake Fault (FALF). Note the displacement of a block between the FALF and the Chicago Fault (Dressler 1984c; Ames, Davidson et al. 2005). The top of the SIC has a scalloped appearance in the footwall of the FALF and, to a lesser extent, in the hanging wall as well. This map pattern suggests northeast-plunging folds. Data are as follows: attitude of FALF = 063° 60°SE (dip from Thomson 1957); attitude of the top of the SIC as measured in the footwall = 302° 41°NE; sinistral strike separation = 4900 m; and pitch of the top of the SIC in the plane of the FALF is 34°E. Note that the estimated dip of the top of the SIC is based on the plunge of folds at the South Fold Nose (53°E) and the mean plunge of folds in the Vermilion and Errington mines (30°E). The approximate values are as follows: dip-slip = 8000 m; heave = 4000 m; and throw = 6928 m. The dip-slip corresponds exactly with the estimated net minimum displacement across the South Range Shear Zone (Shanks and Schwerdtner 1991). In stress fields in which two of the principal stresses are horizontal and one vertical, normal faults dip 60° and reverse faults dip 30°. Accordingly, the principal stresses may have been non-horizontal, or the FALF may have undergone later rotation or it may represent a reactivated normal fault.

Faults of the Errington Group strike approximately 025° and are diagonal with respect to the long axis of the basin. Except for a single fault in the East Range, they are confined to rocks of the Whitewater Group where they locally offset fold axes in the Chelmsford Formation in a dextral sense. The Errington Fault may represent a southeast-dipping reverse fault (Paakki 1992) and, unlike the other faults of the group, extends southwest so as to cut the granophyre. The Errington Fault may, in fact, be atypical of the group.

Rocks of the North Range are cut by northwest-striking faults of the Fecunis Lake Fault Group. The faults extend as far north as Timmins, a distance of over 200 km (Cochrane 1991), yet do not cross the central part of the Sudbury Basin. North-northwest-striking faults in the East Range may be part of the group but their strike length is limited. The Fecunis Lake Fault dips 67°W (as determined from chart B of Coats and Snajdr 1984) and has a sinistral strike separation of 800 m on the base of the SIC. Based on calculations by Cochrane (1991), where net slip is 490 m and pitch of the net slip is 26° N, the fault is classified as a reverse-sinistral oblique-slip fault.

Faults of the Norduna Group strike northwest and parallel to the bulge at the junction between the East and South ranges. The faults have a variety of dip angles, dip direction and sense of displacement. The Norduna Fault shows a sinistral strike separation on the top of the norite of approximately 1 km; although the dip is unknown it is interpreted as a reverse fault, south-side up (Dressler et al. 1991). The Airport Fault, dip 70° NE, has a sinistral displacement sense. The Bailey Corners Fault dips 65°SW (Dressler et al. 1991).

Faults in the Sudbury area appear to have been activated or reactivated during several episodes of brittle deformation that spanned a lengthy time period. The orientation of the principal stress axes must have varied considerably. An understanding of the orientation, amount and sense of displacement of the faults is of practical importance. For example, the *in situ* stress field is sufficiently large whereby mining activity can trigger slip along underground faults (e.g., Creighton Mine). Thus, knowledge of the faults is important for ground-control predictions. Moreover, some faults displace orebodies (Creighton Mine, Fecunis Mine) while others are the loci of Ni-Cu-PGE mineralization (e.g., Garson Mine, Falconbridge Mine).

JOINTS

A joint is an outcrop-scale extension fracture or shear fracture with little or no visible displacement on the shear fracture. Rousell and Everitt (1981) measured joints in rocks of the Whitewater Group and the SIC at 456 stations. (See Rousell (1984a) for a summary of the results.) Plumose structure and slickensides were seldom observed, which prevented a genetic classification of the joints. There are as many as seven joint sets at a given station and sets at virtually all orientations are present. Figure 9.11 represents a summary plot of the major joint sets in nine subareas and indicates that sets with northwest strikes and steep dips predominate. Rocks of the East Range are the most intensely jointed with the sets geometrically related to a local fault, the strike of rock units, diabase dikes or the norite-footwall contact. Mitchell and Mutch (1957) recorded three well-developed joint sets in the norite of the Hardy Mine (subarea 4): parallel to the base of the norite; parallel to the strike and normal to the dip of the same contact; and vertical and striking normal to the contact. The same sets occur at the surface in the North Range (Photo 9.5). The set that is parallel to the base of the norite is too local to plot on Figure 9.11.

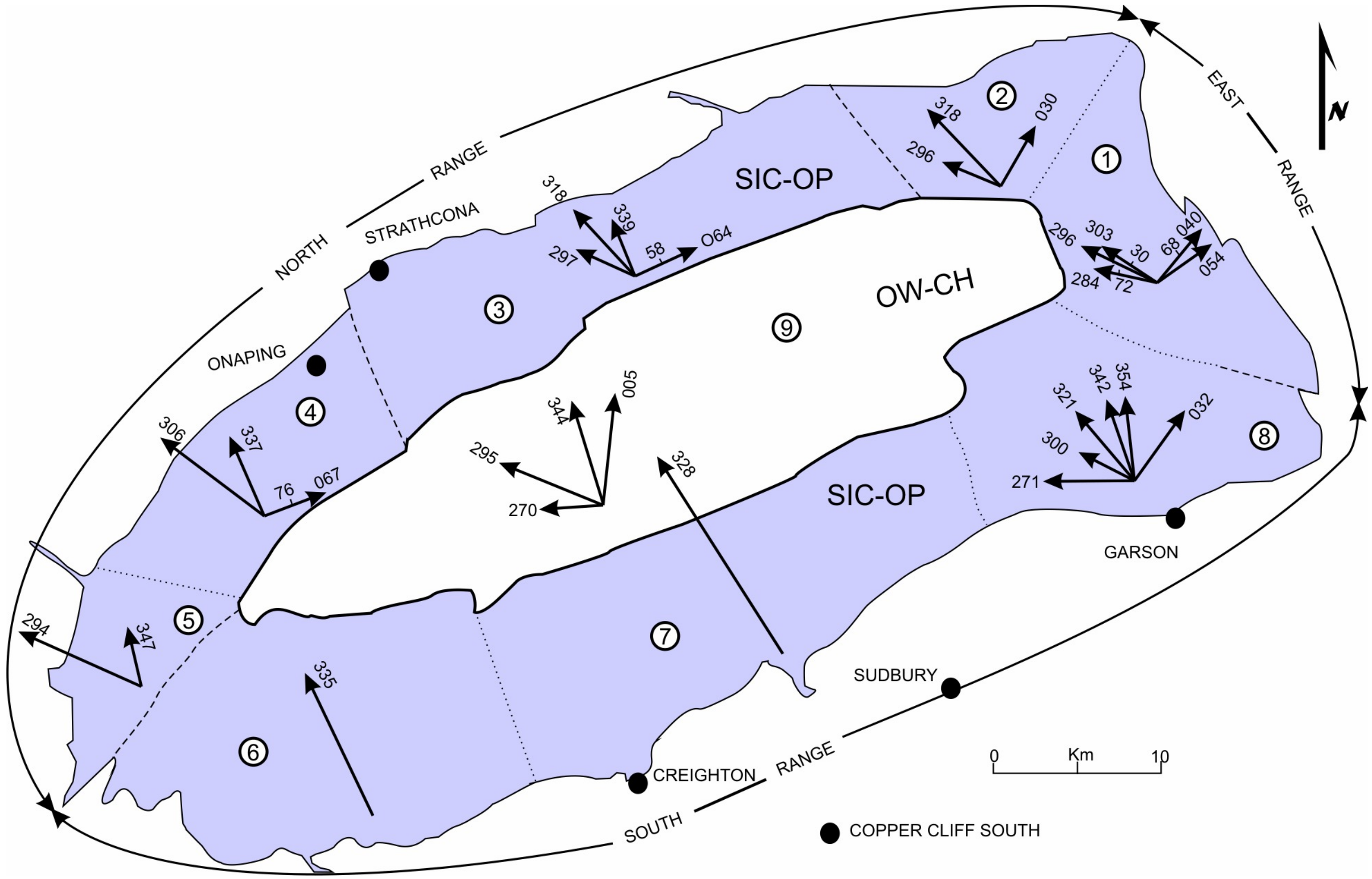


Figure 9.11. Map showing the average orientation of major joint sets inside the Sudbury Basin (*from* Rousell 1984a). The length of strike lines (arrows) are proportional to the strength of maxima on contour diagrams (Rousell and Everitt 1981). Dips are vertical unless otherwise indicated. Subareas are shown as numbers enclosed by circles. SIC-OP = Sudbury Igneous Complex and Onaping Formation, OW-CH = Onwatin and Chelmsford formations.

Other sets in the North Range strike parallel to the strike of faults of the Fecunis Lake Group or to diabase dikes of the Sudbury swarm. In the South Range, the dominant joint set dips vertical and strikes normal to the strike of the tectonic foliation. Of the four joint sets in subarea nine, one set is parallel to diabase dikes, another set represents cross joints on folds, while the other two sets cannot be related to structural features.

A detailed study of joints in the former City of Sudbury (N = 1644), located in the footwall of the South Range, indicates that joints have steep dips with dominant strike directions of 323° and 037° (Rousell et al. 1998). Dressler (1984b) measured joints in 22 subareas in the footwall of the North Range. A regional trend of northwest or west-northwest was identified and is apparently most evident in subareas furthest from the SIC. Many subareas display a random distribution of joint directions. Locally, there is a preferred joint orientation parallel to the basal contact of the SIC.

Photo 9.5. Three joint sets in the norite (SIC) of the North Range. The man is standing on a set which is parallel to the footwall contact of the SIC, his hand rests on a vertical set that strike normal to this contact and the third set, in shadow behind the man, strikes parallel and dips normal to the contact.

LATE-STAGE FRACTURES

Cochrane (1991) identified a system of late-stage faults and joints (called fractures) that apparently constitute the youngest structural features in the Sudbury area. The late-stage faults consist of steeply dipping, north-northeast- to east-striking strike-slip faults and north-northwest-striking thrust faults. Displacement varies from centimetres to over 100 metres. Steeply dipping late-stage joints strike from west-northwest to north-northeast (Figure 9.12). The joints occur in *en echelon* patterns that form distinct zones and may be filled with calcite and/or sulphide minerals. Apparently some "joints" have displacements of several metres, in which case they may be classified as shear fractures. Note that seismic activity, not directly related to mining, correlates with late-stage fractures.

The approximate orientation of the three principal stresses ($\sigma_1 > \sigma_2 > \sigma_3$) can be determined from the orientation of the faults. Thus, for strike-slip faults σ_1 trends 067.5°, σ_2 is vertical and σ_3 is horizontal, whereas for thrust faults σ_1 trends 056°, with σ_2 horizontal and σ_3 vertical. Accordingly, the trend of σ_1 is approximately east-northeast (Figure 9.12). The late-stage joints, with their wide strike range (90°), cannot be easily related to the proposed stress field. Some may be incipient reverse faults but it is unlikely they represent extension fractures.

The orientation of the principal stresses (overcoring method) from five mines in the area (Cochrane 1991) are plotted on the stereogram of Figure 9.13. The mean orientation of σ_1 (Ramsay 1967), obtained from all plots except those from Strathcona Mine (M∂ on stereogram), is, in terms of plunge and trend, 9° W 252°. This value of σ_1 is in close agreement with that obtained from the late-stage faults (Figure 9.12). The mean value of σ_1 (Mb on the stereogram) from the Strathcona Mine (plots 4, 5, 6, Figure 9.13) is 13° E 124°. The intermediate and minimum principal stresses show considerable variation but σ_2 tends to be subhorizontal and σ_3 subvertical, which corresponds to the stress field for reverse faults.

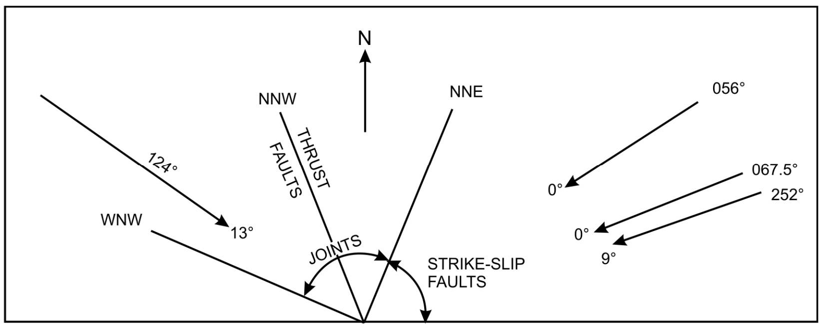


Figure 9.12. Sketch showing the orientation of late-stage faults and joints and the approximate orientations of the maximum principal stress, in terms of plunge and trend. Stress orientations are as follows: 1) 0° 056° (normal to the strike of thrust faults); 2) 0° 067.5° (bisects the strike range of strike-slip faults); 3) 9° SW 252° (average of data from the Onaping, Creighton, Copper Cliff and Garson mines, see Figure 9.13); and 4) 13° SE 124° (average of data from the Strathcona Mine, see Figure 9.13). *Data from* Cochrane (1991).

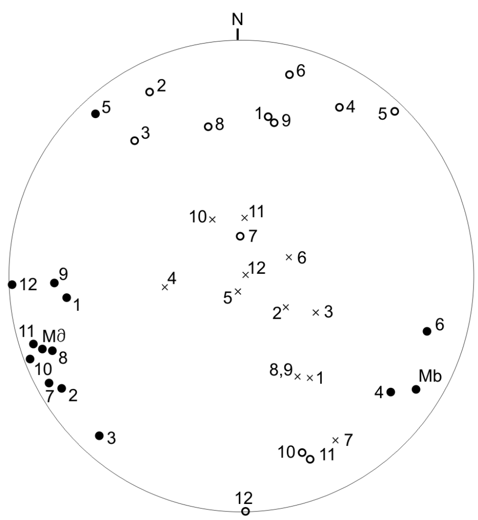


Figure 9.13. Lower hemisphere equal-area plot of the maximum (closed circle), intermediate (open circle) and minimum (cross) principal stresses from the following mines: Onaping = 1, 2, 3; Strathcona = 4, 5, 6; Creighton = 7, 8, 9; Copper Cliff South = 10, 11; and Garson = 12. M∂ = mean value of σ_1 from Onaping, Creighton, Copper Cliff South and Garson mines and Mb is the mean value from Strathcona Mine. For the location of the mines see Figure 9.11. *Data from* Cochrane (1991, Table 6-1). Mean values calculated by the method of Ramsay (1967).

Conclusions

Prior to the Sudbury event (1850 Ma) the Paleoproterozoic rocks of the Southern Province, including outliers of the Huronian Supergroup northwest of the Sudbury Basin, were folded about east- to northeast-trending axes and locally about north-northeast- to north-northwest-trending axes. This folding may have occurred during, or immediately prior to, the emplacement of the Nipissing Gabbro at approximately 2219 Ma (Card et al. 1972) and may be related to the Blezardian Orogeny (2140 Ma, Stockwell 1982; revised age 2400 to 2200, Riller and Schwerdtner 1997) or possibly to an early phase of the Penokean Orogeny (1900 to 1700 Ma, Bennett et al. 1991).

After the Sudbury Event, ductile deformation of the central and southern parts of the Sudbury Basin, by northwest-directed stresses, resulted in shortening and uplift. The basin was also weakly deformed by stresses oriented in a northeast-southwest direction. This resulted in the following: folding about northeast-trending axes and more widely spaced northwest-trending axes formed northeast-striking doubly plunging folds; the development of a prominent S-plane (fracture cleavage, slaty cleavage, foliation); the formation of a largely down-dip stretching lineation; and the development of South Range Shear Zone and satellite shears.

The post-Sudbury Event may be related to a late stage of the Penokean Orogeny. However, apart from the age of shearing at the Lindsley Mine (roughly 1658 Ma), the timing of post-1850 Ma ductile deformation is poorly constrained. The post-Sudbury Event deformation was largely the result of northwest-directed principal stresses. In contrast, the orientation of the maximum principal stress related to late faults as well as to fractures produced by overcoring (*in situ* stress field) is west-southwest–east-northeast.

Representatives of all five fault groups, discussed above, cut the SIC and hence were active after the Sudbury Event. The time of original formation is unknown. Faults of the Vermilion Lake Group are steeply dipping reverse faults, whereas those of the other groups represent strike-slip faults, many of which have an oblique-slip component.

For further discussions of the tectonic history of the area see, for example, Chapter 10 (this guide), Riller et al. (1999) and Rousell et al. (1997, 2002) and references therein.

Chapter 10

Metamorphism and Metasomatism

R.S. James and J.P. Golightly

Introduction

The Sudbury area has been affected by several episodes of metamorphism that include regional, contact and shock metamorphism. The effects of an earlier event may be partially obliterated by later events. High to low crustal levels are exposed with metamorphic grades ranging from subgreenschist to granulite facies. Gold mineralization is associated with regional alkali metasomatism and base metal mineralization by hydrothermal activity inside the Sudbury Basin.

Superior Province

The Superior Province, located north of the Sudbury Basin, consists of the Levack Gneiss Complex (LGC) and the Cartier granite (see Figures 1.1 and 1.2). The LGC occurs as a 0.5 to 5 km wide arcuate belt that fringes the Sudbury Igneous Complex (SIC) in the North and East ranges (Figure 10.1). The Cartier granite (2642 Ma; Meldrum et al. 1997) lies north of, and intrudes, the LGC. A mass of metagabbro located north of Joe Lake (Figure 10.1) is probably a member of the early Paleoproterozoic (2470-2490 Ma) East Bull Lake suite of intrusions.

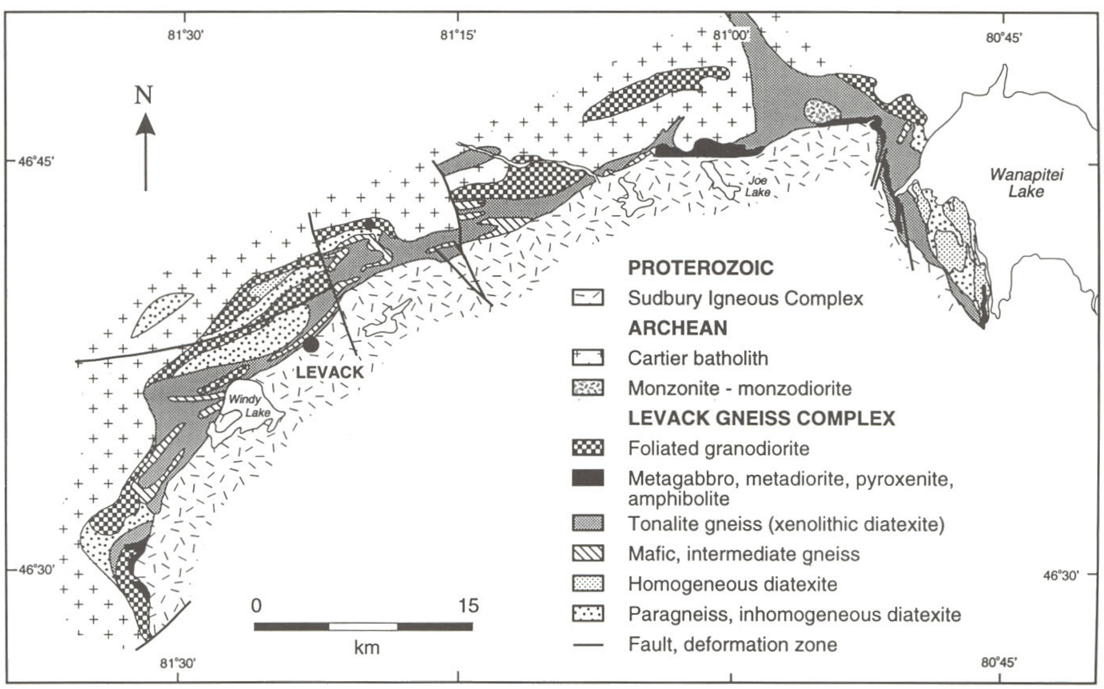


Figure 10.1. Major rock units, Levack Gneiss Complex (*from* Card 1994).

REGIONAL TECTONOMETAMORPHISM

The LGC comprises several units (Figure 10.1), including high-grade gneissic and foliated supracrustal rocks interspersed with metatexites and diatexites. Rock compositions suggest that the terrain originally represented a supracrustal sedimentary sequence that was intruded by bodies of silicic and basic magma. Minerals formed by regional metamorphism include plagioclase, quartz, biotite, hornblende, ortho- and clinopyroxene and almandine garnet. Table 10.1 sets out a summary of the mineral assemblages in the major rock units (see Kretz 1983 for an explanation of abbreviations.) Note the lack of pelitic assemblages.

Table 10.1. Mineral assemblages for major rock units in the Levack Gneiss Complex (*after* Card 1994).

Tonalite gneiss and related diatexite	pl-qz-opx-cpx-hbl-bt-minor grt; granoblastic to foliated texture.
Intermediate and mafic gneiss	pl-hbl-opx-cpx-bt-qz-scarce grt; massive to weakly foliated layered rocks; migmatites have several generations of leucosome.
Paragneiss and related diatexite	pl-qz-bt –cpx ± gt; tonalite to granite leucosomes; scarce quartz-magnetite, metamorphosed iron formation.
Metagabbro	pl-cpx-hbl; massive to weakly foliated; occurs as metre-scale boudins and metre- to kilometre-scale bodies.
Foliated granodiorite	pl-qz ± kfs, amphibole

The above mineral assemblages formed during several tectonometamorphic events as illustrated by a structural map of a local area in the footwall of the North Range (Figure 10.2). There, a penetrative S_1 gneissic foliation, at granulite facies, is folded by east-trending, shallow-plunging F_2 folds which are refolded by open to tight northwest-plunging F_3 folds that have a weak axial-planar cleavage. Granulite facies metamorphism occurred between 2711 Ma (Krogh et al. 1984) and 2648 Ma (Ames et al. 1997), at a depth of 21 to 28 km, and at 750 to 800°C (James et al. 1992). This was followed by amphibolite facies metamorphism, migmatization and F_2 foliation formation at 2647 Ma (Krogh et al. 1984), at 6 to 11 km depth and 525 to 600°C (James et al. 1992). The significant uplift between the two events may be related to the emplacement of the Cartier granite, dated at 2642 Ma (Meldrum et al. 1997). The F_3 cleavage may be related to a post-SIC episode of greenschist metamorphism that has overprinted the LGC and Cartier granite. Plagioclase-phyric dikes, presumably part of the Matachewan swarm (2475 Ma; Heaman 1995) and Sudbury Breccia (1850 Ma) cut all structures and indicate a late Archean to early Paleoproterozoic age for the regional tectonometamorphism.

SHOCK/CONTACT METAMORPHISM

A 0 to 3 km wide zone of highly shocked and thermally recrystallized rocks occurs at the northwest margin of the SIC. The zone is attributed to bolide impact and the heat of recrystallization of the resulting melt sheet (SIC) at 1850 Ma. The aureole is recorded in the LGC as well as in associated intrusive bodies such as the Cartier granite and gabbro-anorthosite intrusions. Locally, the rocks are bleached and display a spotted hornfelsic texture.

Shock features in the rocks of Levack Township, which survived contact metamorphism, extend as much as 6 km into the footwall from the margin of the SIC (Figure 10.3). They include: planar features in quartz and plagioclase; kink bands in biotite; and shatter cones (one locality only). Note that kink bands are common tectonic structures. Textural and optical data also suggest the formation of maskelynite (shock-induced amorphous plagioclase). Shock pressures as high as 7 Gpa (Figure 10.3) were calculated by the method of Robertson and Grieve (1977).

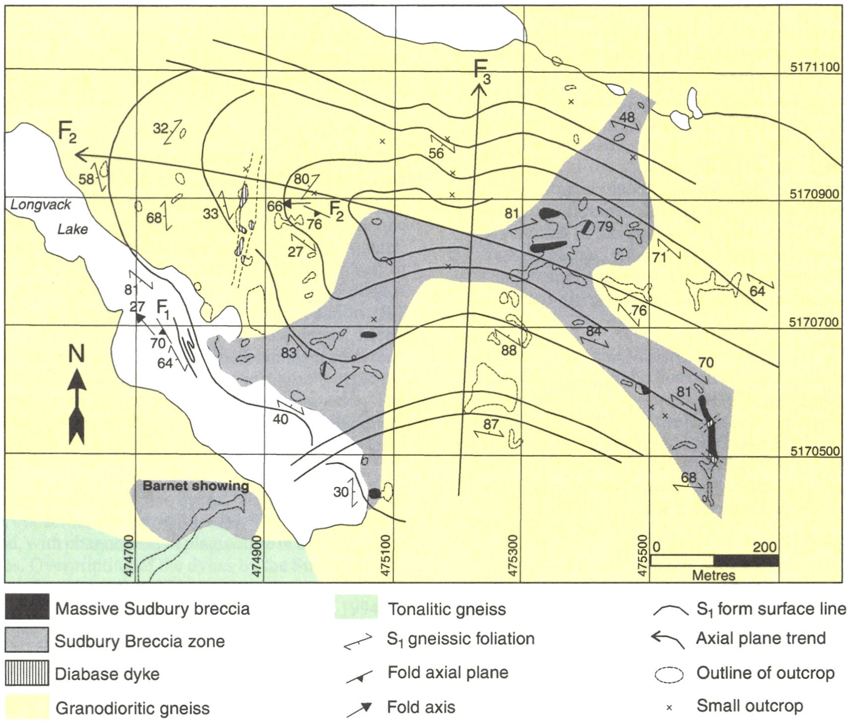


Figure 10.2. Geology of Archean gneisses and Sudbury Breccia zone, east of Longvack Lake (*from* Legault et al. 2003).

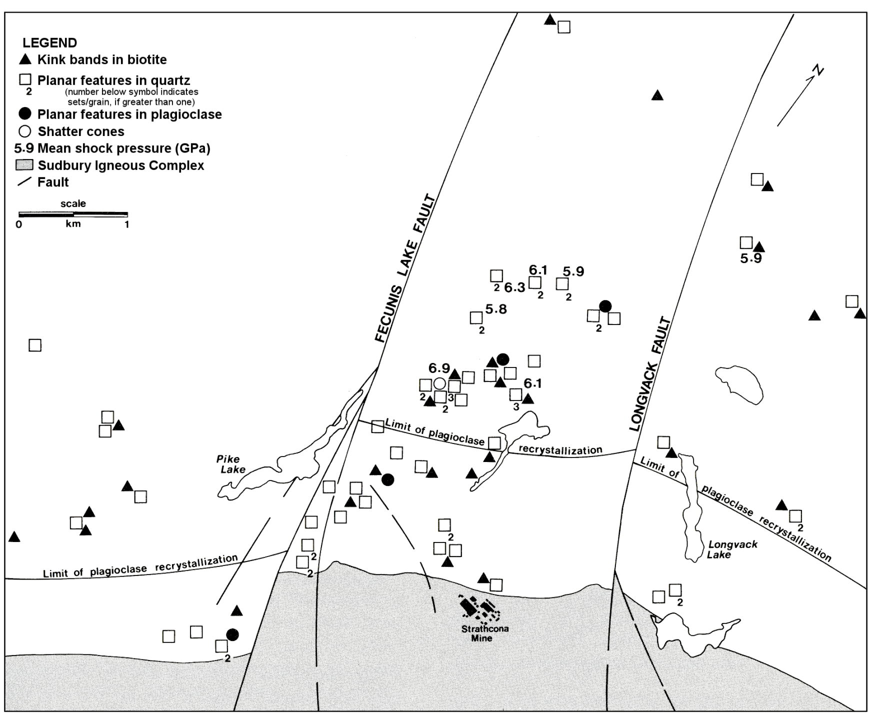


Figure 10.3. Zoning of shock metamorphic features in the footwall rocks of Levack Township (*from* Dressler 1984b).

Three to four contact metamorphic zones have been identified (Boast and Spray 2006; Dressler 1984b). The variable width of the aureole is due to post-SIC, northwest-directed thrust faulting (Boast and Spray 2006). Dressler (1984b) mapped the distribution of contact metamorphic facies in Levack Township (Figure 10.4). Boast and Spray (2006) reported similar data in the Windy Lake area but with the addition of an anatectite facies. The facies are as follows.

1. Anatectite facies (Footwall Breccia): width ≤ 25 m; annealed and/or retrogressed mafic clasts in a felsic igneous matrix; pl, ± ap, mag, px in gabbro-anorthosite areas (Boast and Spray 2006).

2. Pyroxene hornfels facies: width ≤ 180 m; cpx + pl ± brown and green hbl, bt, qz (Dressler 1984b); pl + qz + opx ± hbl, bt (Card 1984); opx + cpx + pl (An_{30-70}) ± mag, ap; myrmekite, rare crd + opx assemblages (Boast and Spray 2006).

3. Hornblende hornfels facies: width ≤ 900 m; pl + hbl + qz + bt (Card 1994; Dressler 1984b); cpx + hbl + pl (An_{30-70}) ± qz (Boast and Spray 2006).

4. Albite-epidote hornfels facies: width ≤ 1 km; pl (An_{10-50}) + qz ± bt, ep, chl; pl (An_{30-50}) ± qz, ep, chl.

Idioblastic, stubby, decussate plagioclase crystals completely replace original plagioclase and/or maskelynite in all but the lowest grade facies, and increase in grain size with increasing metamorphic grade. Large, strained quartz crystals are replaced by clear, unstrained polygonal aggregates throughout all four contact facies. Planar features in plagioclase and quartz are scarce or absent as a result of the thermal recrystallization. Metamorphic pyroxene, hornblende and biotite form clear, unstrained crystals. Clinopyroxene forms characteristic granoblastic aggregates which replace strained, shocked clinopyroxene, particularly in the pyroxene hornfels zone. Kink-banded biotite and planar-shocked features in quartz are common outside the contact aureole. Granophyre melt-patches at the SIC contact (anatectite facies) have refractory remnants with assemblages such as cordierite-orthopyroxene-biotite-sillimanite-hercynite-corundum. This assemblage suggests temperatures of 900°C and a depth of 5 to 11 km (James et al. 1992).

Rousell et al. (2003) studied the effects of contact metamorphism on the matrix of Sudbury Breccia in an area near Levack Mine. In an inner zone 60 to 140 m wide (pyroxene hornfels facies), a mosaic of twinned plagioclase grains have an average diameter of 0.1 mm and a maximum size of 0.5 mm, whereas in the outer zone (albite-epidote hornfels facies) the average diameter is only 0.03 mm with the maximum at 0.1 mm.

Southern Province

The Paleoproterozoic Southern Province is bounded by the Archean Superior Province to the northwest and the Mesoproterozoic Grenville Province to the southeast (Figure 1.1). The Province consists of northeast-striking supracrustal rocks of the Huronian Supergroup (>2450 to 2200 Ma; Bennett et al. 1991), Nipissing gabbro (2210 Ma; Noble and Lightfoot 1992), SIC (1850 Ma; Krogh et al. 1984) and the Whitewater Group (post 1850 Ma) inside the Sudbury Basin. Fresh olivine diabase dikes of the Sudbury swarm (1238 Ma; Krogh et al. 1987) represent a minimum age for the rocks and for metamorphic events.

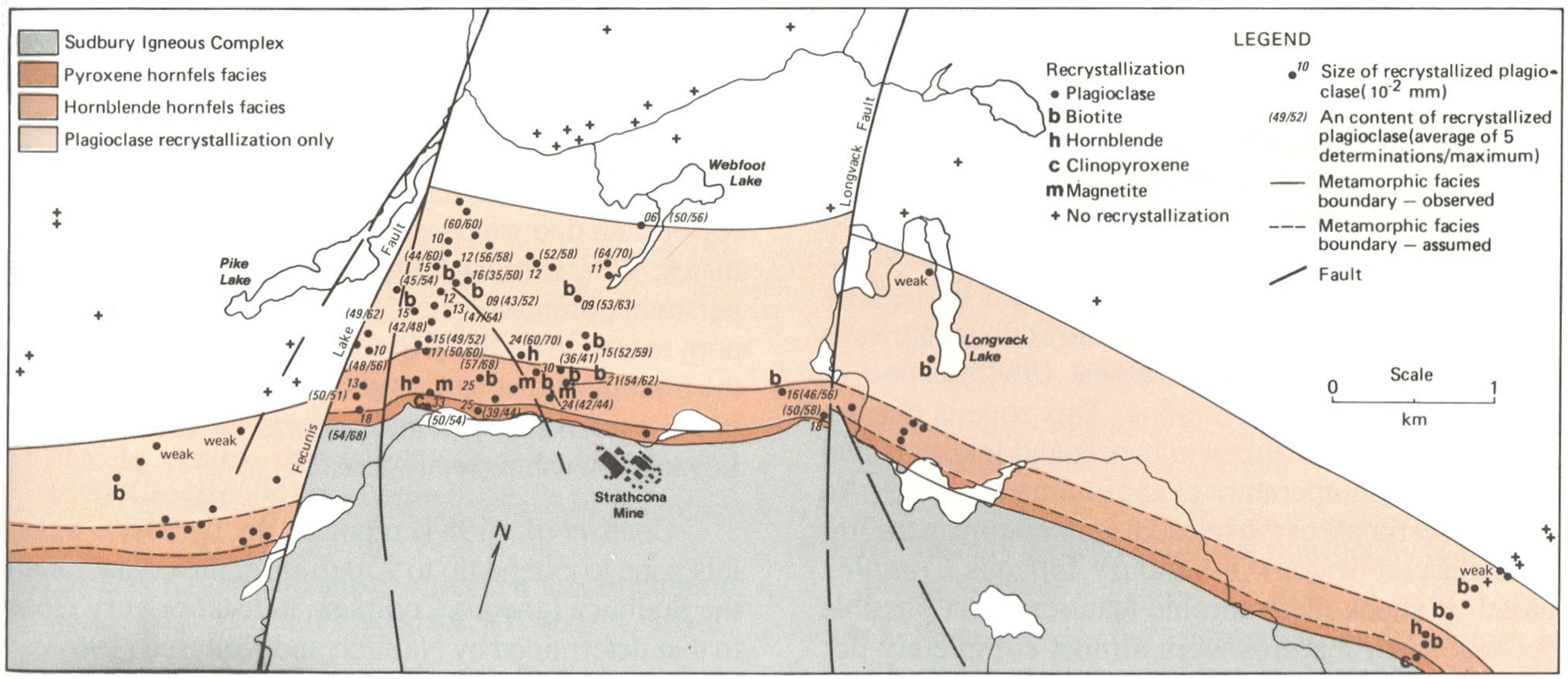


Figure 10.4. Contact metamorphism, Levack Township (*from* Dressler 1984b).

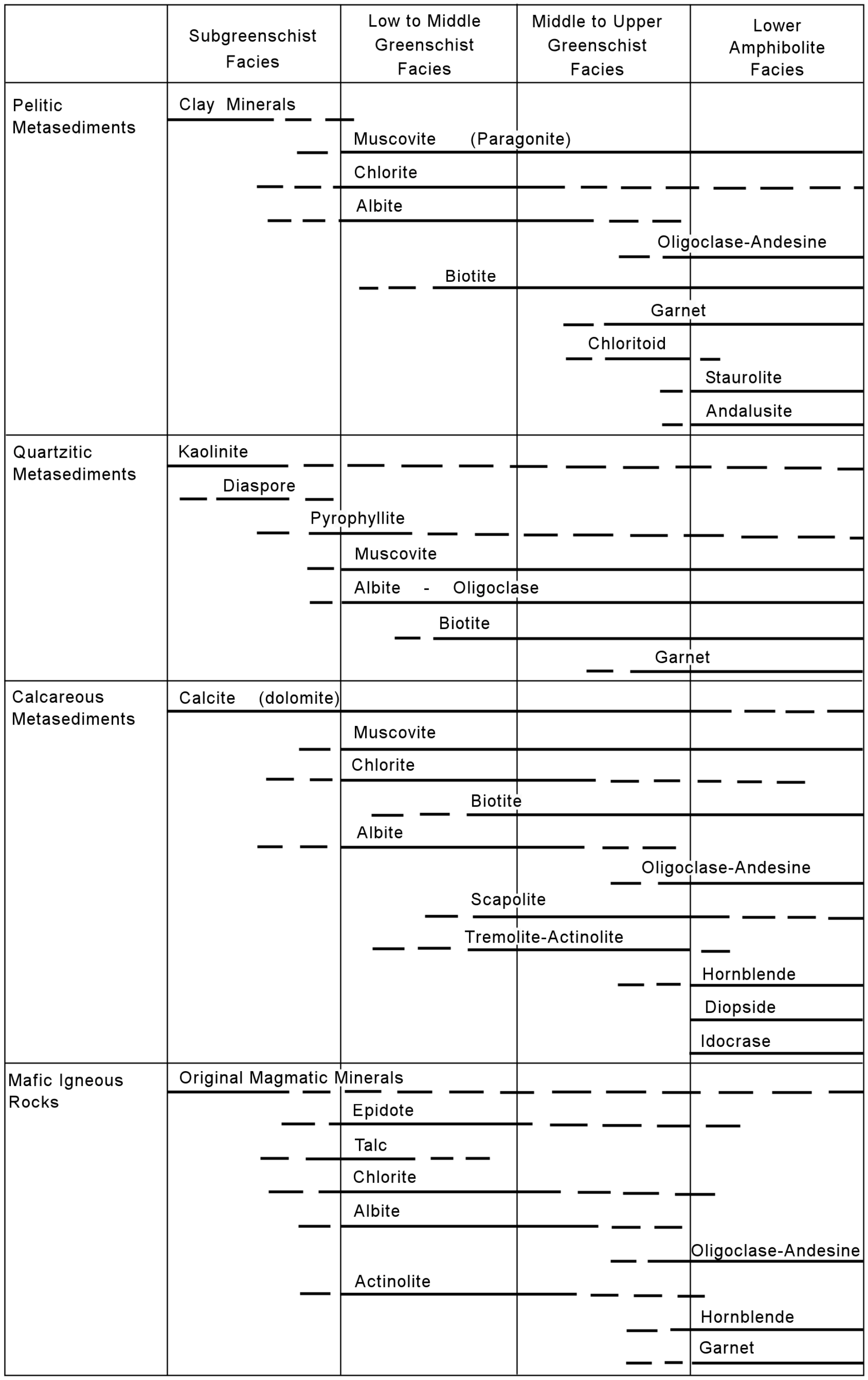
The precise timing of tectonometamorphic events in the Southern Province remains unresolved. Much of the deformation is attributed to the Penokean Orogeny. However, there are several estimates as to its range, such as 1900 to 1700 Ma (Bennett et al. 1991) or 2400 to 2200 Ma (Williams et al. 1992). Rousell et al. (1997) concluded that tectonometamorphic events occurred before and after the Sudbury Event (1850 Ma), with the former event more aerially extensive than the latter.

Patterns of metamorphism of the eastern part of the Southern Province, from Sault Ste. Marie to Cobalt, are shown in Figure 10.5 (Card 1978a). The reader is referred to the following material for local details: Fox (1971), Brocum and Dalziel (1974), Zolnai et al. (1984), Fleet et al. (1987), Shanks and Schwerdtner (1991), Riller and Schwerdtner (1997), and Jackson (2001). Easton (2000) summarized the foregoing data and incorporated it into a tectonometamorphic map of Ontario (Easton and Berman 2004).

Table 10.2 sets out the metamorphic assemblages for rock types in the eastern Southern Province. Notable mineral occurrences include the following: staurolite, chloritoid, andalusite and almandine garnet in metapelites; kyanite in quartzites; scapolite, diopside and idocrase in calcareous metasediments; and garnet, actinolite, hornblende and epidote in mafic igneous rocks. Also, very low temperature diagenetic minerals such as kaolinite, diaspore and pyrophyllite occur in quartzitic metasediments. In general, original igneous textures and structures are well preserved even in the highest grade rocks. Porphyroblasts of staurolite, chloritoid, garnet, andalusite and kyanite are locally preserved in pelitic rocks (e.g., McKim Formation). Elsewhere, these phases are replaced by chlorite or white mica. In mafic igneous rocks, including volcanic rocks of the Elsie Mountain and Stobie formations (see Figure 4.2 for units of the Huronian Supergroup), Nipissing gabbro and norite and gabbro of the SIC, one or more metamorphic events produced texturally complex assemblages of calcic and non-calcic amphiboles, chlorite and epidote-group minerals.

In the eastern Southern Province the metamorphic grade varies from subgreenschist facies to lower amphibolite facies (Figures 10.5 and 10.6). Two large domains of subgreenschist facies flank a central area (Sudbury–Cutler) of lower to mid-greenschist facies to lower amphibolite facies (Figure 10.5). The elongate nodal distribution of the amphibolite rocks in this area appears to be due to a combination of at least two periods of regional folding accompanied by extensional and later compressional movement on northeast-striking faults of the Murray Fault set. The occurrence of oligoclase in upper greenschist pelitic and mafic assemblages and the stable co-existence of staurolite and andalusite ± kyanite and rare sillimanite in pelites suggest a low pressure, intermediate type of regional metamorphism. Several authors (Fox 1971; Jackson 2001; Easton 2006) have estimated metamorphic conditions in local areas situated 45 to 60 km east-northeast of Cutler (Figure 10.5). The assemblages reported by the latter two investigators also indicate the "low pressure intermediate type" of Miyashiro (1961) or bathozone 3 of Carmichael (1978). Mineral equilibria involving garnet, biotite, plagioclase and white mica allowed Jackson (2001) to estimate the following peak metamorphic conditions: T approximately 500° to 560°C; P approximately 1.5 to 3 kb; and depth roughly 5 to 11 km. Fox (1971) employed Schreinemakers analysis to estimate the (cd, bi, ga, chl) invariant point at T = 500° to 550°C, P = 4.45 to 4.75 kb and depth = 15 to 17 km. Easton (2006), utilizing the P-T grid after Spear (1993) for the system KFMASH, estimated that the highest grade metapelites (kyanite, sillimanite) formed at bathozone 4 or 5 (Carmichael 1978) at T approximately 600° to 625°C, P approximately 4 to 6 kb, and depth approximately 14 to 21 km. This assemblage represents the deepest part of the orogenic belt recognized to date.

Table 10.2. Metamorphic minerals present in the major rock types of the eastern Southern Province at various metamorphic grades (*from* Card 1978a).



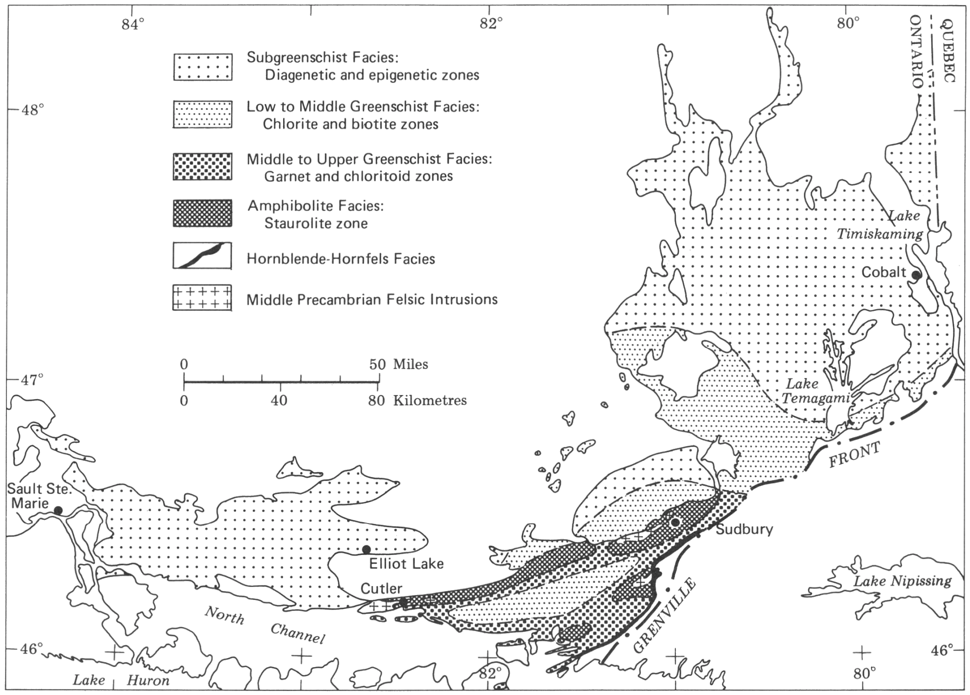


Figure 10.5. Patterns of metamorphism in the eastern part of the Southern Province (*from* Card 1978a).

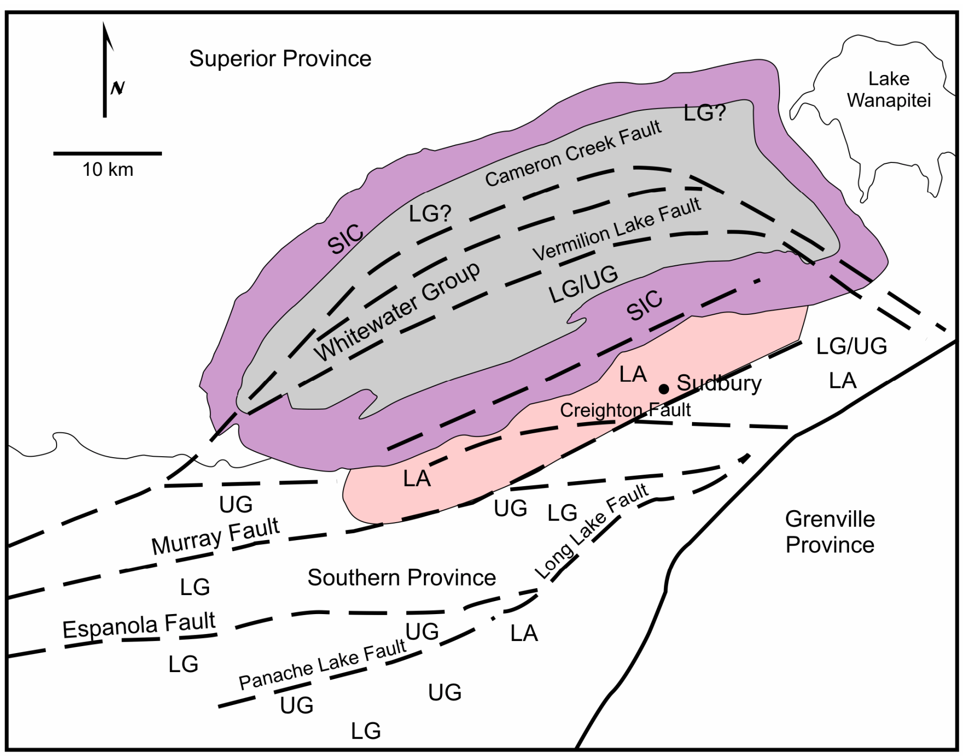


Figure 10.6. Map showing the distribution of regional metamorphic facies and major faults in the Sudbury area. LG = lower greenschist facies, UG = upper greenschist facies, LA = lower amphibolite facies.

GREATER SUDBURY AREA

An elliptical zone of amphibolite facies rocks lies along the southern margin of the SIC (Figure 10.6). There, norites of the SIC and basalts of the Elsie Mountain and Stobie formations (Huronian Supergroup) form amphibolites (hornblende + plagioclase assemblages), whereas pelites in the Stobie and McKim formations developed staurolite ± white mica ± andalusite / cordierite assemblages. Note that in the basalts, immediately adjacent to the SIC, relict outlines of pyroxene porphyroblasts / clasts, replaced by one or more generations of amphibole, attest to the presence of a relict contact aureole. In the area between the amphibolite facies zone and the Long Lake Fault (Figure 10.6), metasediments of the Hough Lake Group are within the greenschist facies, i.e., the assemblages qz, and two or more of mu, plg, kspar, bio and chl. South of the Long Lake Fault the metamorphic grade is abruptly higher.

Staurolite-andalusite assemblages occur in metapelite layers due to south-side-up thrust faulting along the Long Lake Fault (Fueton and Redmond 1992, 1997). These greenschist and lower amphibolite assemblages formed as a result of Penokean regional metamorphism (Card 1978a). The staurolite and andalusite porphyroblasts are late, syn- to post-tectonic with respect to the F_1 folding associated with this event. On the southeast side of this zone lies a narrow belt of hornblende hornfels facies rocks (Figure 10.5) that formed as a result of the emplacement of the Chief Lake Batholith (1464 Ma). Folded meta-greywackes and semi-pelites of the Whitewater Group (Sudbury Basin) and the SIC of the North Range (Figure 10.6) are at lower- to upper-greenschist facies (Card 1978a; Fleet et al. 1987). An axial planar penetrative foliation (S_2), defined by micas and chlorite, is best observed in the Onwatin slates and portions of the Chelmsford Formation.

Estimates of the age of minerals and fabrics formed during and after the Penokean Orogeny and described in this section, are as follows:

1) 1815 ± 15 Ma ($^{207}Pb/^{206}Pb$ age from the brown core of titanite grains in amphibolitized norite, Lindsley Mine, South Range, SIC; Bailey et al. 2004); and

2) 1806 ± 24 to 1803 ± 11 Ma (U-Th total electron microprobe analyses of cores of monazite, Mississagi Formation, amphibolite facies, south of the Long Lake Fault (LLF); Piercey et al. 2007).

Card (1964, 1978a) and Jackson (2001) identified a post-Penokean tectonometamorphic event in the Cutler–Sudbury region of the eastern Southern Province (Figure 10.5). In the Sudbury area, staurolite, andalusite and garnet porphyroblasts are largely replaced by post F_1 white mica ± chlorite. These rocks may or may not exhibit an S_2 foliation. In the same assemblages, south of the LLF, the alteration to white mica post-dates a crenulation cleavage (S_2) that folds S_1 and encloses staurolite/andalusite (Fueton and Redmond 1997). West of Sudbury, in May Township, this greenschist facies penetrative foliation and lineation are axial planar to folded S_1 fabrics and Penokean-age isograds. Piercey et al. (2007) have reported $^{40}Ar/^{39}Ar$ ages for S_2-type white mica and hornblende, from several localities in the Cutler–Sudbury region, which indicate ages from 1700 to 1799 Ma.

In the southern Lake Superior region, an area with a history of protracted convergence, Piercey et al. (2007) recognize three Paleoproterozoic tectonothermal events: Penokean (~1850 Ma); Yavapai (~1750 Ma); and Mazatzal (~1650 Ma). They report the following (Yavapai) ages from the Sudbury–Cutler area: 1799 to 1700 Ma ($^{40}Ar/^{39}Ar$) from S_2 white mica and hornblende; and 1770 to 1740 Ma (Pb monazite) from whole rocks. According to Piercey et al. (2007, p.139) "…… this metamorphic and magmatic (Cutler Batholith) pulse is the most significant Paleoproterozoic tectonothermal event recorded in the rocks of the (eastern) Southern Province".

The South Range Shear Zone (Figure 9.1) cuts the Onaping Formation, the SIC and Penokean / Yavapai metamorphic zones. Lower amphibolite facies metamorphism developed within the shear zone. Mineral assemblages include andesine, cummingtonite, tschermakite/paragasite (amphibolite), almandine garnet and biotite (Fleet et al. 1987). Bailey et al. (2004) suggest that the Thayer Lindsley Shear Zone, part of the South Range Shear Zone, formed during the Mazatzal Orogeny, as the colourless rims that surround the brown cores of titanium grains (described above) gave $^{207}Pb/^{206}Pb$ ages of 1658 ± 68 Ma.

A penetrative foliation and a penetrative south-plunging lineation (elongate quartz grains, stretched clasts) occur in Huronian supracrustal rocks south of the Long Lake Fault. These structures also occur in the Chief Lake Batholith (1464 Ma), but are absent in local Sudbury dikes (1265 Ma), indicating a post-Mazatzal age for the deformation.

Alkali Metasomatism

A complex regional metasomatic event, dated at 1700 Ma (Schandl et al. 1994), can be traced from Sault Ste. Marie eastward to Temagami. Veins, breccias and massive alteration zones, metres to tens of metres in width, are widespread in rocks of the Huronian Supergroup. The zones commonly trend parallel to northeast-striking fractures and are characterized by pervasive albitization that may be locally overprinted by carbonate, quartz, chlorite, biotite and ore minerals (Gates 1991; Schandl et al. 1992, 1994). The rocks are typically tan to reddish in colour due to finely disseminated hematite and/or the oxidation products of iron sulphides. Accessory minerals associated with albite alteration include monazite, rutile and minor detrital zircon. Accessory phases in the chloritized rocks are gadolinite (yttrium-beryllium-LREE-rich silicate), bastnaesite and synchysite (REE-rich fluoro-carbonates). In the Espanola area, albitization, extensive carbonate (calcite, ankerite), minor biotite and late pyrite, with hematite rims, characterize the alteration. In contrast, chlorite is the major alteration component in the Sudbury–Wanapitei area. Metals associated with the metasomatism include gold, copper, nickel, cobalt and chromium (Meyer et al. 1986, 1987). U-Pb ages of 1700 ± 2 Ma, from gold occurrences in the Sudbury area (Schandl et al. 1994), suggest an age between the Mazatzal and the Yavapai orogenies.

Shock Metamorphism

The following features in the Sudbury Structure may result exclusively from large-scale extraterrestrial impact and at the present time most of them are difficult to explain by any other known mechanism. Note that some features can also develop by endogenic processes, including pseudotachylite (faulting), biotite kink bands (deformation) and diamonds (kimberlite intrusions). All are presumed to form at approximately 1850 Ma, which is the age of the SIC melt sheet.

Megascopic features include:

1) shatter cones (Dietz 1964; Guy-Bray 1966);
2) Sudbury Breccia - pseudotachylite and clastic;
3) melt bodies; and
4) suevite in the Onaping Formation (Ames et al. 2002).

Microscopic features include:

5) biotite kink bands;
6) quartz deformation lamellae;
7) plagioclase deformation lamellae;
8) devitrified maskelynite; and
9) microdiamonds in the carbonaceous Dowling Member of the Onaping Formation (Masaitis et al. 1999).

Geochemical features are as follows:

10) PGE anomaly in the Onaping Formation (Mungall et al. 2004);
11) fullerenes and trapped primordial He in the carbon of the Onaping Formation (Becker et al. 1996; Elsila et al. 2005); and
12) crustal isotope signatures in the SIC (Faggart et al. 1985).

Of these twelve impact-related features, only the megascopic features and maskelynite can be observed in the field. Both are the product of low shock pressure and hence are the most distant features from the centre of the Sudbury Structure. The distribution of shatter cones and Sudbury Breccia are shown in Figure 7.2. For details of shatter cone distribution, orientation and strength of development see Figure 11.3B and Dressler (1984b) or Guy-Bray (1966). Shatter cones are most common in the footwall of the South Range (quartzite of the Mississagi Formation) and have apical axes which are, in general, perpendicular to the crater floor. Small cones form an assembly that are related to the axis of a master cone. Melt bodies occur in the South Range Breccia Belt and at the base of the Onaping Formation in the North Range. Of course, the largest melt body is the SIC itself. Shock metamorphic micro-textures and minerals are found only in the country rocks of the North Range beyond the contact metamorphic aureole of the SIC, and in country rock fragments in the Onaping Formation. (See Figure 10.3 for zoning of shock metamorphic features in Levack Township, North Range footwall.) Planar features in quartz and plagioclase suggest shock pressures of 6.1 to 6.9 Gpa. The outer limit of plagioclase recrystallization probably coincides with the outer limit of maskelynite formation (recrystallized by contact metamorphism). Maskelynite is most evident in coarse-grained mafic rocks where its recrystallized equivalent imparts a distinctive porcellaneous appearance to the rock (seen in anorthositic gabbro a few hundred metres south of the Chicago Mine in Drury Township).

Chapter 11

Evolution of Ideas About the Origin of the Sudbury Igneous Complex and its Associated Ni-Cu-PGE Mineralization

A.J. Naldrett

Introduction

The presence of sulphides was initially reported at Sudbury in an 1856 Geological Survey of Canada (GSC) report (Murray 1857) as documented in the following quotation:

> "Previous to my visit to Whitefish lake, I had been informed by Mr Salter that local attraction of a magnet had been observed by himself while running the meridian line and he expressed it to be his opinion that the presence of a large body of iron ore was the immediate cause. When therefore, I came to the part indicated by Mr Salter, I made a very careful examination not only in the direction of the meridian line but for a considerable distance on each side of it, and the result of my examination was that the local attraction, which I found exactly as described by Mr Salter, was owing to an immense mass of magnetic trap. Specimens of this trap given to Mr Hunt for analysis and the result of the investigation shows that it contains magnetic iron ore and magnetic iron pyrites generally disseminated throughout the rock, the former in very small grains: titaniferous iron was found associated with the magnetic ore, and a small quantity of nickel and copper with the pyrites."

This report was not followed up and it was not until 1883, during construction of the Canadian Pacific Railway, that massive sulphides were uncovered at Sudbury in the vicinity of what became the Murray Mine. A number of companies then started operations, the largest of which was the Canadian Copper Company. Production commenced in 1886. It is ironic that the Canadian Copper Company did not realize that they were mining nickel in addition to copper until a shipment of ore sent from Creighton Mine to the Orford Copper Co. in New Jersey proved difficult to smelt, and, on analysis, was shown to contain nickel (Holloway et al. 1917)! The nickel proved something of an embarrassment because world demand at the time was only about 1000 tons per year, but Samuel J. Ritchie, President of Canadian Copper, had contacts with General B.F. Tracy, Secretary of the US Navy, which resulted in the recognition of the importance of ferronickel alloys in armour plate. By 1905, Sudbury production was exceeding that of the laterite deposits in New Caledonia, an island about 1000 km offshore from Australia.

Origin of the Sudbury Igneous Complex

EARLY DAYS

In 1886, the GSC followed up on the early work of Murray by sending R. Bell and A.E. Barlow to begin a programme of regional mapping. The results of this were published in 1891 (Bell 1891a), where the first reference is made to the basinal shape of the Sudbury structure. Barlow (1907) recognized the subdivision of the Sudbury Igneous Complex (SIC) into two principal units: a lower noritic and overlying granophyric layer. A.P. Coleman commenced mapping for the Ontario Bureau of Mines in 1902, and his initial map (Coleman 1905) was the first to show that the SIC outcropped around the whole of the Sudbury basin, and, in fact, defined the structure. He argued that the SIC had been intruded as a sill along the contact between the overlying breccias and sediments (Figure 11.1A), which he correlated with the Animikie of the Lake Superior area, and older basement rocks. His revision of his 1905 map (Coleman 1913), of which a simplified copy is shown here as Figure 11.2, differs very little in the disposition of the rock types from present-day maps.

Although many early workers regarded the SIC as the product of differentiation of a single magma, the high proportion of granophyre to norite bothered geologists, as did the absence of what they regarded as "Animikie"-type strata (the Onaping, Onwatin and Chelmsford formations) outside the structure. Knight (1917) proposed that the Sudbury basin was a caldera, that the centre of the structure had collapsed, carrying the "Animikie" formations with it, and thus preserving them from the erosion that had removed them outside the basin (see Figure 11.1B). He suggested that the norite and granophyre were not units of a sill, but were two ring-dike-like injections of magma up the circular fracture system encircling the collapse. The relative proportions of granophyre and norite continued to concern geologists well into the 1950s. Phemister (1926) endorsed Knight's viewpoint of two ring-dike injections, re-affirming this view 10 years later (Phemister 1937). W.H. Collins of the GSC undertook one of the most exhaustive studies of the SIC ever published (Collins 1934, 1935, 1936, 1937) and concluded that it was a differentiated sill. Wilson (1956) sought to address the question of the relative proportions of granophyre and norite by proposing that the SIC has the form of a lopolith (that is, a funnel-shaped intrusion) and that the high proportion of granophyre exposed at surface is counter-balanced by a mass of ultramafic rock hidden at depth (Figure 11.1C). This view was challenged by Thomson and Williams (1959) who, in their paper "The Myth of the Sudbury Lopolith", reaffirmed the caldera model. They pointed to the enormous release of energy that was required to form the Onaping Formation, and suggested that it was a series of glowing avalanche deposits (pyroclastic flows) into the caldera.

While most early work was concentrated on the SIC and its ores, the 1950s saw increasing attention being paid to the breccias developed at Sudbury. These include the dikes of Sudbury breccia developed in country rocks external to the basin, and the Onaping Formation within the structure. Speers (1957) concluded that the former were the result of high-pressure fluids related to diatreme or volcanic activity. The early interpretation of Burrows and Rickaby (1930) that the Onaping Formation was volcanic was reinforced, as discussed above, by the work of Thomson (1957) and Williams (1957), who concluded that it was an accumulation of more than 300 cubic miles of pyroclastic flows, whose abrupt eruption from an underlying magma chamber had given rise to the subsidence that is now preserved as the Sudbury basin. All geologists of this period recognized that the formation of the Sudbury structure had been accompanied by an enormous release of energy.

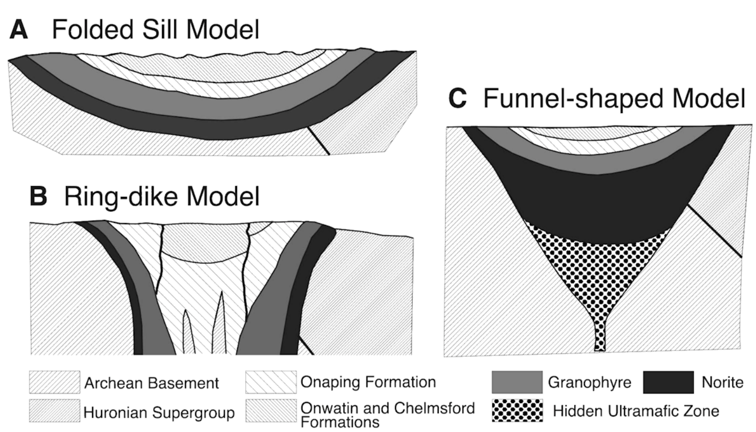


Figure 11.1. The folded sill, ring-dike and funnel-shaped intrusion models that dominated early thinking about the Sudbury Igneous Complex.

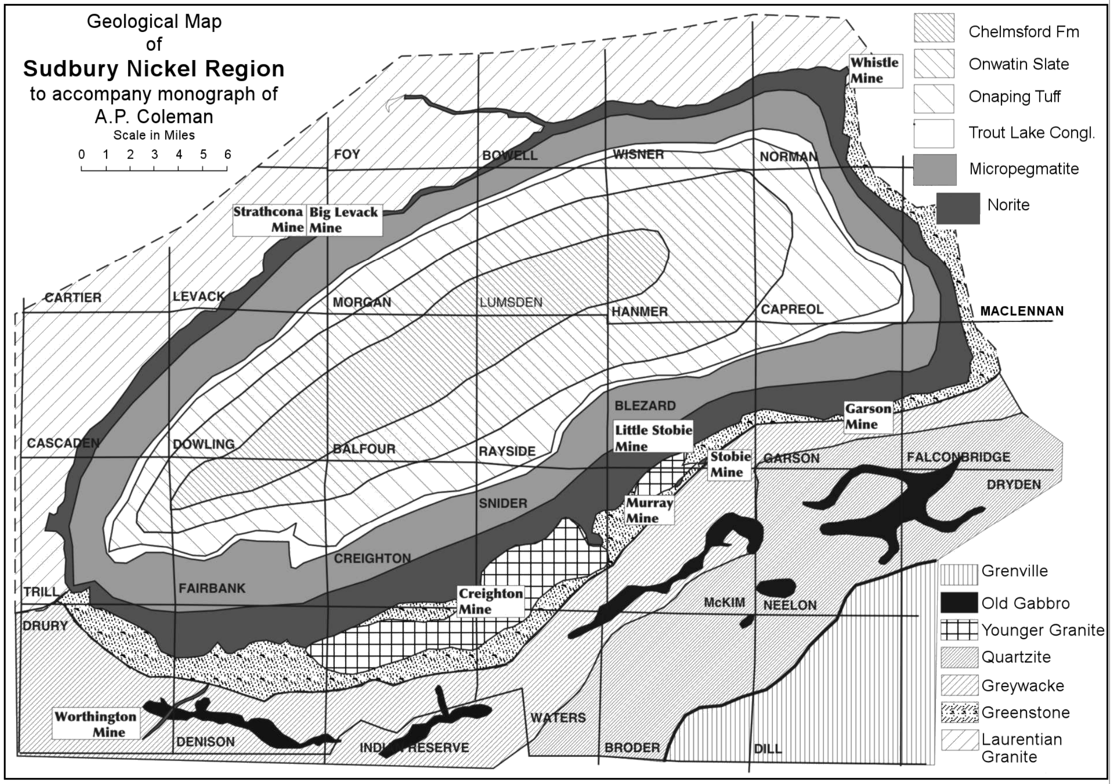


Figure 11.2. A simplified copy of A.P. Coleman's 1913 revision of his 1905 map of the Sudbury area (Coleman 1913).

THE ASTROBLEME HYPOTHESIS

Robert Dietz first came to Sudbury in May 1962. He was a US Navy oceanographer, who, as a sideline, had developed an interest in astroblemes, the scars on the Earth's surface left by the impact of meteorites and comets. He had been working on the Vredefort dome in South Africa, and came to Sudbury to look for shatter cones, which he believed were the result of the enormous shock pressures developed on impact. Regrettably, on his first visit he had selected to look at the Onaping Formation, which, being post-impact, was the wrong environment to investigate, but the following year, after more thought, he returned to Sudbury and was successful in identifying shatter cones in quartzite south of Kelly Lake. His 1964 paper (Dietz 1964) was the first proposing that the Sudbury structure was an astrobleme. The timing of this paper coincided with the awakening interest of NASA in the types of rock fabrics and structures that astronauts might find on the moon, and many scientists interested in extraterrestrial impact came to Sudbury to test out their theories. The result was that over the next 10 years, visitors and local Sudbury geologists accumulated a tremendous amount of data relating to an impact origin. These data include the distribution of Sudbury breccia, which is equated to the Vredefort pseudotachylite (breccia dikes that are the consequence of shock melting) (Figure 11.3A), the wide distribution of shatter cones in the country rocks around Sudbury (Figure 11.3B), the presence and distribution of "shock-deformation" structures in individual minerals (Figure 11.3C), and the analogies that can be drawn between the Onaping Formation and impact breccias as they were known from the Riess crater in Bavaria (Avermann 1999).

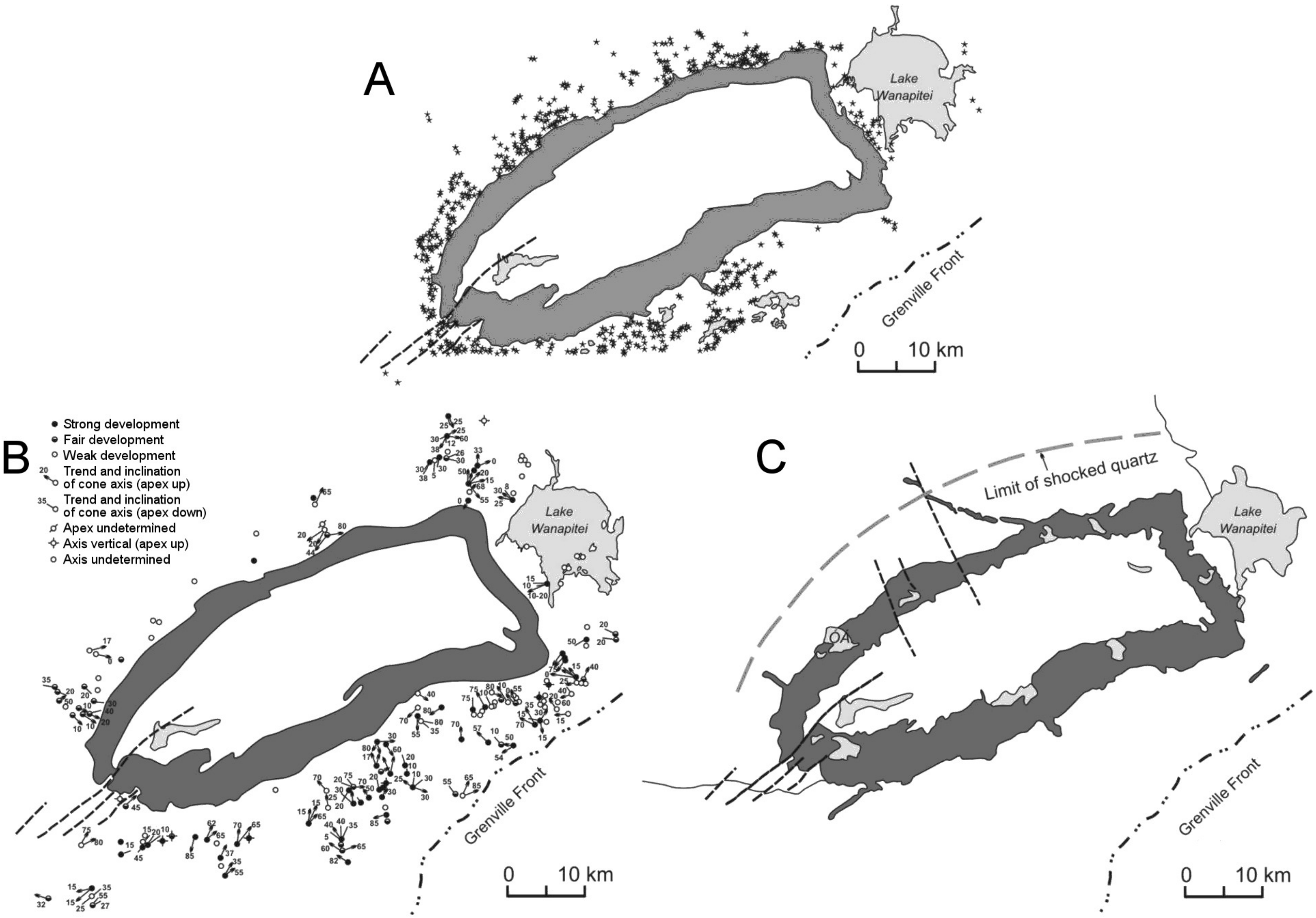


Figure 11.3. Evidence supporting astrobleme origin of the SIC. **A.** Distribution of zones of Sudbury breccia (shown by stars) around the margin of the SIC (*after* Dressler 1984b). **B.** Distribution of shatter cones and their orientation in the Sudbury area (*after* Dressler 1984b). **C.** Limit of distribution of shocked quartz in the Sudbury area (*after* Grieve 1994).

The impact hypothesis generated strong debate over the next 20 years. Much of this debate was focused on the interpretation of structures within the Onaping Formation and was comprehensively summarised by Muir and Peredery (1984) in "*The Geology and Ore Deposits of the Sudbury Structure*" (Pye et al. 1984). Despite the joint nature of their paper, Muir endorsed a volcanogenic origin for the Sudbury structure, basing his endorsement primarily on localised cross-cutting stratification within the Onaping that suggests formation as a result of repetitive explosive events that are incompatible with a single, catastrophic impact. Peredery supported an impact origin, referring to an increasing body of evidence indicative of impact-related shock metamorphism, noting the overall lack of regional stratification in the Onaping Formation, and interpreting the localised cross-cutting stratigraphy to high-energy events occurring during the immediately post-impact filling of the crater.

As discussed above, two aspects of Sudbury geology had always puzzled investigators: (i) the source of the energy that had given rise to so much brecciation; and (ii) the large proportion of granophyre to mafic rocks within the SIC. An impact origin provided a plausible answer for the energy involved in the brecciation.

THE SUDBURY IGNEOUS COMPLEX

With regard to the SIC itself, Naldrett et al. (1970) showed that the "norite" forming the lower part of the SIC exhibits cryptic variation indicative of *in situ* differentiation (Figure 11.4), an observation that remains unique to Sudbury amongst other impact structures. The early trace element studies of Kuo and Crocket (1979) showed that the chondrite-normalised REE profiles of the SIC rocks are much steeper than uncontaminated mantle basalt, which they interpreted to indicate that the initial SIC magma had incorporated a high proportion of country rocks (Figure 11.5). This explained both its quartz-rich nature and the high proportion of granophyre. Faggart et al. (1985), reporting on Sm-Nd and Rb-Sr isotopic studies of the SIC, noted that both the Main Mass and Sublayer of the SIC contained less-radiogenic Nd and more-radiogenic Sr than the estimate for Bulk Earth. They interpreted this to indicate that the whole of the complex was an impact melt. Naldrett et al. (1986) pointed out that major, trace element and isotopic data are explicable as the result of the contamination of flood basalt magma by a mix of country rocks exposed at the present erosion level of the SIC. Chai and Eckstrand (1994) documented a marked compositional break between the Quartz Gabbro and the overlying Granophyre, and argued that this implied derivation from two different magmas originating from different sources. They postulated that the norite and Quartz Gabbro were the product of a primary mantle melt that had become contaminated by Archean granulites typical of the lower crust, whereas the Granophyre was an upper crustal, impact melt. However, Lightfoot et al. (1997a) noted that, with the exception of Sr, P, Eu and Ti (which are very dependent on addition or removal of plagioclase, apatite or Fe-Ti oxides) the Felsic Norite, Quartz Gabbro and Granophyre have extremely similar trace element patterns. In particular, they drew attention to the similarity in Th/Zr ratios (0.04-0.05) between gabbro-noritic rocks of the SIC and the Granophyre which, they argued, would be an extraordinary coincidence if they had been contaminated by, or derived from different crustal reservoirs.

In 1990, Sudbury became the target of one of Canada's Lithoprobe projects, whereby a series of vibroseis traverses were conducted across the SIC. The base of the SIC, and the contact between the granophyre and underlying gabbro/norite were well defined by the seismic data, and Milkereit et al. (1992) interpreted the results to indicate that the base of the complex as seen on the North Range extended continuously at least as far south as the southern margin (Figure 11.6). The South Range had been thrust northward over the North Range along a zone of plastic deformation known as the South Range Shear Zone (Shanks and Schwerdtner 1991). The implication of the Lithoprobe study was that the original size of the Sudbury structure was far greater than the current dimensions (30 x 60 km), and of the order of 200 km. This led Grieve et al. (1991) to comment that the volume of impact melt likely to have formed during the generation of a crater of this magnitude was equal to or greater than the volume of the SIC, and to suggest that the whole of the SIC is an impact melt, with no primary mantle material. Subsequently, Grieve (1994) showed, using least-squares mixing models, that the average composition of the SIC corresponds to a mix of an Archean granite-greenstone terrain, with possibly a small component of Huronian cover rocks.

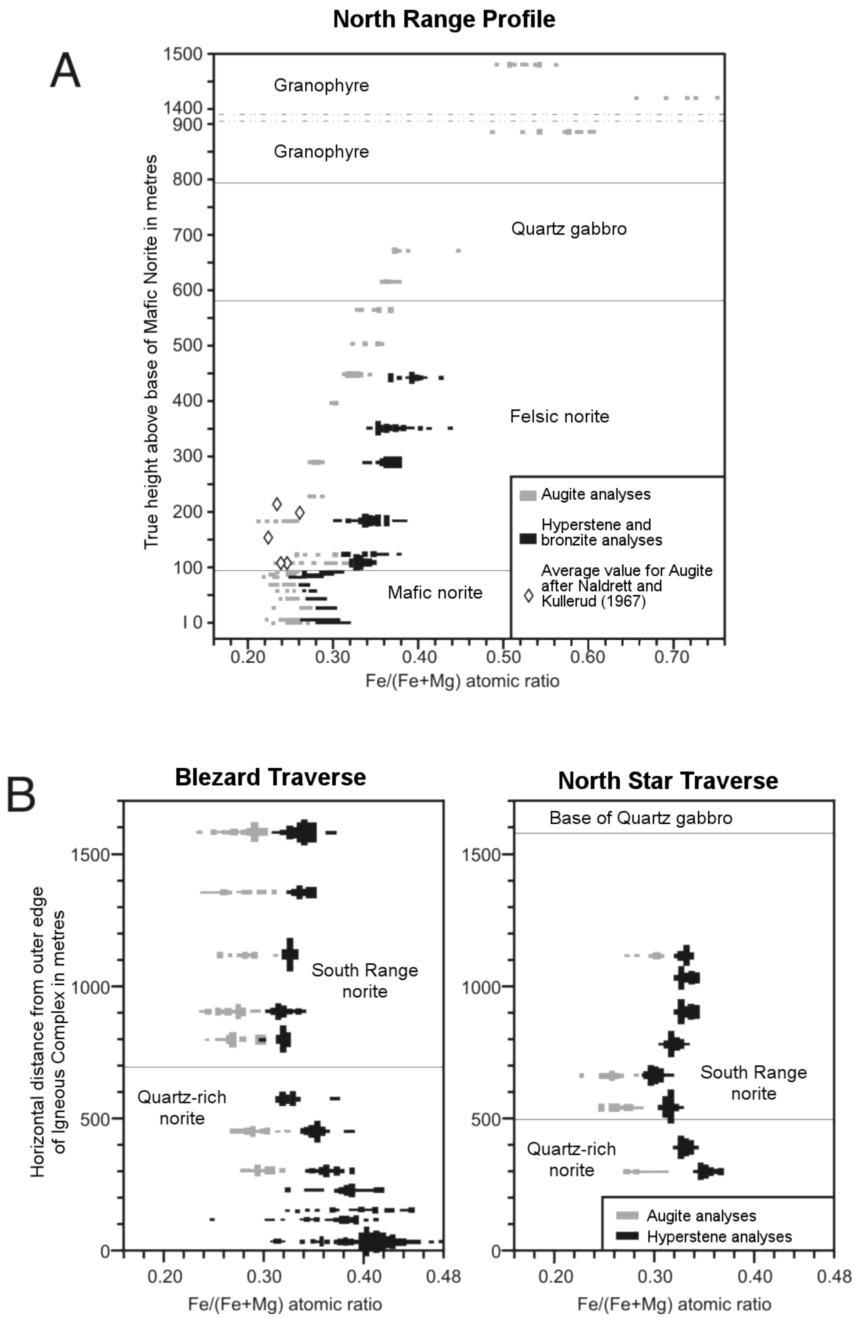


Figure 11.4. Plots of FeO/(FeO+MgO) of ortho- and clinopyroxene through **A.** a profile through the North Range of the SIC, and **B.** through the Blezard and North Star traverses across the South Range of the SIC. Note that distances in A are height above the base of the Mafic Norite while those in B are distances from the outer southern margin of the SIC measured horizontally. *After* Naldrett et al. (1970).

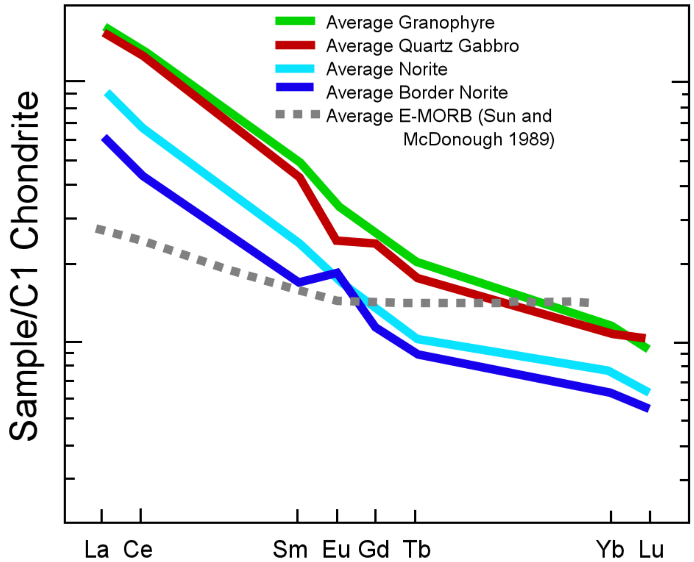


Figure 11.5. REE profiles of average Sudbury rocks, *modified after* Kuo and Crocket (1979).

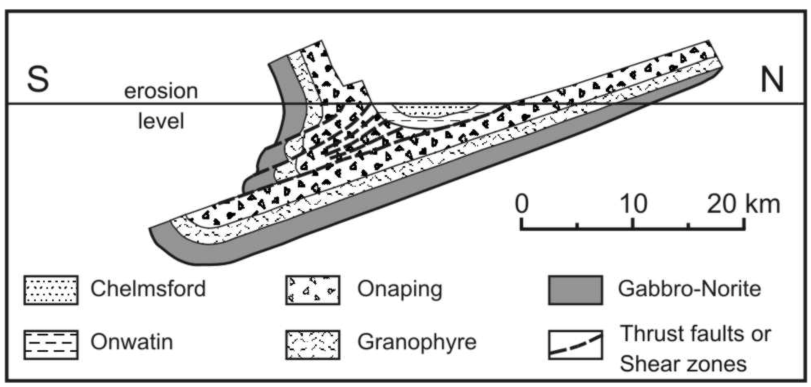


Figure 11.6. North-south cross-section of the Sudbury structure as suggested by seismic data (*after* Shanks and Schwerdtner 1991).

Walker et al. (1991) showed that Re-Os isotope systematics required that the ancient crust had contributed 60-75% of the osmium contained in the McCreedy West and Falconbridge ores, and probably nearly 100% of that contained in the Strathcona ores. Dickin et al. (1992) pointed out that, given the uncertainties involved in assumptions about the osmium content and isotopic composition of continental crust, the osmium isotopic data are consistent with 100% of the osmium being of crustal origin. Morgan et al. (2002) concluded on the basis of a very precise Negative Thermal Ionization Mass Spectrometry (NTIMS) study of $^{186}Os/^{188}Os$ and $^{187}Os/^{188}Os$ ratios in samples of Sudbury ore that the osmium is clearly crustal, and can be explained as having been derived from a binary mixture of Archean Superior and Huronian metasedimentary rocks (Figure 11.7). However, they found that ores with low $^{190}Pt/^{188}Os$ ratios from two deposits, Falconbridge and McCreedy West, had $^{186}Os/^{188}Os$ ratios that were substantially superchondritic. They suggested that this could be due to the admixture of a third component, most likely Archean or early Proterozoic mafic rock with $^{190}Pt/^{188}Os$ approximately equal to 1, that had also been sampled by the impact.

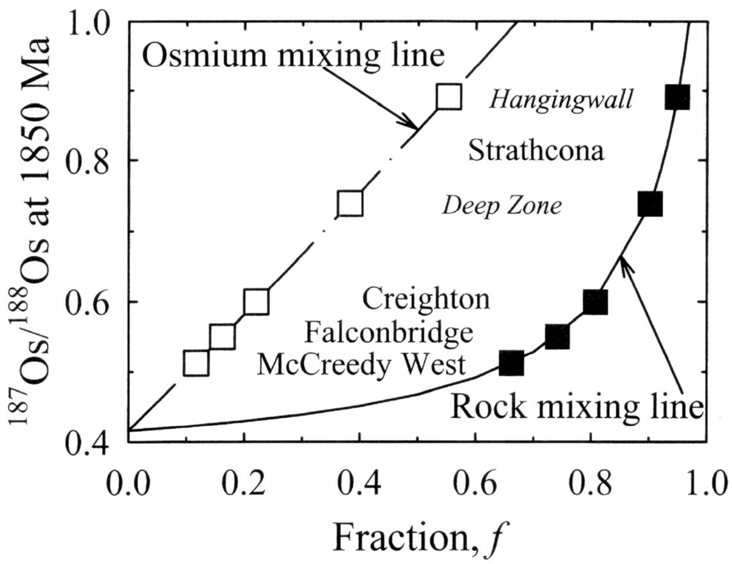


Figure 11.7. Mixing model for $^{187}Os/^{188}Os$ in the Sudbury ores assuming an Archean end member with 00.72 ppb Os and $(^{187}Os/^{188}Os)_{1850\ Ma} = 1.319$, and a Proterozoic end member with 0.108 ppb Os and $(^{187}Os/^{188}Os)_{1850\ Ma} = 0.416$. *From* Morgan et al. 2002 who used data for Creighton from Dickin et al. 1992.

Ariskin et al. (1999) modeled the crystallization of the SIC, using the computer program COMAGMAT-3.5 (Ariskin 1997) that had been developed with respect to other differentiated mafic bodies. When they used Naldrett and Hewins' (1984) marginal Quartz-rich Norite as a starting composition, their modelling gave a very close approximation to both the chemical variations and appearance and disappearance of phases in the lower, mafic portion of the SIC. However, whatever initial magma composition they chose, they were unable to account for the proportion of granophyre exposed at surface. They concluded that the units of the SIC, excluding the granophyre, were the consequence of fractional crystallization of a magma similar to the Quartz-rich Norite in composition, and that the granophyre is largely the consequence of subsequent assimilation in the hot environment of an impact crater.

Ivanov and Deutsch (1999) used a modified version of Amsden et al.'s (1980) SALE computer programme for fluid flow (i) to construct a time-dependent model for the formation of the transient crater, and thus show the progression in the distortion of layers in the target area and the position of isotherms close to and beneath the crater following impact; (ii) to model the maximum shock pressures experienced in the target area; and (iii) to model the temperature evolution within and beneath the melt pool from 10^4 to 10^7 years after impact (Figure 11.8). Assuming acoustic fluidization of the target rocks, their modeling showed that a substantial amount of lower crustal material could have been brought to surface as the result of fluid flow and not merely as inclusions or impact melt. This would account for the 5 to 10 km wide ring of granulitic gneiss that surrounds much of the northern perimeter of the SIC. Also, the transient crater would have had a diameter of 75 to 100 km, but would subsequently have collapsed to give rise to a crater with a 200 km diameter. An important point raised by their study is that the initial temperature of the impact melt would have been about 1727°C, and that super-liquidus temperatures (above 1180°C) would have persisted for 100 000 to 250 000 years after impact (see the implications with respect to sulphide settling that are discussed below).

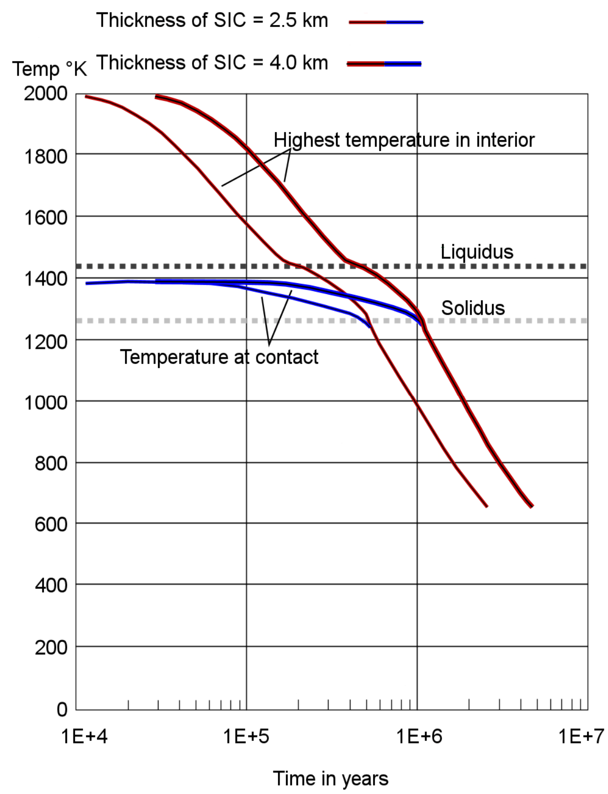


Figure 11.8. Ivanov and Deutch's (1999) calculations as the variation in the temperature of the interior of the SIC with time.

Addressing the fact that the fractional crystallisation model cannot account for the granophyre within the SIC, Naldrett (1999) suggested that at an early stage, the Sudbury magma separated into two layers, an upper layer of felsic composition and a lower layer of more mafic composition, as a result of the process of doubly diffusive convection (McBirney 1987, 1985; McBirney and Noyes 1979). While the overall magma was superheated, considerable assimilation of the overlying Onaping Formation occurred, adding to the mass of the granophyre. Crystallisation then proceeded from the top downward within the granophyre and from the sides and base inward in the case of the lower layer (the sides of the granophyre are not observable). Zieg and Marsh (2005) and Marsh (2006) have concluded that the separation into two layers occurred because the initial impact melt was inhomogeneous, composed of unmixed globules of disparate composition that had resulted from the melting of felsic and mafic portions of the target rocks. With time, the less dense felsic globules rose to form an upper, felsic layer. Homogenisation and subsequent crystallisation within the two layers gave rise to the granophyric and noritic zones of the SIC that we see today.

SUBLAYER AND OFFSET DIKES

The evolution of thinking about the Sublayer and Offset dikes requires a special mention because of the importance of these units as the host to the bulk of the mineralisation at Sudbury. The term "Sublayer" was first introduced into the Sudbury literature by Souch, Podolsky et al. (1969) who remarked that it comprised bodies of distinctive fine- to medium-grained, inclusion-bearing norite and gabbro, up to 200-300 m thick, that occurred intermittently around the outer perimeter of the SIC. Subsequently, the Sublayer has been shown to have a distinctive chemical composition (see below). Traditionally, the Offset Dikes, which contain a similar population of inclusions, have been grouped within the Sublayer.

However, Lightfoot et al. (1997a) showed that the quartz diorite, which comprises the outer portions of many of the Offset Dikes, is close to the average composition of the Main Mass of the SIC in its trace element content, and is distinctly different than that of the Contact Sublayer.

Geological relationships between the Sublayer and the Main Mass give conflicting evidence as to the relative ages of the two units. Zones of marginal Quartz-rich Norite have been observed apparently included within Sublayer, and zones of Sublayer have been observed enclosed in norite of the Main Mass of the Complex. Intrusive contacts between the Main Mass and Sublayer are never marked by fine-grained chill zones, suggesting that they were nearly contemporaneous. On the North Range, the distinction between Main Mass and Sublayer is consistently clear, with the Sublayer having the finer grain size and lower quartz content. This distinction is not necessarily the case on the South Range, where a number of researchers have commented on gradations between the two units (Slaught 1951; Cochrane 1984).

The inclusions in the Sublayer can be divided into two groups: those of obviously local derivation, and those composed of mafic and ultramafic rocks, many of which do not outcrop in the immediate Sudbury area. Scribbins et al. (1984) described the mafic-ultramafic inclusions as ranging from peridotite, through clino- and orthopyroxenites, to olivine gabbro and norite. Most either display cataclastic or cumulus textures.

The match between the relative proportions of trace elements in the Felsic Norite and Sublayer is much less close than between Felsic Norite and the other units of the SIC, including the offsets. The Sublayer rocks are poorer in LREE (Light Rare Earth Elements) and LILE (Large Ion Lithophile Elements) but have similar HREE (Heavy Rare Earth Elements) and HFSE (High Field Strength Elements) as compared to the Felsic Norite. Lightfoot et al. (1997a) argued that these differences cannot be explained by closed system fractional crystallization or partial melting. Sublayer samples from a single embayment have similar compositions, but samples from different embayments have different compositions. It is likely that these differences are due to the provenance of the bulk of the Sublayer in each case being of very local derivation, a concept that is supported by Nd and Sr isotopic data that are discussed below.

On the basis of their field and geochemical studies, Lightfoot et al. (1997a) argued that there is a close genetic relationship between the Offsets and the Main Mass of the SIC. Lightfoot et al. (2001) and Lightfoot and Farrow (2002) emphasized the difference between the marginal (free of inclusions and nearly free of sulphide) and inclusion- and sulphide-rich variants of offset rocks. They pointed out that the marginal zone to the Worthington offset and their calculated average composition for the Main Mass are nearly identical in major and trace elements, with the exception of the chalcophile elements. They suggested that the marginal offset rock represents the initial composition of the SIC magma, and proposed that the marginal offset rocks were introduced early into fractures resulting from the impact, and that the inclusion- and sulphide-rich phases were introduced later, after sulphide immiscibility had occurred in the Main Mass magma (Figure 11.9).

Tuchscherer and Spray (2002) studied the longest of the Offset dikes, the Foy offset, which extends for at least 30 km from the margin of the SIC. They also suggest that the dike was emplaced very early during the development of the Sudbury structure before significant differentiation had occurred, and that chilled margins of the dike are representative of the initial Sudbury magma. However, their field studies, along with those of Murphy and Spray (2002) on the Whistle–Parkin offset dike indicate that the inclusion- and sulphide-enriched phase preceded the fine-grained phase, albeit by only a very short time interval. The implications of these observations, which appear to be contrary to the emerging picture of sulphide immiscibility occurring somewhat later than the formation of the impact melt, remain to be assessed.

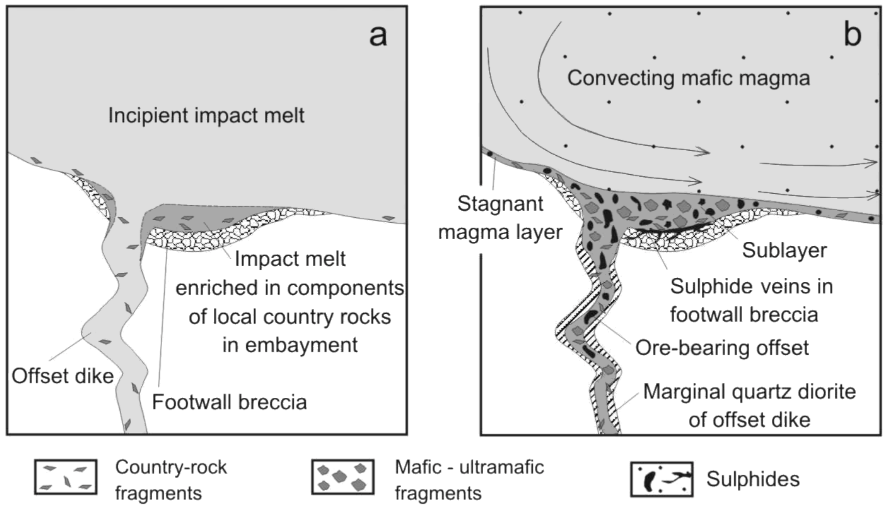


Figure 11.9. Model for development of Offset dikes and embayments in the footwall of the Sudbury Igneous Complex **(a)** at the time of impact and **(b)** after the onset of sulphide immiscibility and settling. *Adapted from* the model of Lightfoot et al. (2001, their Figure 12).

Origin of the Ores

The earliest workers at Sudbury were strong advocates of the magmatic sulphide hypothesis. Bell (1891b) described the ores and proposed that they were the result of the differentiation of the magma responsible for the SIC, together with separation and settling of a dense sulphide melt. Barlow (1907) recognized the following: (1) the subdivision of the SIC into two principal units (norite and overlying granophyre); (2) the occurrence of the orebodies in embayments along the lower contact of the norite; and (3) the idea of settling of sulphide liquid. Coleman (1905, 1913) was perhaps the most ardent advocate that gravitationally induced sulphide liquid settling was in part responsible for the origin of the ores. His arguments included the restriction of the sulphides to the outer (= lower) contact of the SIC, the intimate mixture of sulphide and norite in all proportions, the lack of hydrothermal alteration, and the occurrence of the largest concentrations of sulphides in embayments of the lower contact into the footwall.

The magmatic viewpoint of the early field geologists at Sudbury did not go unchallenged. Dickson (1903) suggested that the mineralisation had been deposited by hot aqueous fluids. C.W. Knight, in his contribution to the Report of the Royal Ontario Nickel Commission (Knight 1917), described the deposits, pointed to the almost universal occurrence of the ores in breccias in which partial replacement of the fragments could be observed, and concluded that they were hydrothermal. In a footnote, he cites a review of Ni-Cu deposits by Tolman and Rogers (1916) who concluded that “ores of this class throughout the world have been introduced at a late magmatic stage by mineralisers, and that the ore minerals replace silicates” (Knight’s words). Wandke and Hoffman (1924) made an extensive microscopic study of samples from Sudbury and also concluded that they were hydrothermal in origin.

One of the major arguments used to support a late-stage hydrothermal emplacement of the ores was their occurrence in breccias composed of Murray and Creighton granite that, on the basis of apophyses of this material cutting the lower contact of the norite, were regarded as younger than the SIC. It was only when dating showed that these intrusions were in fact older than the SIC that this argument was invalidated. The latest date for the intrusions is 2.420 Ga (Smith et al. 1999). The apophyses cutting the norite are now recognised as due to remobilisation of the granite by the heat of the impact.

Throughout the 1930s, 1940s and 1950s, arguments about the magmatic versus hydrothermal origin of the Sudbury ores persisted, with those who mapped the SIC on surface tending to favour a magmatic origin [Coleman, Moore and Walker (1929); Collins (1937)], whereas those who were most familiar with the mines [e.g., Yates (Chief geologist of INCO) 1938, 1948; Davidson (Chief Geologist of Falconbridge) 1948; and Lochhead (Falconbridge Chief Sudbury Mines Geologist) 1955] favouring a hydrothermal origin.

The appearance of J.E. Hawley's seminal study, "The Sudbury ores: Their mineralogy and origin" (Hawley 1962) convinced the greater part of subsequent generations of mineral deposit geologists as to the magmatic origin of the Sudbury ores. This work comprises (Part I) a review of previous work and outstanding problems at Sudbury, (Part II) an extensive study of sulphide mineralogy conducted and co-authored jointly with R.L. Stanton, and (Part III) Hawley's personal conclusions as to the origin of the ores. This is not to say that all argument as to a hydrothermal origin of certain aspects of the mineralisation at Sudbury became stifled, as is discussed below.

It has now become an established fact that the majority of rich concentrations of magmatic, Ni-Cu sulphide deposits, including those at Kambalda, Western Australia; Noril'sk, Siberia; Raglan, Ungava; Voisey's Bay, Labrador; and Jinchuan, China, occur within structures through which magma has flowed, rather than being distributed semi-uniformly at the base of sills (Naldrett 2004). Sudbury is the major exception in this regard. Referring to Ivanov and Deutch's (1999) modeling of the cooling of the SIC (Figure 11.8), Naldrett (1999) pointed out that if sulphide immiscibility had occurred during the interval when the Sudbury impact melt was between the initial temperature and its liquidus, the sulphides would have settled through the high-temperature, low-viscosity magma rapidly, unobstructed by silicates (except close to the contact), thus accounting for the basal sulphide concentrations observed at Sudbury. Li and Ripley (2005) applied their revised equation for calculating the solubility of sulphur in a mafic magma to the Sudbury impact melt. Using Keays and Lightfoot's (2004) weighted average of the sulphur content of mafic and felsic norite, they showed that sulphide immiscibility would occur at 1500°C, over 300°C above the liquidus temperature (Figure 11.10).

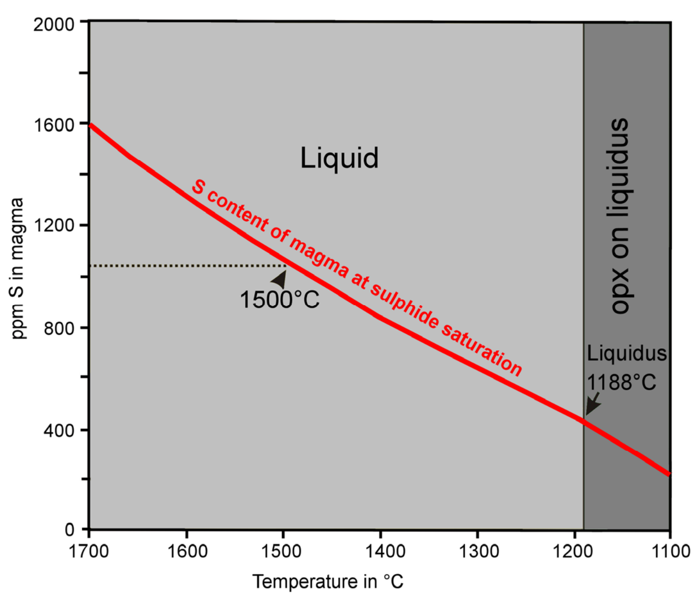


Figure 11.10. The sulphur content of the SIC at sulphide saturation plotted against temperature. Given Keays and Lightfoot's (2004) estimate of the sulphur content, sulphide immiscibility would be expected to occur at a temperature 300°C above the liquidus temperature (*after* Li and Ripley 2005).

Lightfoot et al. (2001) and Keays and Lightfoot (2004) found that the tenor of Ni and Cu in sulphides in a borehole from the eastern part of the SIC decreased upward asymptotically from values of 4 to 8 wt % of both metals within ore deposits and in immediately overlying norite to a value of around 1 wt % in norite 1200 m above the base. They interpreted these data to imply that the sulphides had segregated progressively from a body of magma that, as a result of the segregation, was itself becoming progressively depleted in Ni and Cu. Using established partition coefficients, it is possible to estimate the Ni and Cu contents of the SIC magma at the time of segregation of the sulphides. As shown in Figure 11.11, at the time of the segregation of early sulphides, these estimates indicate that the magma contained 120 ppm Ni and 50 ppm Cu, concentrations that are significantly higher than would be expected from the wholesale melting of typical continental crust. Naldrett (2004) suggested that a mafic/ultramafic intrusion belonging to the series of 2.45 Ga intrusions that occur close to the presumed zone of rifting that gave rise to the Huronian ocean south of Sudbury, existed in the target area at depth. These intrusions contain deposits of Ni-Cu-PGE sulphide that are currently being explored. Impact melting of one of these intrusions, and incorporation of the melt in the overall impact melt could account for the higher-than-normal Ni and Cu contents, and also the comparatively high sulphur content of the SIC. Several small bodies belonging to this class of intrusion are located intermittently around the outer margin of the SIC.

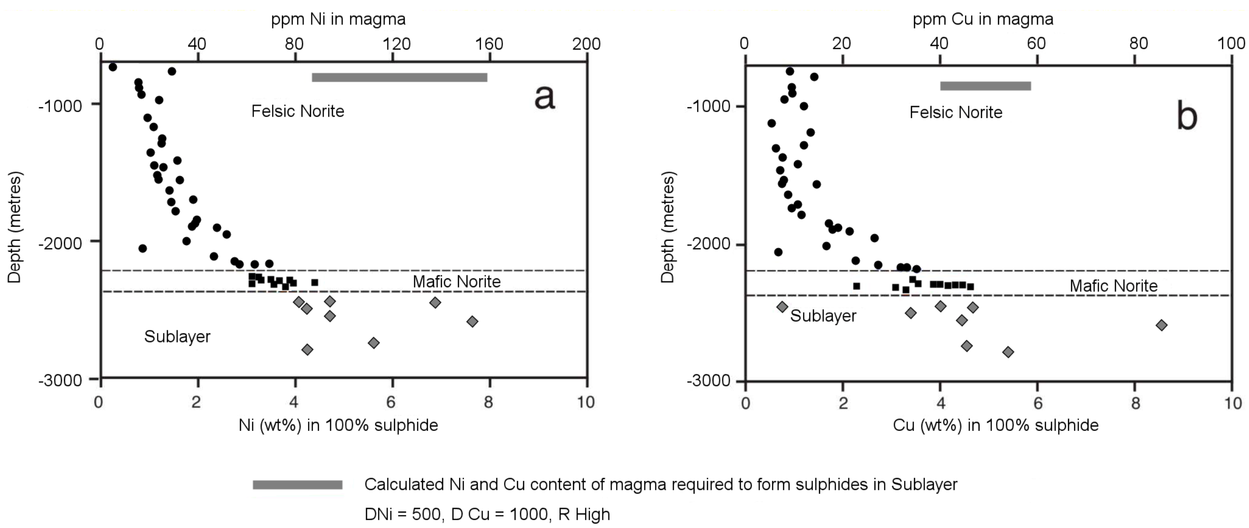


Figure 11.11. Variation of the calculated Ni and Cu contents of the sulphides in the ores and overlying units of the SIC (*after* Keays and Lightfoot 2004). The Ni and Cu content of the source magma is proportional to the Ni and Cu content of the sulphides, provided that the sulphides have had completely free access to the magma (i.e., that the R factor was high). If the access has not been completely free, the Ni and Cu content of the magma must have been higher than that dictated by simple proportionality.

Following his 1962 monograph [Hawley, op. cit.], Hawley (1965) published a second seminal paper on the Sudbury ores in which he described "upside-down" zoning in the Frood–Stobie deposit, and was the first to suggest that this might be due to fractional crystallisation of a Fe-Ni-Cu sulphide liquid, with the fractionated liquid being expelled downward. Keays and Crocket (1970) were amongst the first to study the distribution of PGE in the Sudbury ores. They attributed variations in the distribution to PGE fractionation during the fractional crystallisation of a sulphide liquid, referencing Hawley's 1965 paper. Subsequently, this concept was enlarged upon by Chyi and Crocket (1976). Hoffman et al. (1979) documented the progressive enrichment in Cu, Pd, Pt and Au on proceeding from hanging wall to footwall at the McCreedy deposit (known as Levack West at the time) and Abel et al. (1979) described copper-rich veins in the footwall at the Strathcona deposit. Abel et al.'s paper was the first account of mineralisation that is detached from and lies stratigraphically below the contact ore bodies at Sudbury; subsequently it has transpired that this ore type, which has come to be known as "footwall copper" ore, is an important feature of the northern and eastern "ranges" of the Sudbury structure. Naldrett et al. (1982) showed that the zoning from top to base (hanging wall to footwall) across many Sudbury deposits could be interpreted as the result of the residual liquid migrating towards the footwall. The Falconbridge deposit stood out amongst all others that had been studied in being deficient in the incompatible elements, and they suggested that as much as half of the deposit was missing, possibly as a result of faulting.

While the magmatic origin of the principal ore deposits at Sudbury is now firmly established, there is still debate about the origin of the Cu-, Pt-, Pd- and Au-rich veins in the footwall of the SIC. Li and Naldrett (1993), Naldrett et al. (1994), and many other authors attributed the veins to the migration of a Cu-rich residual away from the site of crystallisation of pyrrhotite-rich ore bodies at the contact. Although recognising that the veins contain chloride-rich fluid inclusions, and have 5 to 20 cm zones of alteration at their margins, authors favouring a magmatic origin have attributed these effects to fluids concentrated within the fractionated sulphide liquid.

Farrow and Watkinson (1992, 1997) and Watkinson (1999) found that the fluid inclusions are extremely saline and have trapping temperatures revealing multiple trapping events between 100^o and 400^{o}C. They cited stable isotope data that are consistent with fluids derived from mixed formational brines, and also with fluids that have equilibrated at high temperature with igneous and metamorphic rocks. They proposed that during the cooling of the SIC and its ores, a hydrothermal system became established that reworked magmatic sulphides and re-deposited Cu, Ni and PGE in veins in the footwall.

Recently, experimental evidence has been developed that supports the derivation of chloride-rich brines from a fractionating sulphide melt. Mungall and Brenan (2003) studied the partitioning of halogens between silicate magma and sulphide melt, and showed that while halogens prefer the silicate magma, sufficient Cl is partitioned into the sulphide melt, that, after 95% fractionation, the residual melt will contain significant Cl. Experimental container problems mean that the water content of Fe-S-O melts is difficult to determine. Naldrett (1969) reported experimental determinations of the solidus of pyrrhotite under aqueous and dry conditions that indicated that water had no detectable effect on the temperature, concluding that less than 10% water dissolves in the sulphide melt. However, Wykes and Mavrogenes (2005) showed that the presence of H_2O at 1.5 GPa lowers the eutectic in the FeS-PbS-ZnS system by 35°C, which they interpret to indicate that significant H_2O dissolves in the melt. They commented that application of their observations to Fe-Ni-Cu-S-O melts suggests that significant water could be present in a fractionated Cu-rich, Fe-Ni-Cu-S-O liquid. Combined with the observations of Mungall and Brenan (2003), one must conclude that sufficient water and chloride may be present in fractionated sulphide melts to cause the observed alteration. The implications of this concept to the broad aureoles of PGE-enrichment that are currently being discovered at Sudbury around footwall-copper ores remains to be evaluated.

Conclusions

In conclusion, there is widespread acceptance that:

1. The Sudbury structure is the consequence of extraterrestrial impact.

2. The crater had an original diameter, after the collapse of the initial transient crater, of in excess of 200 km.

3. The SIC is predominantly an impact melt. The granophyre probably separated from a more mafic magma at an early stage and was augmented by incorporation of the products of melting of the overlying Onaping Formation.

4. The Onaping Formation represents a combination of a basal surge deposit, fall-back breccia (suevite) and suevite that has been reworked in an aqueous environment.

5. The offsets are the result of the early emplacement of impact melt along fractures resulting from impact and subsequent crater readjustment.

6. Sulphide immiscibility occurred during cooling of the SIC, probably before much of the SIC had reached its liquidus temperature.

7. The sulphides settled into embayments in the impact crater wall and were injected into early-emplaced melt occupying the offset fractures.

8. Continuous segregation of sulphide led to a depletion of chalcophile metals within the remaining impact melt.

9. The ores fractionated as they cooled, giving rise to a Cu-, Pt-, Pd- and Au-enriched residual liquid which moved into the footwall, exploiting impact breccia and zones of pseudotachylite-like "Sudbury Breccia".

Some significant questions remain unanswered. These include:

1. The Sudbury structure lacks the central uplift that characterises all known impact craters of equivalent size. It is possible that the documented north-northwest thrusting of the South Range over the North Range has resulted in a central uplift being covered by the allocthonous rocks, but there is no evidence to support this.

2. The fractional crystallisation exhibited by the felsic norites and quartz gabbro of the SIC has not been documented in melt sheets of other impact craters that are apparently of equivalent size to the Sudbury structure.

3. The concentrations of Ni and Cu that appear to have been present in the initial impact melt exceed those to be expected in a melt of average Archean-Proterozoic crust. As suggested in this chapter, the Ni and Cu may have come from a mafic/ultramafic Paleoproterozoic intrusion that existed in the target area, but definitive evidence of this is lacking.

Chapter 12

The Ni-Cu-PGE Deposits of the Sudbury Igneous Complex

J.P. Golightly

Exploration and Mining History

The Sudbury district hosts some of the world's largest Ni-Cu-PGE magmatic sulphide deposits. The combined production and resource exceeds 1648 million metric tonnes grading approximately 1.2% Ni and 1.1% Cu (Lightfoot et al. 1997a). By 1994 the district had produced slightly more than 8 million tonnes each of Ni and Cu, and about 3184, 284, 279 and 100 tonnes of Ag, Pd, Pt and Au, respectively. Significant amounts of Os, Ir, Rh, Ru, Se, Te, S and Fe have also been recovered. Based on 2005 average metal values, more than 77 deposits have produced an estimated CDN$ 330 billion worth of metal in the past century (Ames and Farrow 2007). According to Brock Greenwell, Statistics Analyst, Ministry of Northern Development, Mines and Forestry (personal communication, 2008), the total value of production to 2007 was CDN$ 225 billion.

The first mineralization in the area was discovered by a surveyor (1856) and described by Murray (1857) of the Geological Survey of Canada. Several decades later the site was found to lie only 200 m west of the open pit of the Creighton Mine (Giblin 1984). The first discovery of mineralization, which led to the development of a mine, was made in 1883 during construction of the Canadian Pacific Railway. A rail-cut exposed high grade mineralization, which was later (1884) developed as the Murray Mine. By 1999, after 112 years of exploration, approximately 116 deposits have been found (Table 12.1). See Figure 1.2 for the locations of major deposits. From the late 1920s until around 2000, all significant magmatic sulphide deposits of the Sudbury Structure were the property of either INCO Ltd (now VALE INCO) or Falconbridge Ltd. (now XSTRATA nickel). The Creighton Mine, in continuous production since 1901, is the Western Hemisphere's deepest mine, with current production on the 2256 m level. Significant discoveries in the past 25 years include: 1) several deep deposits (>1000 m) along the Sudbury Igneous Complex (SIC)–footwall contact (e.g., Victor Deep, Lindsley, Trillabelle); 2) deposits in offset quartz diorite dikes (e.g., Kelly Lake); and 3) the entire class of previously unrecognized footwall vein systems (e.g., McCreedy East, Victor Deep). Since that time, some of the properties owned by the two major companies have been farmed out to junior mining companies, including FNX Mining Company Inc., Aurora Platinum Corp. and First Nickel Inc. They, and other juniors, have also been successful in finding new deposits or the extensions of existing deposits or potentially new deposit environments. An example of the latter is the hitherto unknown Trillabelle Radial Offset recently discovered by Wallbridge Mining Ltd.

Deposit Types

The Sudbury Structure hosts several types of Ni-Cu-PGE magmatic sulphide deposits (e.g., Morrison et al. 1994; Morrison 1999; Golightly and Rousell 2000; Naldrett 2004; Ames and Farrow 2007). A convenient classification is as follows: 1) SIC–footwall contact deposits; 2) footwall vein deposits;

3) offset dike deposits; and 4) sheared deposits. Although the SIC is strongly differentiated, it does not appear to contain any stratiform “reef”-type deposits (Lesher and Keays 2002). Approximately two thirds of the deposits are located at or below the base of the SIC (Table 12.1, deposit type 1 + 2) with the bulk of the remaining one third in offsets. Although there are approximately ten times as many SIC–footwall contact deposits as footwall vein deposits, the relatively high Ni-Cu-PGE grades in the latter deposits (Table 12.2) greatly enhances their value. Although deposits in concentric offsets comprise a small fraction of the total number of deposits associated with the SIC, the pre-mining resource of the giant Frood and Stobie deposits is disproportionately large.

Table 12.1. Number of deposits, in size categories, of Sudbury Igneous Complex (SIC) deposit types (*from* Golightly and Rousell 2000). Deposit types: 1) SIC–footwall contact; 2) footwall veins; 3a) radial offsets; 3b) concentric offsets; 4) sheared deposits. Size categories: Giant = several times larger than major, Major = orebodies that have been mined or have undergone mineral reserve calculations, Minor = undeveloped.

Deposit Type	Giant	Major	Minor	Total
1 + 2	Creighton	34	38	73
3a		17	17	34
3b	Frood, Stobie	3	3	6
4		3	-	3
				116

Table 12.2. Average grades in SIC–footwall contact deposits and footwall vein deposits (*data from* Naldrett et al. 1999). SIC = Sudbury Igneous Complex, Massive = > 20% sulphide mineralization, NR = North Range, SR = South Range, 1 = bornite-type veins, 2 = chalcopyrite-type veins, N = number of samples.

	SIC–Footwall Contact				Footwall Vein	
	Massive		Disseminated			
	NR	SR	NR	SR	1	2
wt % sulphide	49.16	47.43	7.66	13.03	15.59	46.82
wt % in sulphide						
Ni	4.78	4.82	5.27	4.94	6.97	6.27
Cu	1.83	4.64	5.77	7.31	49.6	27.3
S	38.0	37.6	37.43	37.3	24.6	34.3
Ni/Cu	2.7	1.0	0.9	0.7	0.1	0.2
ppb in sulphide						
Ir	41	130	65	152	4.5	52
Rh	115	461	151	420	10.5	54
Ru	75	246	208	279	181	104
Os	19	53	50	65	94	128
Pd	754	3457	2584	4407	12 209	15 770
Pt	2398	2503	3668	3898	40 435	30 578
Au	642	3689	2278	2211	5629	1475
N	666	490	334	233	73	315

SIC–FOOTWALL CONTACT DEPOSITS

These deposits occur at or near the contact between the base of the SIC and the underlying footwall rocks. The mineralization occurs as disseminated sulphides in the Contact Sublayer, which increase in concentration downward, and as massive to semi-massive sulphides along the Sublayer–footwall contact or in the underlying Footwall Breccia.

The main sulphide minerals include pyrrhotite (0.2 to 1% Ni) with lesser amounts of pentlandite, chalcopyrite and magnetite, and only small abundances of platinum-group elements. Monoclinic pyrrhotite is more abundant than the hexagonal type. Pentlandite occurs as equant to tabular grains and as exsolved flames in pyrrhotite. Chalcopyrite is present as stringers and veinlets. Mineralization ranges from fine to coarse (blebby) disseminated through semi-massive to massive.

South Range Deposits

Most of the SIC–footwall contact deposits in the South Range, such as the Little Stobie Mine, are hosted by the Sublayer (Figure 12.1). The Sublayer contains abundant, predominately mafic inclusions in a poikilitic to diabasic quartz diorite to norite matrix. Most inclusions are derived from the adjacent footwall except for a small suite of exotic ultramafic xenoliths, which are most abundant near mineral deposits (Photo 12.1). The ultramafic inclusions may have been derived from a deep, unexposed, pre-SIC ultramafic body or from an early cumulus phase of the SIC (Morrison et al. 1994). The thickness of the Sublayer (0 to 700 m) is controlled by the topography of the SIC–footwall contact. It is thickest in footwall embayments or troughs and is thin to absent outside these structures.

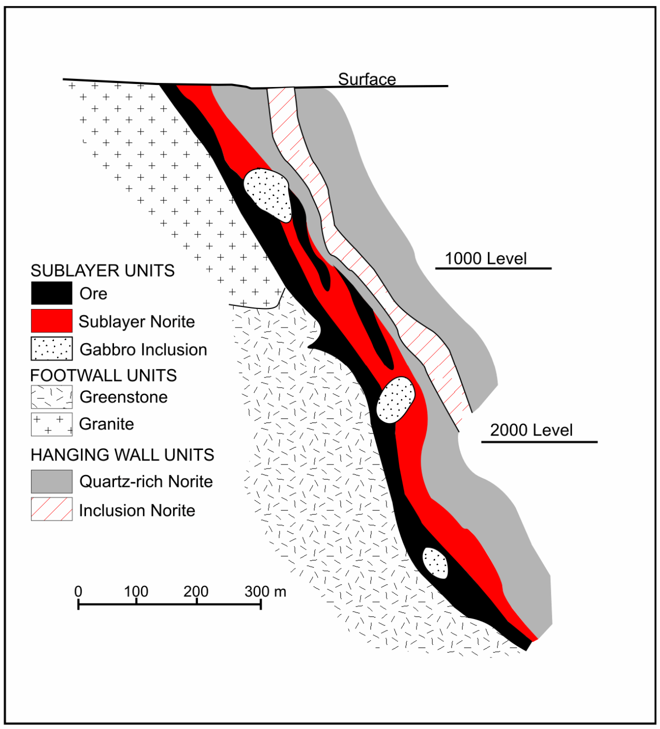


Figure 12.1. Cross-section of the Little Stobie No.1 orebody (South Range) looking west. Level depths are in feet. (*From* Davis 1984.)

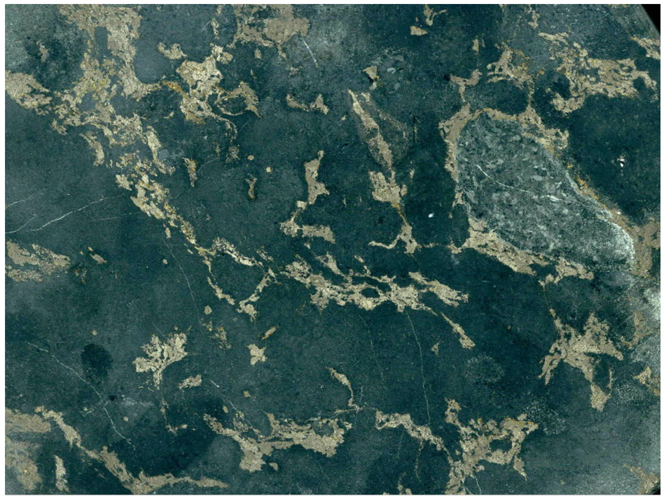


Photo 12.1. Sublayer with sulphide matrix (Creighton Mine). The coarse-grained pyroxenite inclusion is approximately 10 cm wide.

Embayments may be long (roughly 7 km), broad (900 m to <75 m) and relatively shallow (about 1.8 km) such as the Onaping–Levack embayment in the North Range (see maps and cross-sections in Coats and Snajdr 1984). Other embayments, such as that occupied by the Creighton Mine (Figure 12.2), the largest SIC–footwall contact deposit, are steep-sided, as deep as 1 km, and extend more than 4 km down-dip. Note that the smaller Gertrude deposit occupies the upper 2 km of a relatively shallow tributary embayment that plunges northeast and joins the Creighton embayment at a point about 3 km down the main embayment.

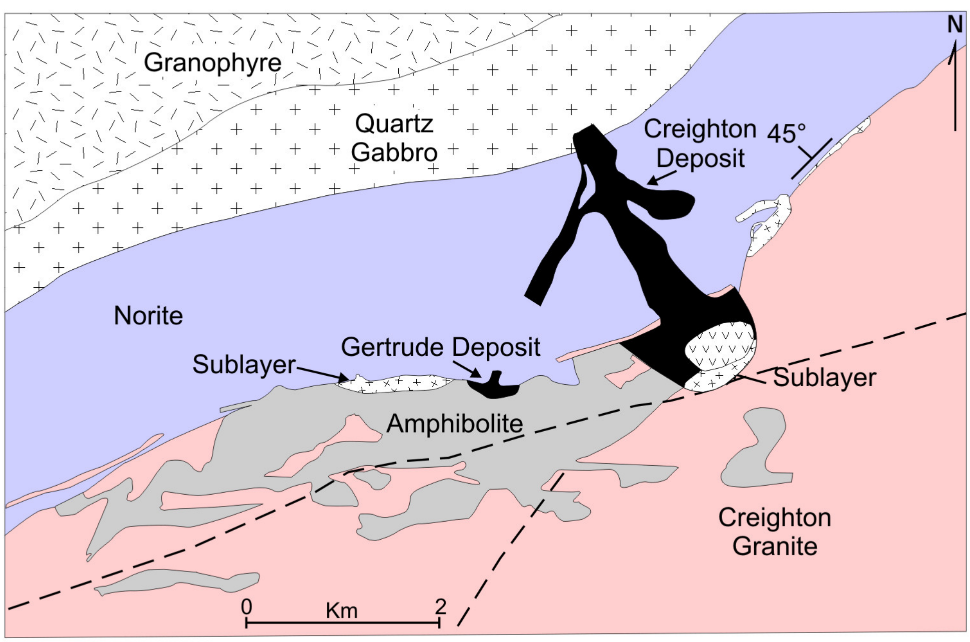


Figure 12.2. Map showing the surface projection (black) and original exposure (V-pattern) of the Creighton orebody (South Range), the largest SIC–footwall contact deposit (*after* Golightly and Rousell 2000). Note the asymmetrical V-shape of the orebody and the location of the Gertrude Mine.

Morrison (1984) suggested that the embayments represent radial channels for ejected material, analogous to the troughs in lunar craters. Smaller (on the scale of 100 m) depressions have been interpreted as terraces formed by slumping of the crater wall (Morrison 1984) or as smaller-scale channelways (Golightly 1994). However, it is likely that the original topography was significantly modified by post-impact thermomechanical erosion processes (Prevec and Cawthorn 2002). The ores occur as elongate, tabular bodies up to 150 m thick, which are localized in smaller, second-order depressions within the embayments and in the lowest levels of the embayments.

North and East Range Deposits

Most of the SIC–footwall contact deposits in the North and East ranges, such as McCreedy West (Figure 12.3), are hosted by Footwall Breccia which underlies, but is closely related to, the Sublayer. The breccia contains felsic and mafic fragments, derived from the underlying Levack Gneiss Complex, plus fragments of Sudbury Breccia, and pyrrhotite-rich sulphides (Photo 12.2).

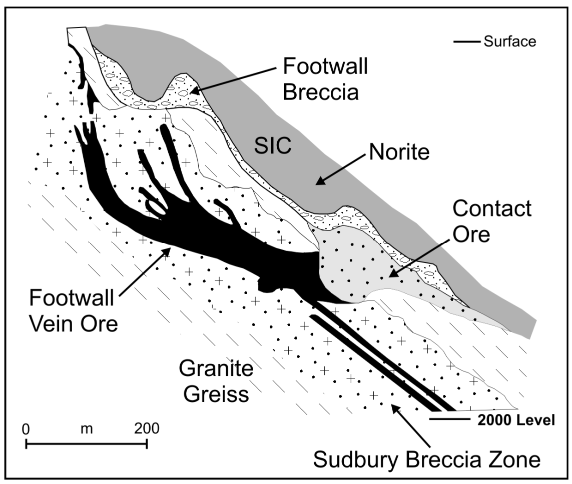


Figure 12.3. Cross-section of the McCreedy West Mine (North Range) looking east (*after* Morrison et al. 1994). Note the continuity between contact ore and footwall vein ore. SIC = Sudbury Igneous Complex. Level depth is in feet.

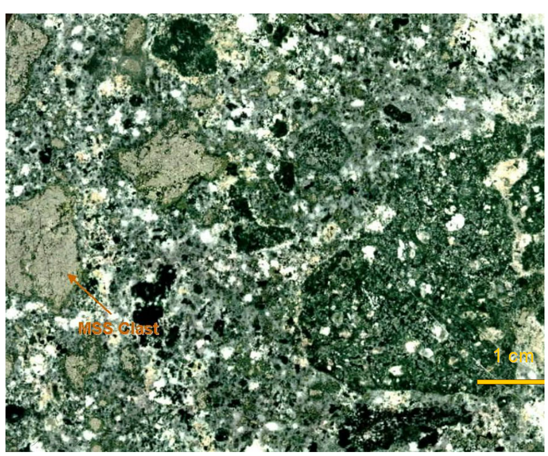


Photo 12.2. Footwall Breccia ("granite breccia") in drill core from the Levack area. A massive sulphide clast consists of pyrrhotite, pentlandite and orange chalcopyrite.

Morrison et al. (1994) suggested that at an early stage of the Sudbury Event, Footwall Breccia-hosted contact mineralization accumulated in depressions along the contact between the Sublayer and the brecciated footwall. Thermal metamorphism by the overlying SIC caused local melting of the breccia matrix, resulting in displacement of the molten matrix by sulphide melt, collapse of the breccia fabric and break-up of partially crystallized Sublayer, sulphide, Sudbury Breccia and Footwall Breccia. Gregory (2005) showed that there are systematic variations in the style, structure, texture and mineralogy of Footwall Breccia-hosted mineralization with proximity to the SIC. A zone-refining process involved progressively less thermal erosion, progressively more mechanical erosion, progressively deeper infiltration of sulphides and an increase in the width of the thermal aureole as the SIC cooled (see discussion by Prevec and Cawthorn 2002). Alternatively, the sulphide clasts may have been derived from pre-SIC orebodies (Pattison 1979).

FOOTWALL DEPOSITS

Mineralization in the footwall occurs as veins and disseminations as much as 700 m from the base of the SIC. There are several subtypes of mineralization (Farrow and Lightfoot 2002), as follows.

1) “Sharp-walled” veins of chalcopyrite + pentlandite ± millerite ± cubanite, which grade into thinner, less common veins of type 2.

2) Bornite ± millerite ± pentlandite ± quartz ± carbonate. Type 2 veins are further from the footwall contact and/or are at the peripheral parts of type 1 veins.

3) Au-PGE-rich, sulphide-poor disseminated mineralization at the most peripheral parts of the footwall systems.

4) “Hybrid” mineralization that appears to be transitional to the other types.

There is considerable evidence of hydrothermal activity in and around these systems, including the occurrence of epidote, pyrite, quartz-bearing chalcopyrite veins, and complex hypersaline fluid inclusions in quartz and chalcopyrite (Li and Naldrett 1993; Morrison et al. 1994; Farrow and Watkinson 1999; Molinar et al. 2001; Hanley et al. 2004). However, it is still uncertain as to what proportion of the mineralization is magmatic or hydrothermal or some combination of the two.

The greatest development of footwall vein deposits is in the Levack embayment on the North Range (Figures 12.3, 12.4, 12.5), in the Victor Deep–Nickel Rim environment in the East Range, and in the Whistle–Parkin environment at the northeast corner of the basin. In detail, the veins tend to form complex networks and stockworks that are usually restricted to zones of Sudbury Breccia that lie sub-parallel to the overlying SIC–footwall contact. The geometry of the vein system seems chaotic in detail, although there are some clear conjugate patterns in the larger (kilometre-scale), spatial distribution of the vein systems (Abel 1981). In the McCreedy West deposit (Figure 12.3) there is a gradual transition from disseminated pyrrhotite-pentlandite-chalcopyrite contact mineralization, hosted within Footwall Breccia in a footwall depression or terrace, into massive chalcopyrite-cubanite-pentlandite hosted by Sudbury Breccia in a footwall vein system.

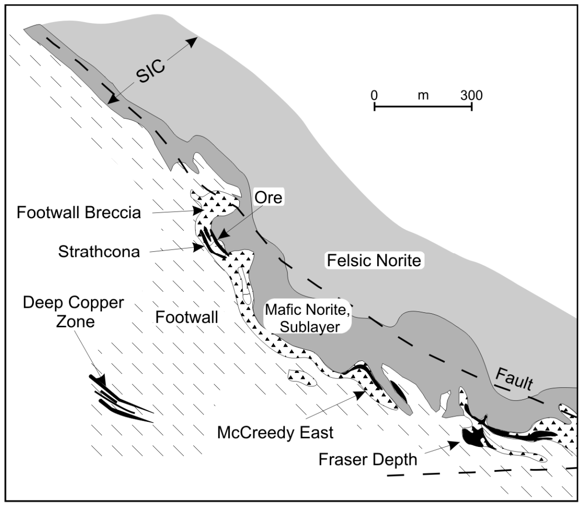


Figure 12.4. Cross-section in the Onaping–Levack area (North Range) looking east (*after* Coats and Snajdr 1984). The Strathcona Deep Copper Zone, a footwall vein deposit, has no known connection to SIC–footwall contact mineralization and is not hosted by Sudbury Breccia. Compare with Figure 12.3.

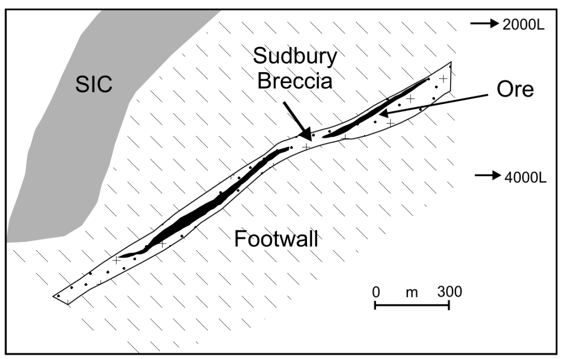


Figure 12.5. Cross-section of the McCreedy East Mine (North Range) looking west; an example of a footwall vein deposit (*after* Morrison et al. 1994). Level depths are in feet.

Footwall deposits also occur in the South Range but are much less common than in the North Range. There are small examples at Creighton Mine (Souch et al. 1969) and at Lindsley Mine. The vein systems in the North Range are laterally extensive, whereas those in the South Range are equidimensional.

Footwall deposits always occur below or near SIC–footwall contact deposits. This suggests that the former deposits were derived by the differentiation of the sulphide liquid that initially precipitated the latter deposits. In the North Range, the Ni/Cu ratio in contact deposits is 2.7, whereas it is only 0.1 to 0.2 in footwall deposits (Table 12.2). Presumably, Cu was removed from contact mineralization and deposited in the footwall veins (Naldrett et al. 1999). It is exceptional, however, to find a direct connection between the two deposit types. The McCreedy West deposit (Figure 12.3) is an example of a continuous connection, whereas a connection has not been established at the Strathcona Deep deposit (Figure 12.4).

Footwall deposits are not known adjacent to Offset deposits. However, Cu- and PGE-rich mineralization, with a strong chemical and mineralogical similarity to typical footwall veins, occur at the base of the Frood deposit (Zurbrigg 1957; Hawley 1962, 1965).

OFFSET DIKE DEPOSITS

The radial Offset dikes consist of fine-grained quartz diorite that was intruded in two phases. The first phase, preserved along the margins of the dikes, contains no sulphide minerals. The second phase, generally preserved in the dike cores but locally present along the margins, contains sulphide mineralization. Inclusions in the first phase also occur in the second phase, but this is rare (Photo 12.3). Coarse-grained, euhedral pyrrhotite, together with variable amounts of pentlandite and chalcopyrite, are disseminated in a fine-grained matrix of quartz diorite. This imparts a porphyritic texture to the rock (Photo 12.4). Angular fragments, consisting of quartz diorite with disseminated sulphide minerals, locally occur in massive sulphide ore.

The deposits of the Copper Cliff radial offset (Figure 12.6) tend to be located near changes in strike of the dike, constrictions in the dike or near cross-cutting faults (Cochrane 1984; Mourre 2000). Individual deposits commonly form steeply plunging lenses with typical dimensions of 1 x 0.2 x 0.1 km.

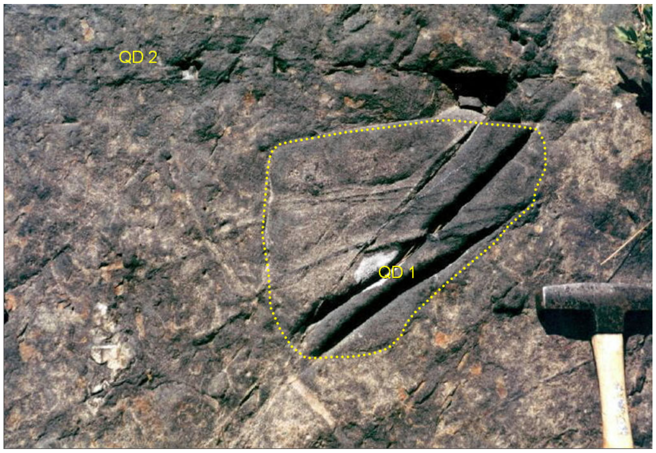


Photo 12.3. Barren quartz diorite inclusion (QD 1) in mineralized quartz diorite (QD 2) near the headframe of the Totten Mine.

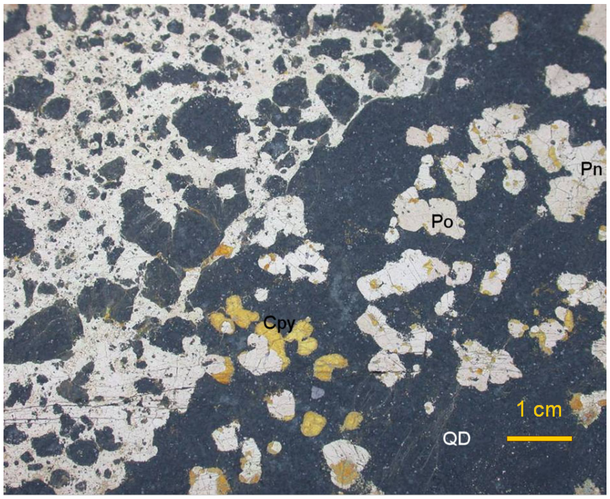


Photo 12.4. The transition from massive ore (upper left) to disseminated ore (lower right). The coarse-grained sulphide minerals, disseminated in a fine-grained matrix of quartz diorite (QD), impart a porphyritic texture to the rock (Frood Mine).

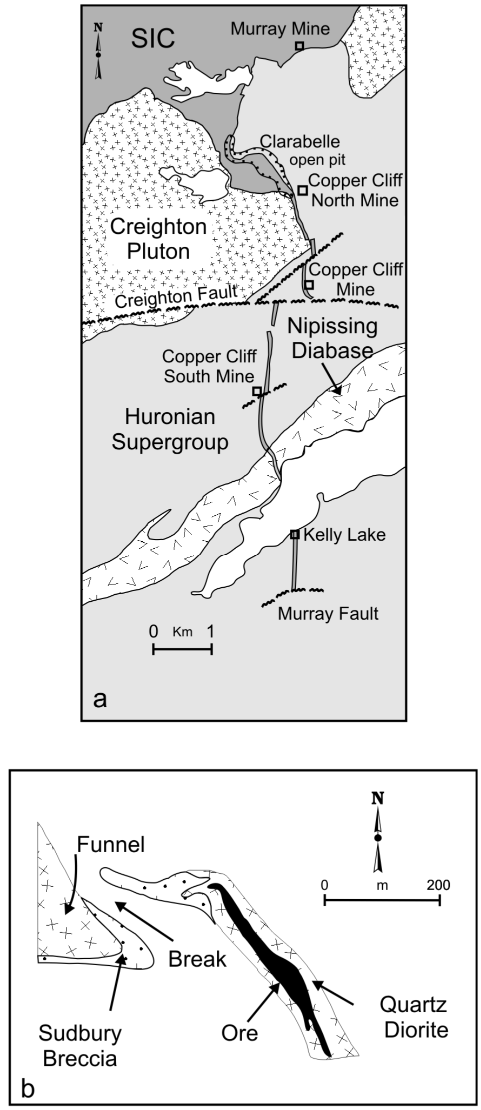


Figure 12.6. a) Geological map of the Copper Cliff radial offset dike. **b)** Plan of the 400-foot level, Copper Cliff North Mine, showing the break between the funnel and the quartz diorite dike (*after* Cochrane 1984).

The Frood and Stobie deposits, two of the largest in the Sudbury district, occur in a concentric offset. The mineralization is associated with quartz diorite bodies within a belt of Sudbury Breccia known as the South Range Breccia Belt (see Chapter 7). Each deposit occupies an inclusion-filled, quartz diorite tongue with individual surface dimensions of 1000 x 50 to 300 m and dip of 60° NW. The deposits taper down-dip for approximately 1 km (Figure 12.7). There is a gradual downward transition (Zurbrigg 1957) from blebby, disseminated pyrrhotite-chalcopyrite-pentlandite (in the ratios 6.4 : 0.88 : 1.0) to massive pyrrhotite-chalcopyrite-pentlandite (5.5 : 1.2 : 1.0) in the margins and deeper portions of the deposits. Massive sulphides in the Frood Mine grade into inclusion-filled massive cubanite-chalcopyrite-pyrrhotite-pentlandite (5.4 : 5.2 : 1.8 : 1.0). Further down, the deposit splits into two roots comprising a silica-rich zone of cubanite-chalcopyrite-pyrrhotite-pentlandite ore with a peripheral sheath, up to 30 cm wide, of maucherite-niccolite ± gersdorfite-pyrrhotite-chalcopyrite with high concentrations of sperrylite and other platinum-group minerals. In the Stobie Mine the ores are richer in pyrrhotite than in Frood Mine, disseminated sulphide minerals predominate, and the massive cubanite and siliceous ore zones are absent.

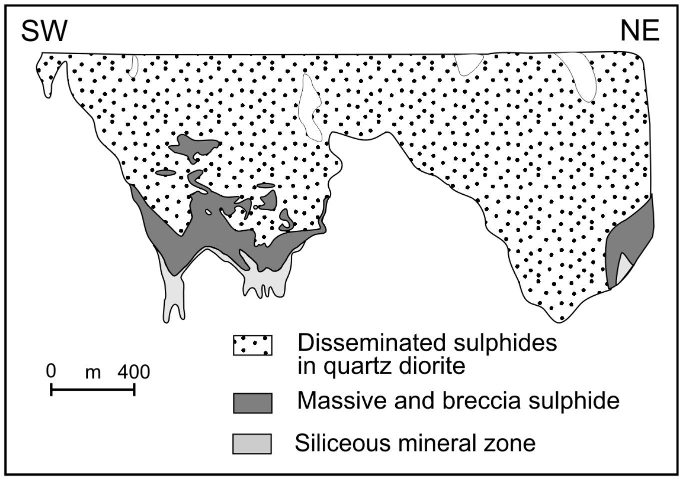


Figure 12.7. Vertical longitudinal section of the Frood (southwest) and Stobie (northeast) mines (South Range); an example of concentric offset deposits (*after* Zurbrigg 1957).

In some offsets, the transition from disseminated to massive sulphide ore is marked by a narrow zone of round quartz diorite blebs, from 0.5 to 5 cm in diameter, in the massive ore (Photo 12.5). According to Hawley (1962, 1965) the feature represents liquid quartz diorite suspended in liquid sulphide. Although uncommon, the texture demonstrates that primary magmatic sulphide/silicate intergrowths may survive the effects of metamorphism and sub-solidus sulphide recrystallization. Hawley (op.cit.) regarded the texture as decisive evidence for the magmatic origin of the sulphides, thus ending a 70 year controversy on the origin of the deposits.

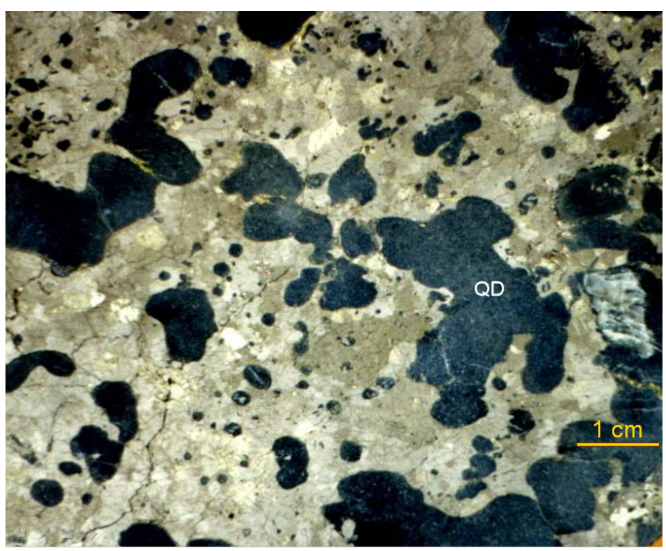


Photo 12.5. Quartz diorite (QD) blebs in a massive matrix of pyrrhotite and pentlandite (Copper Cliff South Mine). Hawley (1962) described similar material from Frood Mine.

SHEARED DEPOSITS

SIC–footwall contact deposits in the eastern third of the South Range were strongly affected by ductile shear. The Falconbridge deposit best illustrates the form of the sheared deposits. The original deposit, which was probably more irregular in shape, has been remobilized to form tabular bodies localized along the shears (Figure 12.8). Deeper in the Falconbridge Mine, the mineralized zone is confined by three faults (Figure 12.9). Marcasite is a locally important mineral in the west end of the Falconbridge deposit. Gersdorfite is also locally important in the Cu-rich ores, which occur entirely in the footwall. See Bailey et al. (2004) for an analysis of shearing in the Lindsley Mine (15 km west-southwest of Falconbridge Mine).

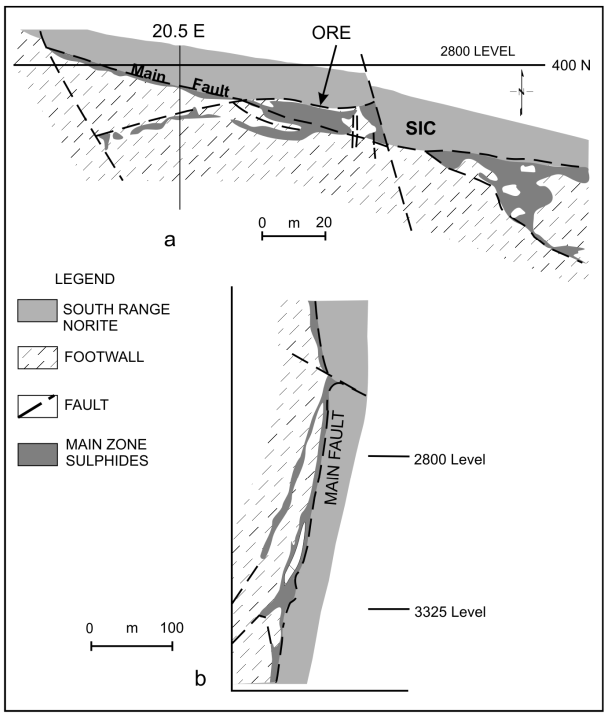


Figure 12.8. a) Geological map of the 2800 foot level and **b)** vertical cross-section (20.5 E) looking west of the Falconbridge Mine (South Range); an example of a sheared deposit (*after* Owen and Coats 1984). Level depths are in feet.

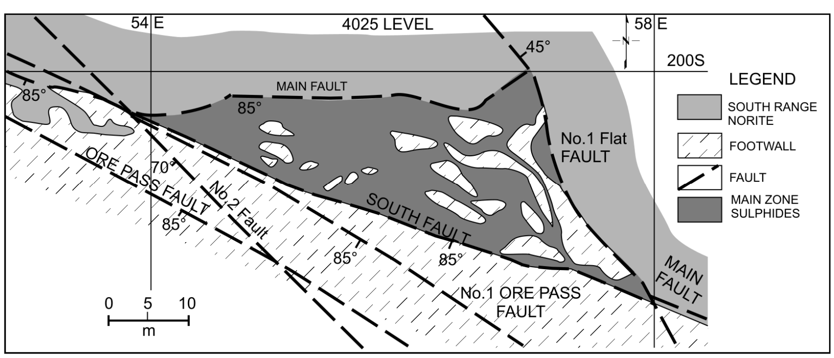


Figure 12.9. Geological map of the 4025 foot level of the Falconbridge Mine showing a wide sulphide zone confined by the Main Fault, the Number 1 Flat Fault and the South Fault (*after* Owen and Coats 1984).

Chapter 13

Mineralization in the Whitewater Group

D.H. Rousell, J.J. Paakki and M.J. Gray

Introduction

The rocks of the Whitewater Group host a variety of mineral occurrences including two former base metal mines (Rousell 1984c, 1984d). The mineralization is diagrammatically represented in a columnar section (Figure 13.1) and may be grouped as follows: 1) disseminated sulphide minerals in the Onaping Formation; 2) sulphide minerals in the Vermilion Formation (Vermilion and Errington mines); 3) pyrite and metals in the Onwatin Formation; 4) anthraxolite veins; and 5) mineralized quartz veins. See Figure 13.2 for the location of occurrences mentioned in the text.

Sulphide Minerals in the Onaping Formation

Disseminated sulphide minerals are present throughout the entire Onaping Formation. These minerals occur mainly as discrete fragments, but there are also sulphide patches and grains within rock and glass fragments and in igneous-textured bodies that occur within the formation. Several of the occurrences are exposed in trenches and pits. One is exposed in an adit 37 m in length ("Morley's Mine"). The sulphide minerals rarely exceed 10% and are mainly on the order of 1% by volume. The average nickel content of pyrrhotite, the main sulphide mineral, is 0.28 wt % (Desborough and Larsen 1970). Note that the average nickel content of pyrrhotite from the Sudbury Igneous Complex (SIC; Duke and Naldrett 1976) and from several Sudbury mines (Hawley 1962) is 0.81 wt % and 1.62 wt %, respectively. Other sulphide minerals include pentlandite, chalcopyrite, pyrite, marcasite, sphalerite and galena. Trace element data from the Onaping Formation are summarized in Rousell (1984c, 1984d). Bussolaro et al. (1984) made a detailed geochemical study of two areas in the North Range and concluded the following: the metal content is low and uneconomic; metal distribution is erratic and without stratigraphic control; and the bulk of the sulphide minerals are fragmental.

Sulphide Minerals in the Vermilion Formation

INTRODUCTION

James Stobie discovered Pb-Zn mineralization near the Vermilion River in 1897. The Vermilion Mine, the Sturdy property and the three shafts of the Errington Mine are located in the southwestern corner of the Sudbury Basin, in the vicinity of the discovery site (Figure 13.2). These Zn-Cu-Pb-Ag-Au deposits are hosted by the Vermilion Formation, a thin unit between the Onaping Formation and the overlying Onwatin Formation. The mines operated from 1926 to 1931 and from 1952 to 1956. Operations ceased due to low metal prices and low metal recoveries due to the fine-grained nature of the ore. The Errington and Vermilion deposits were described by Martin (1957) and Thomson (1957), prior to the current understanding of seafloor deposits.

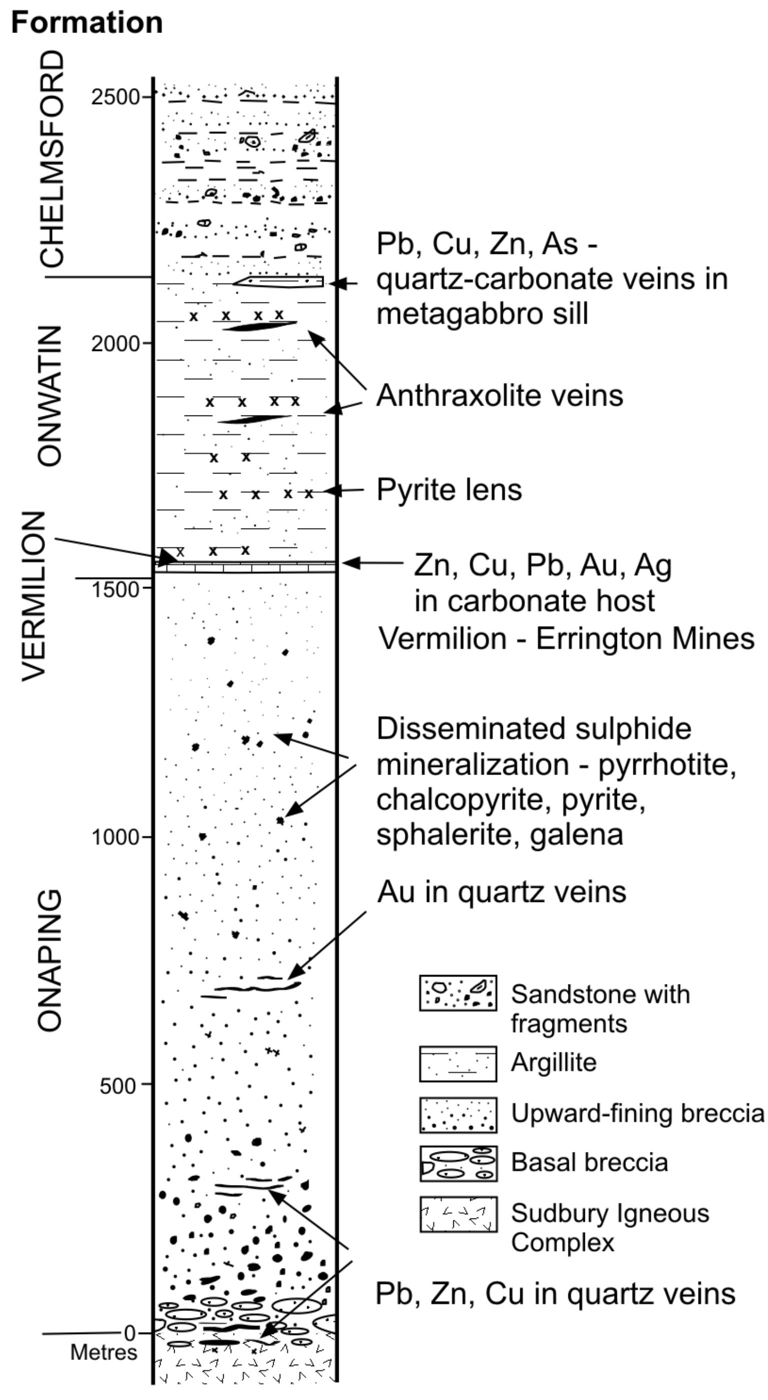


Figure 13.1. Diagrammatic representation of mineralization in the Whitewater Group (*after* Rousell et al. 2002).

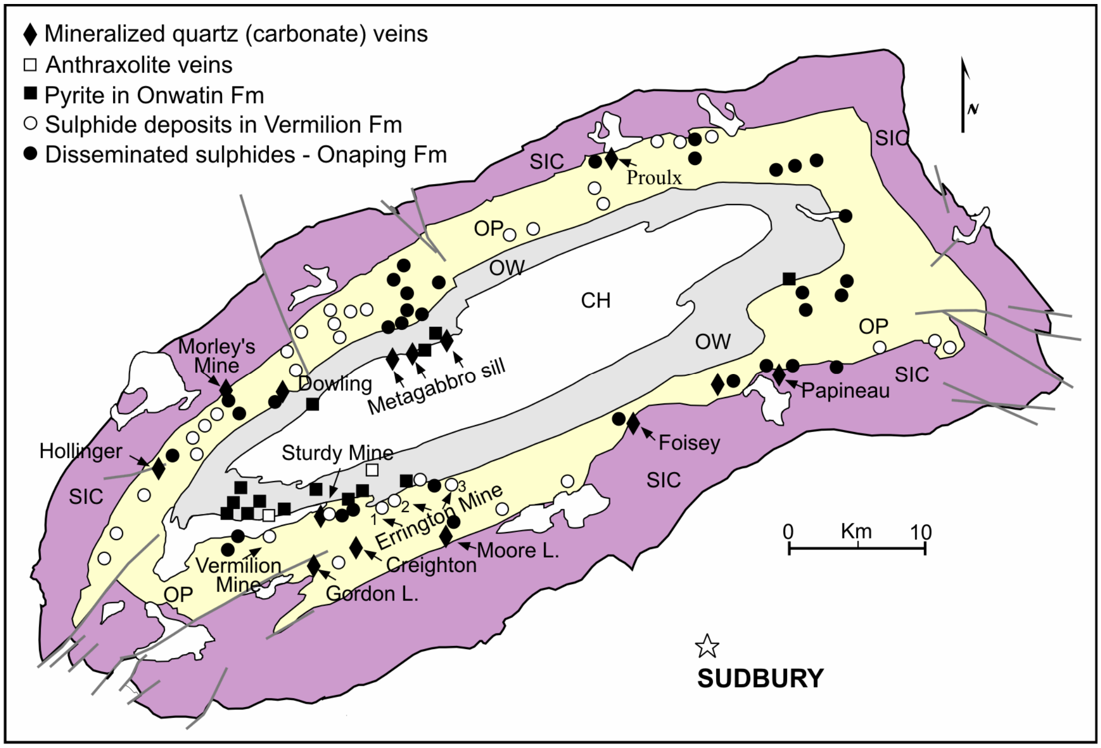


Figure 13.2. Geological map showing the location of mineral occurrences in rocks of the Whitewater Group (*after* Paakki 1992 and Rousell et al. 2002). SIC = Sudbury Igneous Complex; OP = Onaping Formation; OW = Onwatin Formation; and CH = Chelmsford Formation.

VERMILION MINE

The Vermilion deposit, discovered in 1929 by drilling, lies at a depth between 100 and 300 m. The deposit contains an inventory (proven-probable-possible) of 2.44 million tonnes grading 5.11% Zn, 1.49% Cu, 1.37% Pb, 1.10 g/T Au and 66.17 g/T Ag (Severin and Gates 1981). The main sulphide minerals, in order of abundance, are pyrite, sphalerite, chalcopyrite, galena, marcasite and pyrrhotite.

The deposit consists of seven mineralized lenses which strike between 030° and 060° (Figure 13.3a). Two of the lenses, No.4 (Figure 13.3b) and No.6, comprise 85% of the tonnage. The strata-bound lenses are hosted by the Lower Ca-Mg Carbonate member of the Vermilion Formation and the upper 5 m of the underlying Onaping Formation. Note that the carbonate host grades laterally into carbonates that are rich in Mn and Fe. The lenses dip southeast, occur within tight folds and are cut by axial plane faults (Martin 1957).

Early studies (Burrows and Rickaby 1930) suggested that the deposit represented irregular pods, scattered throughout the carbonate host, with the sulphide minerals derived from hydrothermal solutions emanating from the SIC. Later studies suggested that the deposits were structurally and stratigraphically controlled, that the host rocks may be sedimentary in origin with some or all of the carbonate and pyrite the product of hot-spring activity (Martin 1957; Thomson 1957). The stratigraphic succession of basal argillite, banded carbonate with sulphide-mineral layers, chert and black carbonaceous pyritic shale (Onwatin Formation) suggests that the Vermilion–Errington deposits may be related to the so-called Remac-type Pb-Zn deposits (Sangster 1970) and represent sedimentary-exhalative deposits (Rousell 1984c, 1984d).

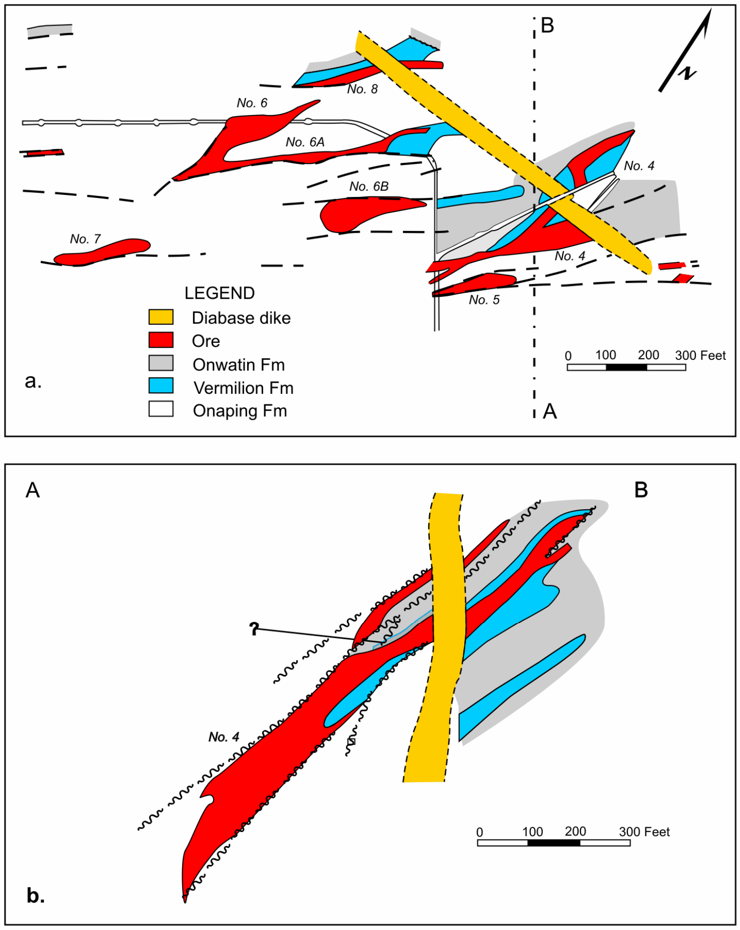


Figure 13.3. Vermilion Mine (*after* Martin 1957). **a)** Plan of the 600-foot (183 m) level. **b)** Vertical section (A-B), looking southwest, through the No.4 orebody. Wavy lines are faults.

Gray (1995) has set out the most recent and comprehensive study of the Vermilion deposit. His summary on the origin of the deposit is as follows (op.cit., p.iv).

> "The Vermilion deposit formed in a sub-seafloor environment by the initial low temperature pyrite-sphalerite infilling of carbonate porosity within sinters followed by later higher temperature chalcopyrite-rich mineralization that replaced earlier pyrite-sphalerite and secondary porosity from hydrothermal dissolution leaching of the carbonates. The carbonate sinter host is attributed to sub-seafloor boiling and carbonate precipitation by venting into an alkaline, anoxic basin and by successive replacement events that enabled the sinter mounds to grow. Carbonates were preserved from dissolution by the emplacement of distal turbidites. The hydrothermal system responsible for the Vermilion deposit was long-lived, the underlying SIC, or precursor intrusion provided the thermal energy required to drive and sustain convection cells within the Onaping formation aquifer. The Vermilion deposit shares some of the features of Cu-Zn-Pb fragmental-dominated VMS deposits such as the Mattabi deposit in Ontario, the Falun deposit in Sweden and some features of the modern seafloor sulphide environment in the Guyamas Basin" (Gulf of California).

ERRINGTON MINE

The Errington sulphide body occurs within the basal portion of the Vermilion Formation. The formation lies conformably between the underlying carbonaceous tuffs of the Onaping Formation and the overlying carbonaceous argillites of the Onwatin Formation (Figure 13.4). According to Severin and Gates (1981) the Errington deposit contained 4.36 million tonnes grading 4.05% Zn, 1.24% Cu, 1.05% Pb and 55 g/t Ag.

The sulphide mineralization consists of multiple lenses of grossly stratiform, pyrite-dominant, semi-massive and massive sulphides hosted by carbonates over a strike length of 3 km. Sulphide lenses are underlain by discrete feeder zones that consist of chalcopyrite and pyrrhotite stringers, silicified rock and iron carbonate alteration. Prominent enrichments in Ba, Zn and CO_2 and depletions in MnO occur in a zone, greater than 0.5 km wide, which surrounds the mineralization. The deposit is interpreted to have formed on the seafloor such that early pyrite and marcasite mineralization was subsequently replaced by higher temperature assemblages including chalcopyrite, sphalerite, pyrrhotite and galena. Silicified rocks overlie the deposit and act as an impermeable cap during mineralization. This inhibited mixing of ascending hydrothermal fluid and overlying seawater and promoted subsurface replacements of early carbonates (Paakki 1992).

Folding has modified the disposition of the mound- to blanket-like deposit. In the area of the No.1 and No.2 shafts, the mineralization occurs in the crest and the southern limb of an anticline (Figure 13.5) which plunges to the east and is overturned to the north. There, the deposit is displaced by south-dipping, axial planar reverse faults and a south-dipping thrust fault (Figures 13.4, 13.5).

The latest exploration work on the Errington Mine was conducted by Falconbridge Ltd. in the early 1990s with the aim of defining a larger deposit. A challenge faced in this geological setting is the lack of effective geophysical targeting methods, as the carbonaceous and highly conductive nature of the strata that encloses the sulphide lenses virtually preclude electromagnetic methods. However, geochemistry and detailed structural mapping can provide vectors to vent sites where sulphide lenses are thickest (Paakki 1992).

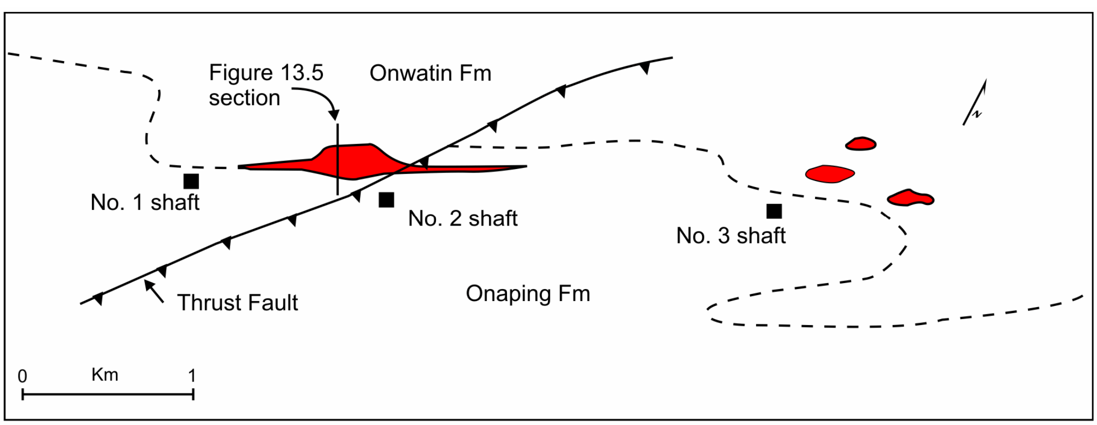


Figure 13.4. Map showing the location of the three shafts of the Errington Mine and a surface projection of the mineralized zones (red-filled areas). Dashed line marks the contact between the Onaping and Onwatin formations.

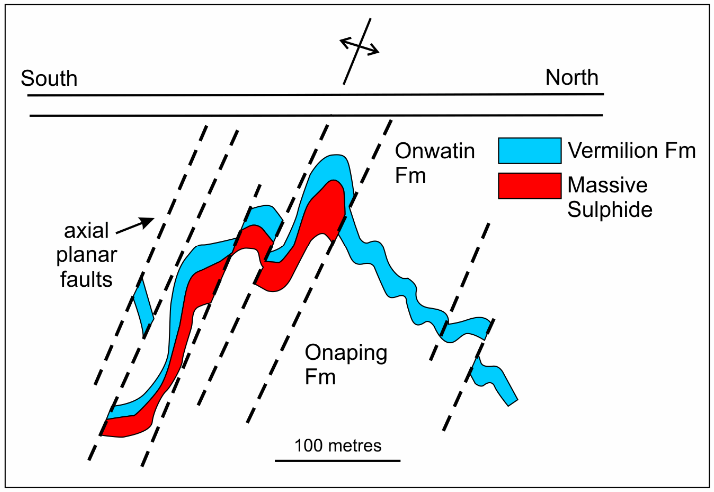


Figure 13.5. Cross-section, looking southwest, through the Errington Mine. See Figure 13.4 for the location of the section.

Pyrite and Metals in the Onwatin Formation

The Onwatin Formation is rich in pyrite. The pyrite occurs as silt-size grains arranged parallel to the bedding planes and as massive lenses, generally 1 to 3 cm thick, but locally as thick as 20 cm. Pyrite cubes, as large as 2 cm in diameter, form nodular masses and are the result of recrystallization. The presence of pyrite along cleavage planes indicates remobilization during metamorphism.

Trace element chemical data, for ten metals, are set out in Rousell (1984d) and compared to the average trace element content of shale (Krauskopf 1979). Particularly high relative values are present, in the formation, for Ag, Zn and Cu, with moderately high values for Ni, Co, Pb, Ba and As.

Anthraxolite Veins

Anthraxolite is a black, dense, lustrous, platy material that consists of approximately 95% carbon. Veins consisting of anthraxolite, together with some quartz and pyrite, are hosted by the Onwatin Formation and occur at two locations. One vein is located north of the Vermilion Mine and the other is north of Errington No.1 shaft. The latter vein is exposed in a 30 m long inclined adit and two smaller adits. Prior to the turn of the 20th century, there was considerable local interest in the anthraxolite as a possible fuel. However, this was precluded as the material does not burn well due to the high ash content (Burrows and Rickaby 1930).

Mineralized Quartz Veins

Sulphide-bearing quartz veins occur at the contact of the SIC and the Onaping Formation (Foisey and Papineau occurrences), near the base of the Onaping Formation (Moore Lake and Proulx occurrences, Morley's Mine and Hollinger) and near the top of the Onaping Formation (Dowling). Quartz veins, supposedly gold-bearing, are present in the lower part of the Onaping Formation (Gordon Lake Mine, Creighton Gold Mine). Quartz veins containing minor pyrrhotite and pyrite are reported from the northern end of the East Range (Rousell 1984c, 1984d). A metagabbro sill, located at the contact between the

Onwatin and Chelmsford formations, hosts mineralized quartz-carbonate veins. Trace element chemical data for veins and host rocks, unless referenced otherwise, are given in Rousell (1984c, 1984d).

At the Foisey occurrence, the most mineralized vein contains sphalerite (8.5 wt % Zn) together with minor amounts of galena and chalcopyrite. Quartz veins and blebs, up to 5 m thick, are numerous at the Papineau property. At the Papineau shaft, a 2.5 m thick quartz vein contains massive arsenopyrite and some pyrite as well as chalcopyrite, malachite and azurite in a mauve-coloured carbonate host rock.

At the Moore Lake occurrence, galena occurs as masses up to 5 cm wide, together with sphalerite, pyrite and chalcopyrite. According to Burrows and Rickaby (1930), a shaft was sunk at the Proulx occurrence to a depth of at least 20 m where two quartz veins contain appreciable amounts of sphalerite, galena and chalcopyrite. Quartz veins in Morley's Mine contain approximately 9% Zn, 12% Pb and 0.4% Cu (Lafleur 1981). Burrows and Rickaby (1930) reported the presence of stringers and lenses of sphalerite and galena, together with minor pyrite, pyrrhotite and chalcopyrite in a trenched gossan. Lafleur and Dressler (1985) refer to work on the property by Hollinger Consolidated Gold Mines Ltd. Near the town of Dowling, narrow, east-trending quartz-filled fractures contain chalcopyrite, sphalerite, pyrrhotite and pyrite (Lafleur and Dressler 1985).

The Gordon Lake and Creighton "gold mines" were abandoned around the turn of the 20th century (Blue 1894). The Gordon Lake site is, presumably, a prominent outcrop of quartz-rich rock located on the western side of the Gordon Lake Road. The 12 m thick mass lies within the Onaping Formation and contains numerous quartz veins and blebs as well as pyrite and limonite. At the Creighton property, where there is a shaft and a large dump of broken rock, quartz veins extend over a width of 12 m and can be traced for 100 m along strike. Chemical analyses of specimens from both properties failed to detect any gold.

A metagabbro sill, as much as 30 m thick, is exposed at three localities over a length of 4 km. The sill is the locus of numerous quartz-carbonate veins and blebs, up to 1.7 m thick, that are commonly intimately mixed with the gabbro. The veins have been explored by numerous trenches and pits. The carbonate is buff-coloured, weathered to a chocolate brown and is probably ankerite or siderite. Sulphide minerals include pyrite, arsenopyrite, chalcopyrite and malachite. Chemical analyses indicate local, but appreciable, Au, Ag, Cu and Zn values.

Mineralized Late-Stage Fractures

According to Cochrane (1991), late-stage fractures strike between west-northwest and north-northeast and are filled by calcite and/or secondary sulphides. They form zones that vary in width from several metres to several tens of metres. In the Garson Mine, fracture zones strike northwest, dip steeply east, and are filled by calcite, galena, marcasite and sphalerite. The zones are parallel to, within, or at the margins of olivine diabase dikes of the Sudbury Swarm (1238 Ma; Krough et al. 1987). Galena-filled fractures also occur in the Creighton Mine and in the Strathcona Mine. It is of interest to note that microseismic activity in the Sudbury mines, not directly related to mining, are associated with the late-stage fractures and faults (Cochrane 1991).

Origin of Mineralization

The mineralization in the Whitewater Group may be separated, in terms of time of formation, as follows (Rousell 1984c, 1984d): 1) fragmentary mineralization in the Onaping Formation; 2) mineralization that formed after deposition of the Onaping Formation and before deposition of the Onwatin Formation

(Vermilion–Errington deposits); 3) syn-depositional mineralization in the Onwatin Formation; and 4) mineralization formed during tectonometamorphism (anthraxolite, quartz veins, mineralized fractures).

Sulphide fragments in the Onaping Formation may be explained, in terms of the meteorite impact theory, as follows. The impact of the bolide brecciated the target rocks, which apparently included some sulphide-rich bodies. This material became airborne ejecta that later fell back into the crater and was incorporated into the Onaping Formation.

Burrows and Rickaby (1930) and Card and Hutchison (1972) suggested that the mineralization in the Vermilion and Errington mines was derived from hydrothermal solutions emanating from the SIC. These solutions were at a lower temperature than those, also derived from the SIC, which formed the Ni-Cu-PGE deposits. Further studies (Martin 1957), after the reopening of the mines in 1952, recognized that the deposits were stratigraphically and structurally controlled. The ore occurs mainly on the south limb of folds that are cut by reverse faults. Brecciation by the faults prepared the rocks for epigenetic mineralization. Much of the chert and carbonate host rocks, as well as pyrite, may be the product of hot-spring activity during the last phase of Onaping volcanism (Martin 1957). Rousell (1984c, 1984d) regarded the Vermilion–Errington mineralization as sedimentary exhalative deposits that formed within the framework of the impact model. The impact site was a shallow sea. Some seawater penetrated to the bottom of the deep transient crater and became trapped by the fall-back breccia. The outer rim of the final crater gave rise to a closed basin which was much larger than the present Sudbury Basin. The abundant carbonaceous material, perhaps derived from floating algal mats, suggests that bottom waters were stagnant and anoxygenic. The trapped brines leached metals as well as silica and carbonate material from the bedrock, rose upward, deposited some Zn, Pb and Cu as they passed through the Onaping Formation and, on reaching the basement floor, precipitated the Vermilion Formation. Relatively recent detailed studies (Paakki 1992; Stoness 1994; Gray 1995) concluded that ascending hydrothermal solutions, derived from seawater or possibly from the SIC, transported the sulphides which then replaced carbonate units. Pyrite and marcasite apparently formed initially, followed by higher temperature assemblages including chalcopyrite, sphalerite and pyrrhotite.

The black, carbonaceous shale of the Onwatin Formation, with the syn-sedimentary pyrite, implies deposition in a deep-water euxinic environment rich in H_2S. The relatively high metal values, compared to average values in shale, may be the result of syngenetic or hydrothermal processes (Rousell 1984c, 1984d).

In the South Range, the rocks of the Onaping Formation are characterized by a northeast-striking, southeast-dipping tectonic foliation that formed during the Penokean Orogeny (1900 to 1700 Ma, Bennett et al. 1991). The numerous quartz veins in the South Range are oriented parallel to the foliation, whereas in the North Range, where the foliation is weak or absent, veins are scarce. Tectonometamorphism remobilized and concentrated carbonaceous material in the Onwatin Formation to form the anthraxolite veins.

The age of the mineralization in the late-stage extension fractures and the origin of the solutions that transported the sulphides is not exactly known.

PART 2 - SELF-GUIDED FIELD TRIP

Introduction

The stops are arranged for convenience of travel rather than in order of any particular rock type, formation or age. Presumably, it would require several days to visit all the stops. A visit to Science North and Dynamic Earth would require at least a half day. The amount of descriptive detail varies, depending on the stop. As this is a self-guided field trip, directions to each stop are set out in some detail. Distances are given in kilometres (km) and should be regarded as approximate. The starting point is located on Ramsey Lake Road, near Laurentian University. All stops are plotted on a geological map, scale 1:100 000 (in pocket).

Universal Transverse Mercator (UTM, NAD 83) locations are given for each stop. The UTMs are reported as Eastings (E) and Northings (N). Northings are always displayed in GPS units as seven digits (e.g., 5162266), but Eastings may be displayed as six digits (e.g., 467205) or seven digits (e.g., 0467205), depending on the make or model of the GPS unit. In this field guide we report Eastings in seven digits, therefore, if your unit displays Eastings in six digits, you will need to drop the initial "0" from all Eastings when entering them into your GPS unit.

All the stops in this guide can be located on four topographic maps published by Energy, Mines and Resources Canada (scale 1:50 000), which show the UTM grid, the Eastings and the Northings. These are: Chelmsford (41-I/11), Capreol (41-I/10), Coniston (41-I/7) and Copper Cliff (41-I/6). These maps may be obtained from Ontario Map Company, 463 Clinton Ave., Sudbury, Ontario, P3B 2S5, tel. 705-673-2103, fax 673-3161 and e-mail ontmapco@sympatico.ca. We recommend the purchase of the road map of Greater Sudbury published by MapArt and available at many service stations, pharmacies and stores. The most recent geological compilation map (scale 1:50 000) of the Sudbury area is Open File 4570 (Ames, Davidson et al. 2005) published by the Geological Survey of Canada at $30.00. Contact GSC Bookstore, 601 Booth St., Room 107, Ottawa, Ontario, K1A 0E8, telephone (toll-free): 1-888-252-4301.

The Ontario Geological Survey (OGS) 1:50 000 geology compilation map 2491 (1984) is a bargain at $4.40 and still shows the basic geological elements. OGS Guidebook 8 (Dressler, Peredery and Muir 1992) includes a handy, coloured geological map (scale 1:100 000) at a cost of $8.00. OGS Special Volume 1 (1984) "*The Geology and Ore Deposits of the Sudbury Structure*" remains a classic compendium of Sudbury geology ($16.50). OGS Special Volume 6 (2002) "*The Physical Environment of the City of Greater Sudbury*" contains overviews of a number of topics including bedrock geology and mineral deposits, Quaternary geology, geotechnical properties, hydrogeology and the effects of the mining and smelting industry on the Sudbury landscape ($33.00). The volume also contains three 1:125 000 scale maps of bedrock geology, surficial geology and lakes, rivers and wetlands. OGS material may be obtained from: Publication Sales, Ministry of Northern Development, Mines and Forestry, Willet Green Miller Centre, 933 Ramsey Lake Road, Sudbury, ON P3E 6B5, telephone (toll free): 1- 888-415-9845, ext. 5691. Direct purchases may be made from the sales office located on the second floor of the Centre. Readers may also be interested in the volume "*Restoration and Recovery of an Industrial Region: Progress in Restoring the Smelter-Damaged Landscape near Sudbury, Canada*"; edited by John M. Gunn and published by Springer-Verlag, 1995.

Several stops are on mining or private properties. Prior permission is required in order to visit these sites, which should be obtained well in advance of a proposed visit. Contacts are as follows.

1. Stop 15. Anthraxolite. Call Marc Gascon, home telephone: (705) 524-5110, cell: 562-0847.

2. Stops 29, 30 and 31. Geology Department, Xstrata Nickel, Sudbury Operations, Falconbridge, ON, P0M 1S0. Telephone: 705-693-2761.

3. Stop 31. Coleman / McCreedy East Mine, Vale Inco Ltd., General Office, 1 Rink Street, Copper Cliff, ON, P0M 1N0 : Att. Coleman Mine Manager. Telephone: 705-692-2200.

Description of Field Trip Stops

Stop 1. Shatter cones (E0501540, N5146116)

Exit from the main entrance to Laurentian University (traffic lights) and turn left on Ramsey Lake Road. Stop 1 is approximately 0.75 km from the traffic lights, on the left (south) side of the road and opposite house #516.

The exposure displays shatter cones in sandstone of the Mississagi Formation (Huronian Supergroup). In the Sudbury area, shatter cones are best-developed in this formation. No hammering please.

Photo S1. Shatter cones.

Stop 2. Nipissing gabbro (E0498917, N5146421)

Continue on Ramsey Lake Road toward Science North, turn left on Paris Street then right on Walford Road and proceed to Regent Street. Go straight through the lights at Regent Street and on to Martindale Road. Drive about 0.8 km to the top of a prominent ridge.

The irregular contact between the gabbro and quartzite of the Mississagi Formation is exposed on the ridge behind "The Auto Doctors" shop. A clean exposure of the gabbro occurs on the east side of Martindale Road adjacent to house #998. Note shears, fractures and quartz veins and blebs.

Photo S2. Nipissing gabbro.

Stop 3. Clastic Sudbury Breccia (E0497081, N5144749)

Return, on Martindale Road, to the base of the ridge and turn right on Southview Drive. Continue on the Drive then turn right on Southview Drive where it intersects Bouchard Street. Continue along Southview Drive for about 1 km past the intersection with Kelly Lake Road (3-way stop sign). Park on open space on the right side of the road. Cross the road and walk about 10 m into the bush.

The breccia is developed in the Mississagi Formation. Note the large, round quartzite fragments in a clastic matrix with prominent flow surfaces.

Cross road back to vehicle and walk another 10 m or so into the bush on the south side of the road. Locate narrow, 4 cm wide dike of Sudbury Breccia with apophyses. Flow structures include sheath folds. To the northeast is an exposure of conglomerate of the Ramsay Lake Formation.

Photo S3. Clastic Sudbury Breccia.

Stop 4. Ramsay Lake Formation (E0494032, N5142216)

Continue on Southview Drive to the Southwest Bypass (Hwy 17) and turn right. Continue on the Bypass for about 2 km. The stop is on the right (north) side of the highway, about 200 m west of the sign for Walden Park.

This glaciated outcrop shows both massive (Photo S4.1) and laminated (Photo S4.2) facies of the diamictite. The presence of cross-stratification (Photo S4.2) in some of the sand-rich units confirms development of a thin layer of moving water beneath a wet-based floating ice-sheet or shelf. The thin mudstone layer (centre of sandstone) may represent deposition in a standing water body in winter.

Photo S4.1. Glaciated exposure of massive (top) to weakly laminated (base) clast-rich diamictite.

Photo S4.2. Cross-stratified granular sandstone and thin mudstone interbeds between clast-rich diamictites.

Stop 5. Rock cut in the McKim Formation (E0489918, N5141817)

Continue on the Southwest Bypass to a cloverleaf intersection between Hwy 17 and Regional Road 55. Stay on Hwy 17 West (4 lanes) for about 0.5 km to a prominent rock cut. Park with care.

The cut is up to 18.1 m high and approximately 435 m long. Note the presence of several fracture sets. Look for plumose structure.

Photo S5. Fractures in the McKim Formation.

Stop 6. Copper Cliff Formation (E0488885, N5141577)

Continue west on Hwy 17, take the first turnoff (unauthorized) and return to the cloverleaf intersection. Take Regional Road (RR) 55 west to RR 24 and turn right, toward Lively. Go under Hwy 17 overpass. The stop is immediately north of the overpass. Park on right side, opposite road to Walden Day Care Centre.

The exposure consists of rhyolite; recrystallization has obscured primary structures. The rhyolite is cut by several thin mafic dikes and is bounded, to the north, by Sudbury Breccia.

Photo S6. Recrystallized rhyolite of the Copper Cliff Formation.

Stop 7. Elsie Mountain Formation (E0488666, N5143229)

Continue north on RR 24, past Lively, and park opposite the water tower. The stop is on the west side of the road.

Mafic lava flows feature prominent, somewhat flattened, pillows with selvages (Photo S7.1), massive flows with flow-top breccias and inter-flow sediments (Photo S7.2).

Photo S7.1. Pillows with selvages in the Elsie Mountain Formation.

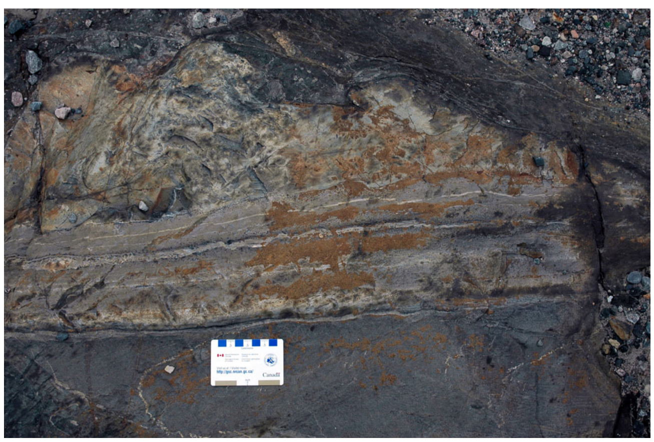


Photo S7.2. Thin, laminated inter-flow strata between a pillowed basalt (base) and a local breccia of chilled pillow selvages beneath a higher flow (top).

Stop 8. Creighton pluton (E0486228, N5144836)

Continue on RR 24. Turn at road to Creighton Mine and park along the sideroad with a plaque displaying the former village of Creighton Mine. Walk to the water tower by following the gravel road whose entrance is blocked by boulders.

Foliated, pink, coarse-grained granite is well exposed between the plaque and the water tower. Migmatite, consisting of granitic material with lenses and layers of black mafic rocks, is present immediately east of the tower. Locally, a 1 m thick dike of pseudotachylite Sudbury Breccia occurs at the contact with the granite.

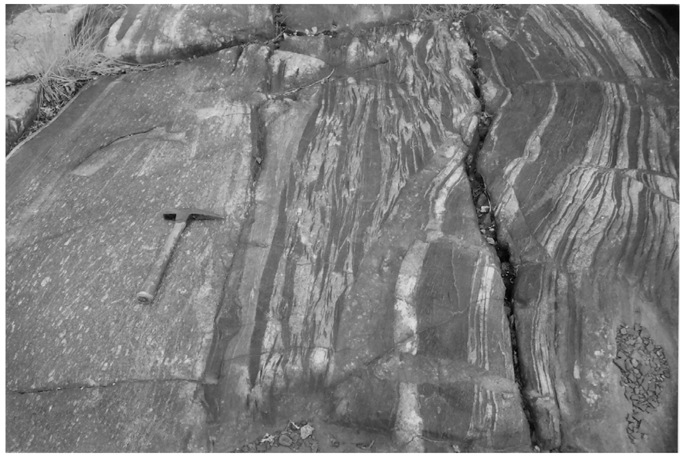


Photo S8. Contact between migmatite and foliated granite.

Stop 9. Pseudotachylite Sudbury Breccia and Creighton pluton (E0485931, N5144612)

Exit Creighton Mine road and turn right on RR 24 for 0.75 km.

Slightly foliated granite with pink feldspar crystals up to 4 cm across is cut by an excellent exposure of pseudotachylite Sudbury Breccia with both sharp (Photo S9) and undulatory contacts.

Photo S9. Sharp contact between pseudotachylite Sudbury Breccia and granite.

Stop 10. Porphyritic mafic dike (Creighton pluton) (E0485704, N5144505)

Continue west on RR 24 for 300 m and stop opposite a small lake (Pump Lake).

The remnants of a mafic dike are exposed on the south side of the road. The dike is replete with greenish, circular to ovoid feldspar crystals up to 10 cm across set in a black, fine-grained matrix. Locally, the mafic porphyry is cut by dikes of the surrounding pink, coarse-grained granite (Photo S10).

Photo S10. Porphyritic mafic dike cut by granite dike; Creighton pluton.

Stop 11. Gertrude Fault (E0484424, N5145048)

Continue west on RR 24 to Hwy 144, turn right and go north for 1 km (approximately 100 m north of turnoff to Gertrude Mine; there is no sign to the mine).

The fault is near the base of the SIC and is sub-horizontal, undulating and anastomosing with gouge as thick as 25 cm.

Photo S11. Gertrude Fault.

Stop 12. South Fold Nose (Vermilion Formation) (E0483638, N5154388)

Continue north on Hwy 144 for about 11 km then turn left (west) on Lavallee Road and go past Errington Street. Turn left (south) on Fire Route K (if you miss this turn you will pass the Colonial Golf Club). Go along Fire Route K for 1.3 km (past a cross road and a road on the right). Turn right (west) on a road at the base of a hill with rock outcrop. The road is wide, straight and may be rutted. Go along this road for about 1 km. If you pass a wrecked auto you have gone too far. Take the trail, on the right side of the road, which is marked by flagging. Walk along the trail (approximately north) for about 10 minutes to a creek. Cross creek to a large, 144 x 30 m mechanically cleared exposure.

The outcrop, known as the South Fold Nose (see Photo 9.2), is on the south limb of a kilometre-scale fold. Structural data (Chapter 9) suggest that the attitude of the axial surface, based on the mean attitude of cleavage, which is assumed to be axial planar, is 084° 82°S. The attitude of the hinge, in terms of the mean plunge and trend of small-scale folds, is 53°E 109°. Shortening estimates from small-scale folds, based on the assumption that they are buckle folds, range from 53 to 83%.

Individual beds of the Grey Argillite Member (Vermilion Formation) are easily traced. The dark-brown weathering of the argillite is due to Mn-staining (3.2 wt % MnO). Some beds have a 1 cm wide, irregular-shaped, black, non-conductive halo. Four slump horizons are visible in the northern portion of the outcrop, indicative of soft-sediment deformation and synsedimentary slope failure. Iron-carbonate laminations are common. They may be the product of CO_2-saturated hydrothermal solutions moving through the beds during burial diagenesis or from the deposition of detrital carbonate derived from the mounds in the Lower Carbonate Member. Two concretionary bodies (3 x 2 cm), presumably formed in mud, suggests that the top is to the southeast.

Stop 13. Mineralized metagabbro sill (E0483082, N5163337)

Return east along Lavallee Road, turn left on Errington Street and go north. Cross Hwy 144 into Chelmsford and turn left on Main Street, then right on Montpellier Road, past the nose of a plunging anticline (Chelmsford Formation). Cross the CPR tracks (note km value on gauge), go past the golf course and Wilderness Road to a locality near the Chelmsford–Onwatin formations contact and 5 km from tracks. Stop at power line and near house #5052. Cross road, and walk along the trail adjacent to and on the right of a fence which borders the property. Trail leaves the fence and veers west. Walk for five minutes to isolated ridge.

The sill, at the contact between the Chelmsford and Onwatin formations, is up to 30 m thick and is exposed at three localities over a length of 4 km. Abundant quartz-carbonate veins have been explored by numerous trenches and pits. Sulphide minerals include pyrite, arsenopyrite, chalcopyrite and malachite. Chemical analyses indicate local, but appreciable, Au, Ag, Cu and Zn values. Quartz veins are exposed on the ridge. There is a deep pit at east end of the ridge.

Stop 14. Anticline in the Vermilion Formation (E0479113, N5153737)

Return to Hwy 144, turn right and travel west for approximately 4 km, then turn left on Vermilion Lake Road. Go south past turn marked “Vermilion” then turn right on Stobie Dam Road (no sign). Go west for about 0.5 km. Note high fence on right, which is part of a mine reclamation initiative. Park and walk south through broken concrete barrier for about 100 m.

At the northeast end of the stop (Figure S14) argillite beds of the Vermilion Formation are cut by a cleavage and folded by an inclined, plunging anticline (Photo S14.1). Attitudes are as follows: hinge = 16°NE 068°; axial plane = 074° 70°S; northwest limb = 092° 57°N; southeast limb = 005° 18°E; and cleavage = 068° 80°SE. Note the cuspate-lobate folds in a thin argillite bed between two thicker beds. At the west end of the stop (Figure S14) hydrothermal products include carbonates, silicification, brecciation and minor sulphide mineralization (Photo S14.2).

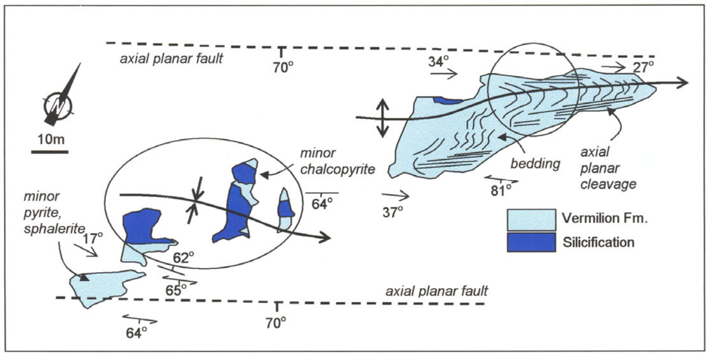


Figure S14. Geological map near the Errington No.1 shaft (Stop 14). Photo S14.1 is located within the circle at the northeast end of the map and Photo S14.2 is located within the circle at the west end of the map. The map area is located about 500 m west and up-plunge from the cross-section of Figure 13.5.

Photo S14.1. Plunging anticline near Errington No.1 shaft.

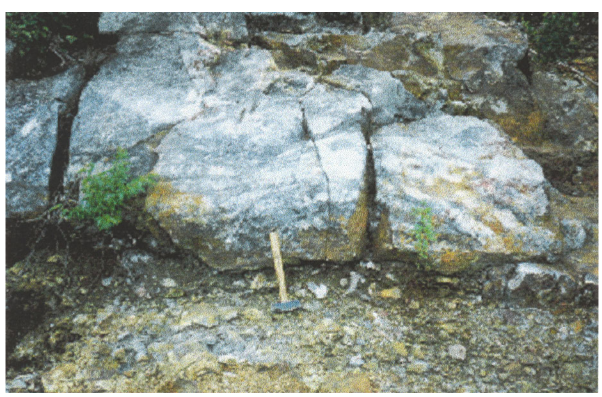


Photo S14.2. Crudely banded hydrothermal silicification and brecciation. The protolith was probably bedded argillite. Errington No.1 shaft area.

NOTE: PERMISSION IS REQUIRED TO VISIT THE NEXT STOP.

Stop 15. Anthraxolite (E0478143, N5155464)

Return east along Stobie Mine Road, turn north along Vermilion Lake Road, then turn left and travel west along Vermilion Lake Road, then turn right and go north along Davey's Road to the end. Walk straight ahead (north) along the wide trail for 4 minutes, then turn right on a trail for 2 minutes to quarry and adits.

The anthraxolite vein is exposed in a 30 m long inclined adit and two smaller inclined adits. Unfortunately, these adits are now blocked by fences. The Onwatin Formation hosts the anthraxolite and is well exposed at this locality. Note the prominent slaty cleavage.

Stop 16. Vermilion Formation (E0476276, N5153688)

Return to Vermilion Lake Road, turn right and go west to Gordon Lake Road. Turn left and go south. Cross the bridge over the Vermilion River and park at the end of the bridge on the left. Walk south, on east side of the road.

"Varved-looking, 1-2 cm thick bands of carbonate and grey argillite representative of the non-carbonaceous Vermilion Formation, and host to the Errington and Vermilion deposits, can be seen at this rare exposure" (Gibbins 1997, p.45, Stop 10). Note the rusty weathering.

Stop 17. Peperite in Dowling Member, Onaping Formation (E0474778, N5152349)

From the bridge over the Vermilion River proceed south for 2.2 km. Park on the left (east) side of the road at the hydro station and walk back to Stop 17, a roadcut on the east side of the road.

Amygdaloidal and spherulitic dikes of andesite intrude the upper units of the Dowling Member as does an amygdaloidal sill with a massive core. The sill is tightly folded with a subvertical axial planar cleavage.

Stop 18. Peperite in the Dowling Member (E0474700, N5152199)

Walk 80 m south, on west side of road, then cross a creek to a series of mechanically stripped exposures.

An extensive peperite body (175 x >1000 m), informally called the Gordon Lake Road peperite, occurs in the upper part of the Dowling Member. It lies 300 m below the Vermilion Mine and trends parallel to the foliation (060°) and subparallel to bedding. Subdivisions of mixing zones (Figure S18) are based on the proportions of andesite versus tuff, and on the texture. Mixing is most intense where the percentage of andesite is greatest. The tuff is silicified and light grey in colour, whereas the andesite is pale green and recessive weathering.

The rocks are intensely altered. Alteration minerals are similar to those in the overlying Vermilion deposit and include Ba-muscovite, Ba-K-feldspar, quartz, Ca-Mg-Mn carbonate minerals, chlorite, biotite and base metal sulphides.

The peperites formed in a shallow, subaqueous environment coeval with an active hydrothermal system that formed Zn-Pb-Cu mineralization. See Photo S18 for a selection of photographs for Stop 18.

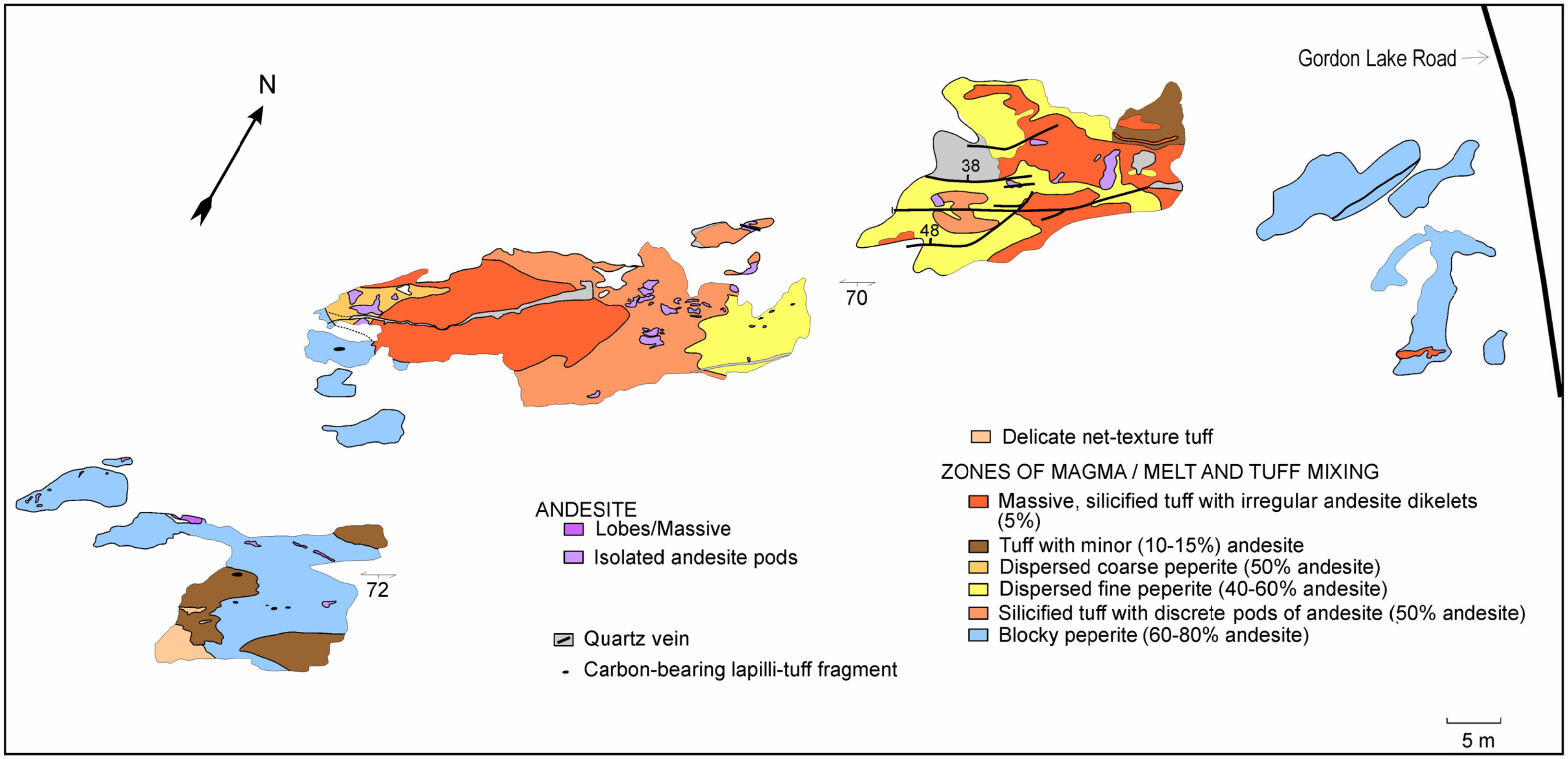
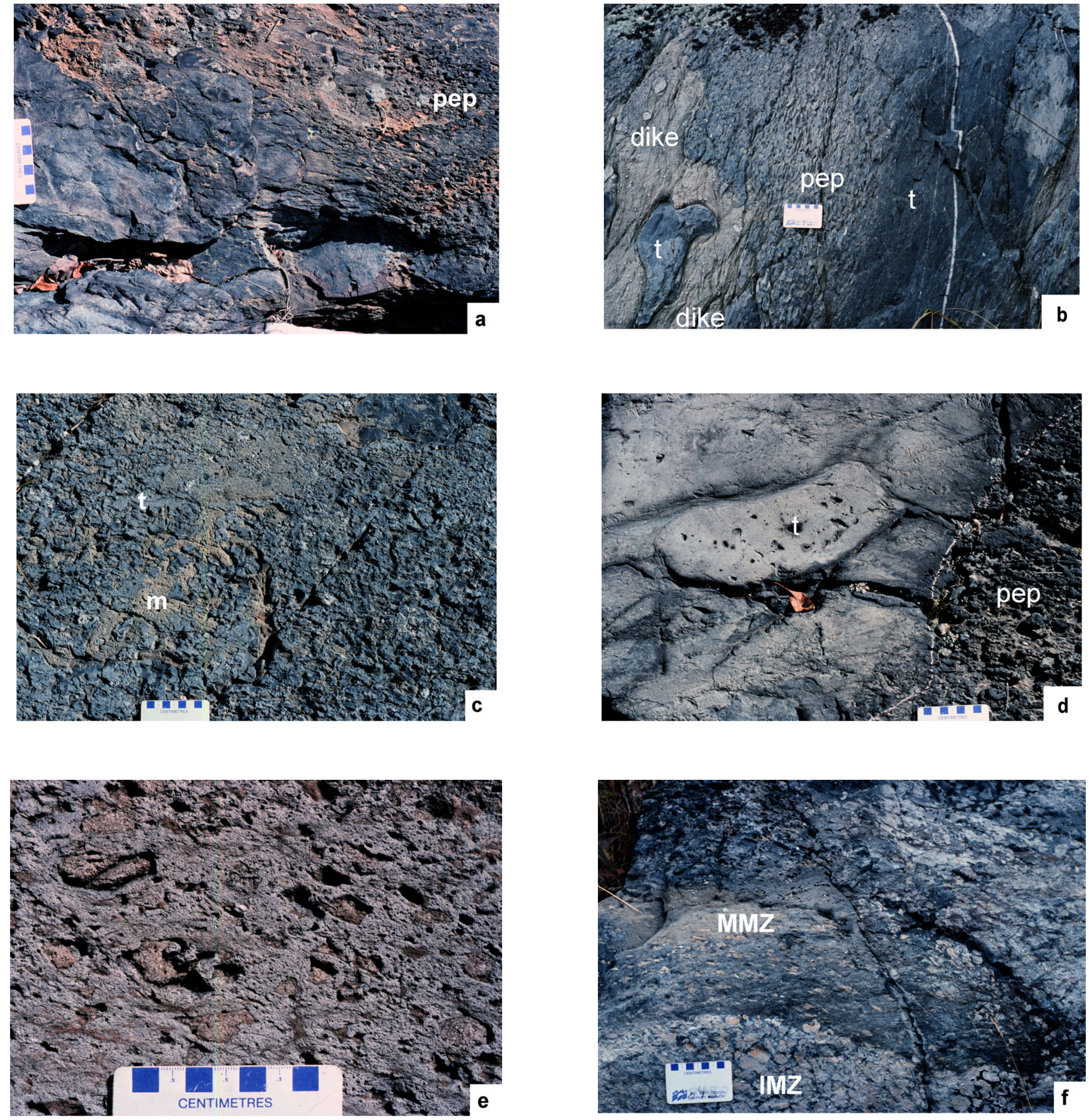


Figure S18. Geological map of the Gordon Lake Road peperite.



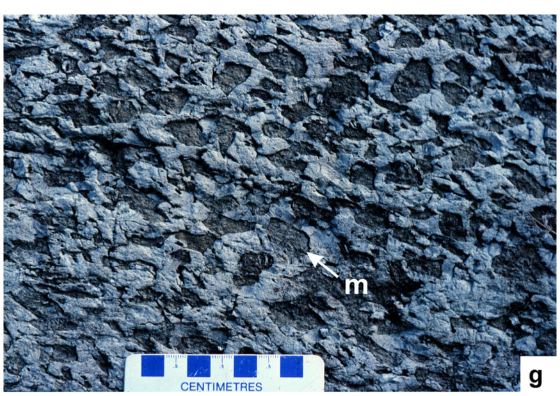


Photo S18. Gordon Lake Road peperite body. **a)** Transitional contact between massive tuff to auto-brecciated tuff to blocky peperite (pep). **b)** Transition from massive, homogeneous tuff (t) to blocky peperite (pep) to xenolithic dike (lighter coloured rock on left). **c)** Peperite lapillistone with green-weathering spherulitic andesite (m) containing angular fragments of silicified tuff (t). **d)** Contact between massive to autobrecciated tuff (t) with blocky peperite (pep). Fractures are chloritized. **e)** Close-up of peperite with delicate texture. **f)** Sharp contact between a zone of moderate mixing (MMZ) and a zone of intense mixing (IMZ) with blocky peperite. **g)** Classic blocky peperite texture. Dark-green, angular, sericitized, recessive-weathering andesite fragments (m = melt) are set in a matrix of grey, silicified tuff.

Stop 19. Kink bands in the Onaping Formation (E0474830, N5150923)

Drive approximately 3.7 km south of the Vermilion River bridge. Stop 19 is at the end of a straight stretch and where road enters a valley. The exposure is on right (west) side of the road and is partially hidden by alders.

Kink bands occur in foliated rocks and are narrow, more or less tabular zones, transverse to the foliation and bounded by subparallel surfaces called kink band boundaries or kink planes, at which the foliation is sharply deflected. Kink bands are small-scale folds with angular hinges and straight, asymmetric limbs. The angles α, β and γ refer to the angle between the foliation outside the kink band and the kink plane, the angle between the foliation inside the kink band and the kink plane, and the angle of rotation of the foliation inside the kink band, respectively.

Mean values for the Onaping kink bands (see Chapter 9, Figure 9.8 and Photo 9.3) are as follows: external foliation = 058° 60°SE; internal foliation = 058° 81°SE; kink planes = 067° 27°NW; hinge lines = 3°NE 060°; α = 84°; β = 73°; and γ = 23°. All the kink bands at the stop display a dextral sense of external rotation and are characterized by a large angle between the external foliation and the kink band boundary and by overrotation (i.e., α > β). The amount of shortening achieved by the kinking is trivial.

Stop 20. Cleavage, biogenic features (?) in Chelmsford Formation (E0477970, N5157864)

Drive north on Gordon Lake Road beyond the Vermilion River bridge. Turn right on Pilion Crescent and continue to Hwy 144 (do not turn left on Simmons Road). Turn left on Hwy 144 and go over railway overpass. Slow down and turn right on narrow road at the end of overpass, then turn left on paved but abandoned road. Drive to prominent outcrops near end of road.

On the north side of the road greywacke and argillite beds are exposed. Note the prominent rough cleavage (Photo S20.1). Careful examination indicates the cleavage is refracted (steepens) and is more closely spaced as it passes from greywacke beds into argillite beds.

Cross to the south side of road and climb up a dipping (30° E) bedding surface. Near the top are patterns with branching arms and others with fine lines (Photo S20.2) that may represent biogenic structures (microbial mats?).

Photo S20.1. Rough cleavage in the Chelmsford Formation.

Photo S20.2. Possible biogenic features in the Chelmsford Formation. Coin for scale.

Stop 21. "Type Section", Chelmsford Formation (E0476425, N5159226)

Return to Hwy 144, turn right and continue west. Cross the Vermilion River and turn left on first road over bridge (unnamed, paved). Drive about 50 m, park, and walk past abandoned building to the river.

This is the best exposure of the Chelmsford Formation. A total of 49 turbidite beds are exposed over a thickness of 54 m with only 8% covered. Sedimentary structures include concretions, exotic slabs, graded beds, cross beds, current marks, flame structures, load casts, ball structures, convolute laminations (Photo S21) and current ripples. See Rousell (1972) for a detailed measured section.

Photo S21. Convolute structure in Chelmsford Formation.

Stop 22. Onwatin Formation (E0475115, N5159258)

Return to Hwy 144, turn left and travel west approximately 1 km. Turn right on Houle Avenue then make an immediate right on Omer Street (parallel to Hwy 144) and go to the end of the street at a turn-around. The stop is on the right.

This exposure is typical of the Onwatin Formation, with steeply dipping slaty cleavage and shallow-dipping, delicate bedding. Fine-grained pyrite is disposed parallel to bedding with local, remobilized pyrite. Note the mafic dike.

Stop 23. A.Y. Jackson Lookout (E0470809, N5159489)

Return to Hwy 144 and proceed west through Dowling, then northwest to the A.Y. Jackson Lookout on the right.

The Lookout is at High Falls on the Onaping River. The location is a convenient lunch stop with a display of rock and mineral specimens. There are several large boulders of representative rock types and Ni-Cu mineralization. Two large blocks in the parking lot are labelled Onaping Formation. Terminology consistent with the most recent geological compilation map (Ames, Davidson et al. 2005) is as follows. The large boulder on the left is interpreted as representative of fall-back breccia of the Sandcherry Member (formerly Grey Member). The boulder beneath the sign is representative of the middle unit of the Dowling Member, an iridium-bearing breccia interpreted to have formed by the collapse of an impact plume (labelled Black Member).

Stop 24. Onaping Formation, Sandcherry Member (E0470676, N5159696)

Stop 24 begins on the west side of Hwy 144 just beyond the entrance to the Lookout and is about 300 m long. Leave the parking lot of the A.Y. Jackson Lookout, turn right on Hwy 144, drive north approximately 0.5 km and park in a large sandy area on the right. Walk back roughly 300 m.

At this locality, the Sandcherry Member is black due to the carbon content (Photo S24). Accretionary lapilli are common; cores of lithic fragments are rimmed by a fine aggregation of particles.

Photo S24. Black Sandcherry Member.

Stop 25. Onaping Formation, Sandcherry Member (E0470714, N5160491)

Return to vehicle and walk north on Hwy 144, past first exposure, for approximately 600 m to the second exposure located at the termination of a prominent ridge. (Do not drive to Stop 25 as parking is lacking.)

A semiconformable and discontinuous silicification zone (1300 x 280 m) occurs near the base of the Onaping Formation. There, equant shards of the Sandcherry Member have undergone silicification, with as much as 71 wt % SiO_2 compared to an average SiO_2 content of 61.5 wt % for unsilicified rocks. Other additions in the zone include Fe, Cu, Zn and S. The resulting assemblage of Fe-poor epidote-actinolite, pyrrhotite, chalcopyrite and sphalerite suggest temperatures of 340°C.

Photo S25. Silicified Sandcherry Member.

Stop 26. SIC: Granophyre–quartz gabbro contact (E0468432, N5162852)

Continue north on Hwy 144, turn left at RR 8 (road to Onaping/Levack). Begin at the prominent exposure at the intersection and on the left side of RR 8. Park in spacious area on the right side of RR 8.

At the southernmost outcrop, the base of the granophyre is a massive, medium-grained, buff to pink rock consisting of approximately three parts micrographic intergrowth (K-feldspar and quartz) to one part tabular, plagioclase phenocrysts. Accessory minerals include amphibole, biotite, chlorite and Fe-Ti oxides. Moving northward and down-section there is a gradual change in the micrographic intergrowth to plagioclase ratio. The contact is arbitrarily placed where the modal plagioclase exceeds that of intergrowth. At the crest of the hill, an oxide- and apatite-rich zone in the quartz gabbro contains several modal % magnetite grains, which makes the rock distinctly magnetic. The large outcrops down the slope belong to the quartz gabbro unit (see also Fedorowich and Nickolic 1999, Stop 5, p.43).

Photo S26. Granophyre–quartz gabbro contact. Note the subvertical fractures. Hammer for scale.

Stop 27. Sublayer (dark norite breccia) and Levack Gneiss Complex (E0467219, N5163320)

Drive north on RR 8, turn left on Onaping Drive, go past Lakeview, to the end of the Drive. Park near the hydro facility. Walk north along the road, past a building to the left, through a gate at old landfill site and continue on road. Large cliff to the right is Archean migmatite of the Levack Gneiss Complex; the cliff to the left is Sublayer.

The Sublayer varies in thickness from 0 to 200 m, with the greatest thickness in embayment structures. Good surface exposures of Sublayer are uncommon. The exposure here (Photo S27.1) is roughly 50 m thick and lies immediately above the Levack Gneiss Complex. Most of the fragments are norite, some are ultramafic with scarce felsic gneiss fragments. Most fragments are in the 0.01 to 0.5 m range and tend to be rounded to sub-rounded; a few are elongate with angular to ragged edges. Fragments locally display a weak preferred orientation sub-parallel to the basement contact. The matrix (30 to 45 modal %) is medium-grained and “noritic” in appearance, but acquires a more felsic appearance toward the footwall (north end of outcrop). Note that there is no embayment structure and little or no Footwall Breccia in the immediate vicinity (Fedorowich and Nickolic 1999, Stop 3, p.42).

The rocks of the Levack Gneiss Complex at this locality are medium-grained, pink to grey gneiss with a swirling foliation. Note the green epidote material (Photo S27.2)

Photo S27.1. Sublayer (dark norite breccia). Knife is 12 cm long.

Photo S27.2. Levack Gneiss Complex. Notebook is 10 cm wide.

Stop 28. North Range felsic norite (E0467205, N5162266)

Return to Hwy 144, turn right and drive north. Turn left at Elk Club Road and park. The stop is at the intersection of Elk Club Road and Hwy 144.

Felsic norite is a grey, massive, medium-grained, hypidiomorphic rock and is the major unit of the Main Mass in the North Range. Felsic norite differs from Main Mass rocks of the South Range as the latter are black or dark green in colour. Note the presence of several generations of alteration veinlets.

Photo S28. Felsic norite. Knife is 27 cm long.

NOTE: PERMISSION IS REQUIRED TO VISIT THE NEXT THREE STOPS.

Stop 29. Sudbury Breccia (SB) and mineralization at Barnet Hill (E0474760, N5170550)

Return south on Hwy 144 and turn left on RR 8 and drive 1.7 km. Just before the Onaping River, turn right on the access road to the Xstrata mine sites. Check in at gate and proceed along the private road to the Strathcona Mill complex.

Drive through the parking lot at the Xstrata Strathcona Mill complex and continue along the road at 035° and through the storage yard (1.0 km). The gravel road branches to the right; follow for 0.8 km as it curves from 030° to 000°. At this point it branches to the right again and is followed for another 0.75 km along a curve from 032° to 355°. An intersection with an east-west trail is encountered. Continue north-northwest for approximately 25 m across this intersection, at which point the road will continue to the north for 0.1 km and will wind gently northeast and north for another 0.3 km. At this point, a steel post marks a narrow trail branching to the right. Park here and take this trail across a light steel bridge crossing a creek and continue for 0.75 km along this trail heading north (a small lake on the left at about 0.25 km) and eventually north-northeast. Flagging tape at the left side of the trail marks a bush path heading at 340° for 0.1 km to the base of the Barnet stripped exposure, which trends north to northeast for approximately 220 m.

At this locality a dike of SB (approximate attitude 050° 80°SE) is enclosed by footwall gneiss and is exposed over a width of about 25 m. In addition, a few parallel veins of SB occur in the hanging wall gneiss (Figure S29). Due to the aphanitic nature of the matrix it is difficult to determine if the original material was a rock flour or a glass. Scattered sulphide mineralization (chalcopyrite, pyrrhotite, pentlandite) in the form of stringers, veins and disseminations occurs mainly in SB but also in the gneiss. Note that the location is an up-plunge surface expression of the Strathcona Deep Copper vein system.

Three types of SB contacts are present. These are: 1) sharp contacts (Photo S29); 2) ductile deflection of the gneissic foliation parallel to SB dikes and emplacement of SB into conjugate shear zones possibly due to north-south compression (see inset of Figure S29); and 3) feldspathized contacts.

Apart from SB, there are three other vein types, all of which cut SB: 1) feldspar-quartz veins (approximate attitude 110° 87°N); 2) sulphide veins (attitudes of 033° 87°N and 123° 87°S) locally synkinematic with feldspar-quartz veins; and 3) epidote-quartz-chlorite veins that have variable strikes and steep dips and cut feldspar-quartz veins (Fedorowich and Nickolic 1999, Stop 1, p.33.)

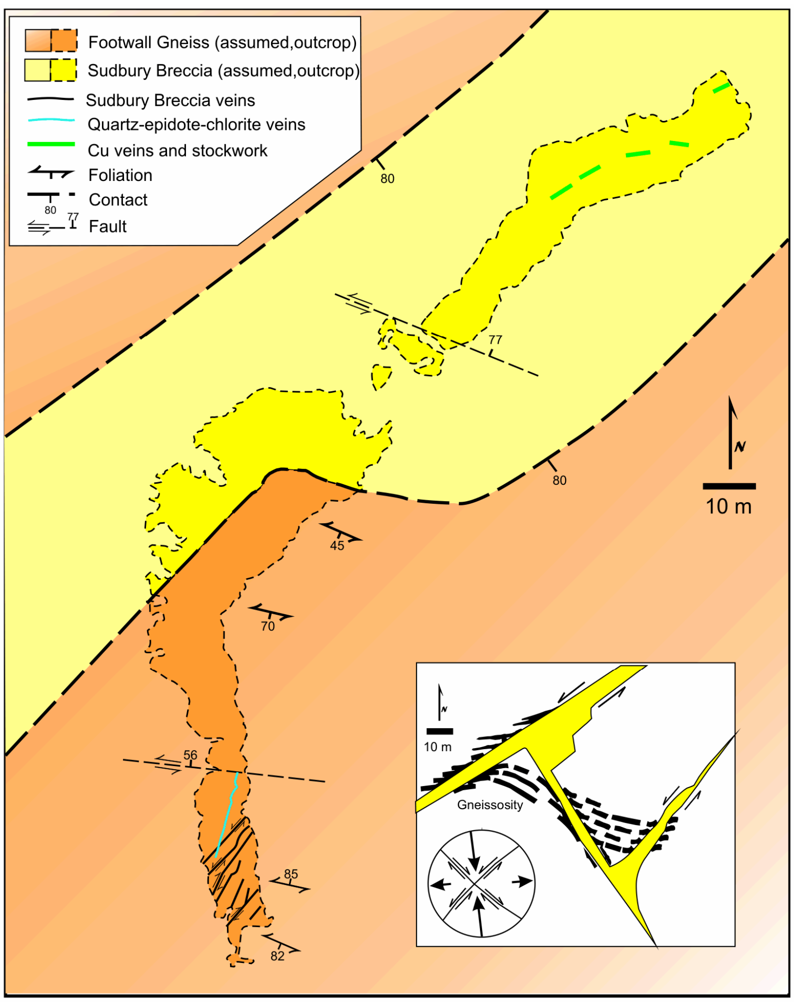


Figure S29. Map of Barnet Hill location (*after* Fedorowich and Nickolic 1999). Inset shows extensional jog in Sudbury Breccia.

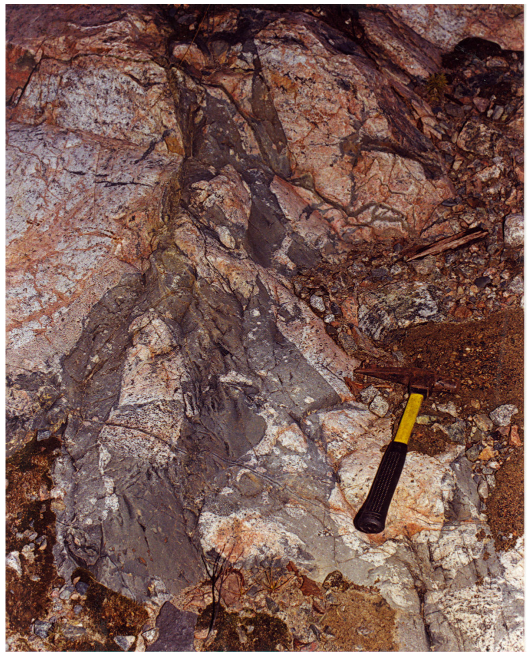


Photo S29. Sudbury Breccia dike (dark colour) cross-cuts granitic footwall gneiss. The sharp contact at the left side of the dike has a 4 cm wide feldspathized margin within the gneiss.

Stop 30. Mineralized Footwall Breccia; Longvack pit to Longvack South Mine

Return to the road where the steel post marks the Barnet trail. Go west-northwest along the road for 150 m, then head due west, off the road, for 80 m. An outcrop ledge rises approximately 4 m above the base elevation and is the remnant of the north wall of the Longvack pit (E0474597, N5170090).

Figure S30 shows the location of four stations. Station 1 is at the north wall of the Longvack pit, station 3 is at the south wall and station 2 is between them. In 1969, 1.21 Mt of ore was extracted (0.94% Ni, 0.6% Cu). The pit was filled and seeded during a reclamation project in the 1990s. Station 4 is located at the site of the former headframe of the Levack South Mine where 2.18 Mt were mined at 0.92% Ni, 0.51% Cu.

Station 1 is at the gradational contact between Footwall Breccia (FB), also known as Late Granite Breccia, and footwall gneisses. There, FB veins, 1 to 3 cm wide and with an attitude of 057° 72°S, penetrate footwall rocks. A vein-parallel foliation in the latter rocks, enhanced by strung-out mafic blebs, define a local strain. Clasts, comprising a variety of rock types derived from the footwall, are hosted by the FB (Photo S30.1).

Station 2. Stockworks of multi-oriented, pinkish-white felsic veins cross-cut a footwall diabase dike (*in situ* brecciation, Photo S30.2) and grade into patchy brecciation where diabase fragments have undergone some rounding and transportation.

Station 3. At this location there is some sulphide staining on the walls. The mineralization is patchy and grades from a halo of fine-grained sulphides, mainly pyrrhotite, to blebby, locally semi-massive accumulations. The sulphides are apparently confined to the FB matrix.

Station 4. The gradational contact (1 to 5 m) between FB and Sublayer (dark norite breccia) is exposed at this locality. Note that the Sublayer contains scattered felsic fragments (Fedorowich and Nickolic 1999, Stop 2, p.39.)

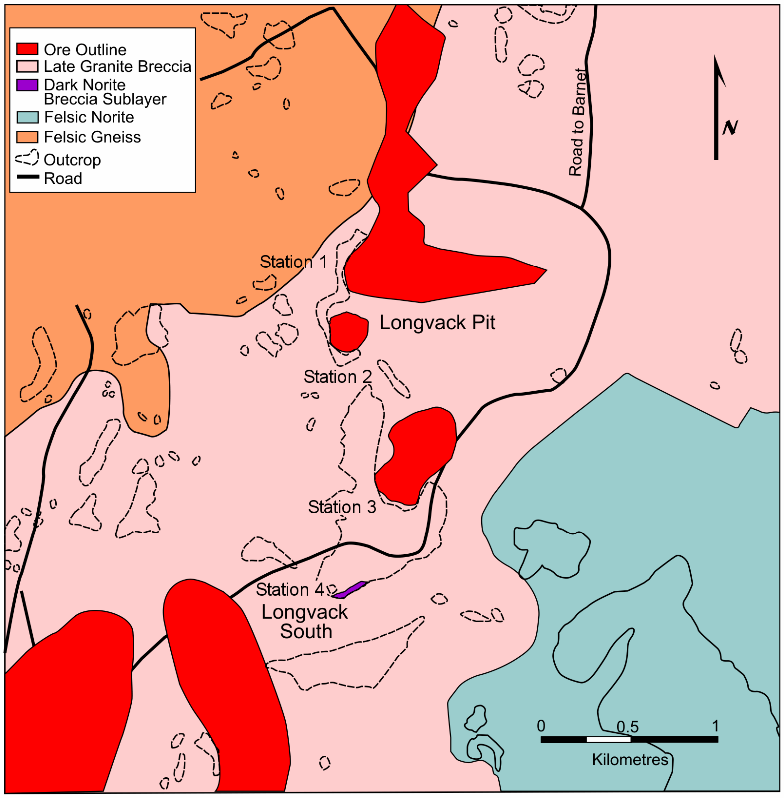


Figure S30. Longvack pit–Longvack South Mine area with station locations (*after* Fedorowich and Nickolic 1999).

Photo S30.1. Footwall breccia at Longvack pit. Coin for scale.

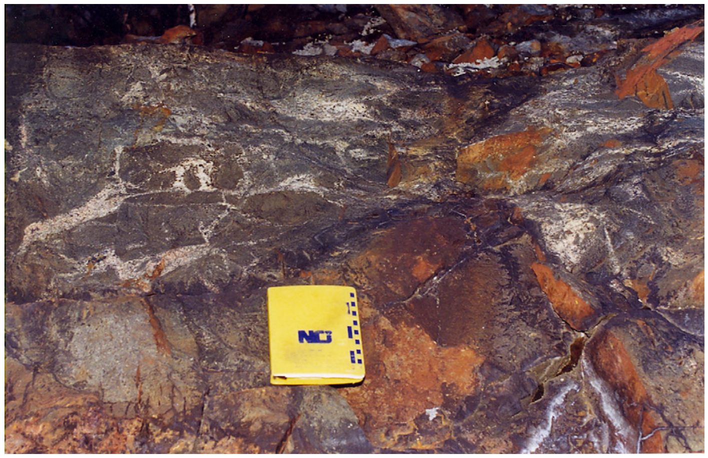


Photo S30.2. Footwall Breccia stockwork veins crosscut a footwall diabase dike.

Stop 31. Footwall Breccia at Strathcona Mine (E0473686, N5169159)

Return to the Strathcona Mill complex and park in the parking lot. Stop 31 is a southeast-facing slope located across the road. As you cross the road a loading-station silo is to the right and a small ore pad is to the left. The base of the slope is approximately 100 m from the parking lot with an up-slope distance of about 80 m.

The exposure of Footwall Breccia (FB) is on the slope facing the Strathcona Mine headframe. The outcrop is almost entirely FB (see Photo 7.1), with some Sublayer (dark norite breccia) at the base of the hill near the roadside. Clasts are matrix-supported, subrounded, range in size from 0.01 to 1 m, and consist mainly of footwall gneiss together with some massive mafic fragments. Locally, there is a weak preferred orientation. In general, the matrix is very fine-grained, has a metamorphic texture, varies from pink to light grey and contains a variety of fragments that are < 2 cm in diameter, with quartz and feldspar the main minerals. Minor sulphide minerals, dominantly pyrrhotite, occur as disseminations, blebs and veinlets (Fedorowich and Nickolic 1999, Stop 3, p.42).

Stop 32. Levack Gneiss Complex, Matachewan dike and Sudbury Breccia (E0464639, N5164228)

Return to Hwy 144 and drive north to Old Cartier Road / Windy Lake Motel. Continue north approximately 1 km and park in a small area on the right. Stop 32 is 30 m to the south. A steel marker is on the outcrop.

At this locality, the gneiss contains abundant amphibolite, pyroxenite and minor paragneiss xenoliths (Photo S32). Large clinopyroxene porphyroblasts are present in both the leucosome and the mafic enclaves. Zircon from local tonalite gneiss yielded a U-Pb age of 2711 Ma, interpreted as the minimum age of the precursor tonalite intrusions (Krogh et al. 1984). Zircon from the leucosome of local tonalite diatexite yielded a U-Pb age of 2661 Ma, interpreted as the age of an early phase of migmatization (Card and Wodicka, in prep.).

The diatexite is cut by a northwest-trending mafic dike that is considered to be part of the Matachewan swarm (2450 Ma). Sudbury Breccia is a minor constituent and consists of thin, pseudotachylite veins with centimetre-scale, locally derived, rock fragments. The veins cut both the gneiss and the dike.

Photo S32. Steeply dipping rocks of the Levack Gneiss Complex. Pen for scale.

Stop 33. Diatexite of the Levack Gneiss Complex (E0460732, N5167429)

Continue north on Hwy 144. Go past a locality where a power line crosses the road (4.5 km north of Old Cartier Road) to a second crossing of the power line at 6 km north of Old Cartier Road. Park on the shoulder or in a sloping area.

Layered, medium- to coarse-grained, inhomogeneous diatexite with inclusions of paragneiss and amphibolite are part of a kilometre-scale enclave of the Levack Gneiss Complex enclosed in, and intruded by, the Cartier granite. Paragneiss of the complex is interpreted as metamorphosed greywacke (Card 1994). The age of the rocks is not known but their occurrence as enclaves within tonalite gneiss indicates they are older than 2700 Ma, the minimum age of the precursor tonalite intrusions (Krogh et al. 1984). Pegmatite dikes that cut the diatexite are probably part of the Cartier suite.

Photo S33. Pale, horizontal layers of diatexite (migmatite) of the Levack Gneiss Complex. Pen for scale.

Stop 34. Cartier granite and Sudbury Breccia (E0458470, N5169129)

Drive 9.5 km north of Old Cartier Road. Exit right at the sign for Crab Lake and park. Note the large graffiti-covered glacial erratic. Walk south, on east side of Hwy 144, for approximately 200 m.

Massive, porphyritic Cartier granite is cut by Matachewan dikes that contain scattered pale-green feldspar grains approximately 2 cm in diameter. Both the granite and the dikes are cut by Sudbury Breccia in the form of anastomosing veins, dikes and irregular masses. The U-Pb age of the granite at this location is 2642 Ma (Meldrum et al. 1997).

Photo S34. Sudbury Breccia cuts granite and mafic dike (left). Notebook is 10 cm wide.

Stop 35. Gowganda Formation with peperite (E0454360, N5178660)

Continue north on Hwy 144 for 21 km north of Old Cartier Road or 12 km north of Cartier. Park, with caution, on the right side of the highway, opposite Percival Lake.

The diffuse, bulbous nature of the contact between sandy conglomerates of the Gowganda Formation and Nipissing diabase, combined with the lack of a clear chilled margin, suggests that the diabase intruded wet, unconsolidated gravel to form peperite (Photo S35.1). Contact framework conglomerates pass up-section into massive, thick-bedded units of clast-poor mixtite that formed as sub-glacial melt-out tills. The highly lenticular unit of clast-supported boulder conglomerate in the middle of the outcrop may have formed within a sub-glacial conduit to produce an esker (Photo S35.2).

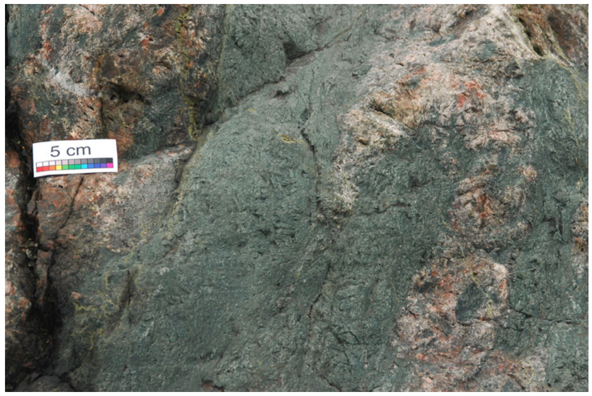


Photo S35.1. Contact between Nipissing diabase (peperite) and Gowganda Formation.

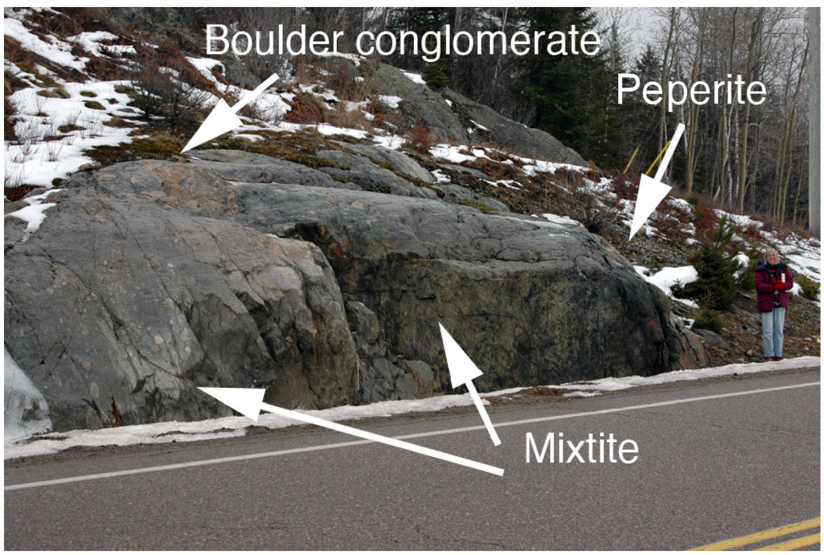


Photo S35.2. Outcrop of Gowganda Formation and Nipissing diabase.

Stop 36. Cartier granite (rapakivi texture), Gowganda Formation (E0453617, N5180228)

Continue north about 1 km and turn into bush road on the left (west) side of the highway, immediately south of the railway tracks.

This small outcrop exposes a thin sequence of stratified, clast-rich diamictites, overlain and interbedded with massive and horizontally laminated medium-, coarse-, and minor fine-grained sandstone (lithic arkose). Note the wide variety of subangular to rounded clast types. Most of the conglomerate is clast-supported and was deposited by traction currents in a sub-glacial setting, beneath flotation ice (Photo S36.1). Locally, rare unconsolidated mudstone interclasts were transported in sub-glacial flows and picked up an armour of small pebbles.

Walk along the east side of Hwy 144 for about 200 m to a prominent rockcut. This cut is an excellent exposure of several phases of the Cartier batholith (Photo S36.2). Of particular interest is a feldspar porphyry, which consists of 50% euhedral crystals of pink potassium feldspar up to 5 cm in length, locally mantled by green plagioclase (rapakivi texture).

Photo S36.1. Typical poorly sorted sandy, cobbly, large pebble conglomerate at the north end of the outcrop. Note crude stratification and local grading.

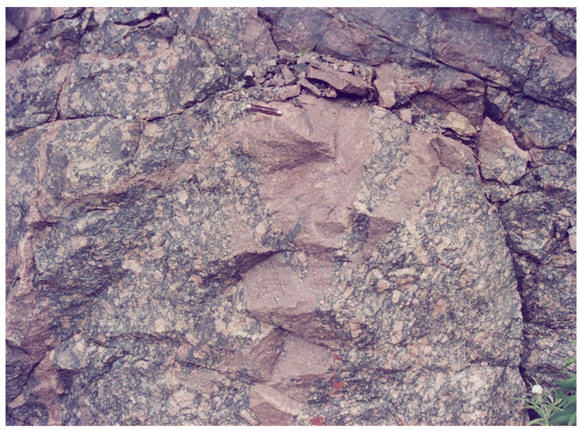


Photo S36.2. Fine-grained granite cuts feldspar porphyry. Note the preferred orientation of the feldspar crystals.

Stop 37. Cartier batholith and Matachewan dikes, Benny greenstone belt (E0453660, N5182160)

Continue approximately 2 km north of railway tracks to a rockcut on the west side of Hwy 144 at the junction with the road to Benny (no sign).

This stop illustrates the relationship between low-grade Archean metavolcanic rocks of the Benny greenstone belt, the Neoarchean Cartier batholith and the Paleoproterozoic Matachewan dikes. Coarse andesitic tuff-breccia of the Benny belt contains lithic fragments that are flattened and elongated within the dominant foliation plane. The tuff-breccia is cut by an undeformed, 10 m thick, northwest-trending porphyritic diabase dike with chilled margins. Locally the diabase is irregular (Photo S37). The trend and texture indicate that the dike belongs to the Matachewan swarm. An intrusive contact between the foliated tuff-breccia and a massive, monzonitic phase of the Cartier batholith is exposed at the southern end of the outcrop. Granite dikes, presumably offshoots of the batholith, are also present.

The attitude of the foliation in the tuff-breccia changes from 122° 56°S at the northern part of the outcrop to 045° 53°S near the contact with the Cartier batholith. Close to the contact, a C/S fabric, showing dextral displacement, is developed in the tuff-breccia. The outcrop lies just north of the Benny deformation zone, which shows sinistral displacement and which may have been active as early as Neoarchean (Kellet and Rivard 1996). The fabric elements present in the metavolcanic rocks were probably developed prior to, or during, the early stages of emplacement of the Cartier batholith (Ames et al. 1997).

Photo S37. Irregular mafic dike cuts tuff. Knife is 12 cm long.

Stop 38. Onaping Formation (E0492969, N5158191)

Return south on Hwy 144 and go through Chelmsford. Continue southeast on RR 35 past Azilda (on the right). Turn left on Gagnon Street, then north on Montee Rouleau. The stop is 4 km north of the railroad tracks and past a landfill site to the top of a hill.

Here the Onaping Formation displays a strong foliation (065° 55° SE) and a down-dip stretching lineation (Photo S38). This L-S fabric developed in the South Range shear zone, a southeast-dipping zone of reverse shear.

Photo S38. Onaping Formation with foliation and down-dip stretching lineation.

Stop 39. Discovery Site (E0495896, N5151733)

Return to RR 35, turn left, and continue south to the Discovery Site on the left and opposite the Clarabelle open pit, now filled with water.

Nickel-copper sulphide mineralization was first reported from the vicinity of Creighton Mine in 1856 by A. Murray of the Geological Survey of Canada. This report was ignored until 1883 when, during construction of the Canadian Pacific Railway, a rock-cut exposed a mineralized zone, which was subsequently developed as the Murray Mine. The original discovery site was located close to the rim of the water-filled open pit, which can be seen west of the highway. In the mid 1970s the road, rail line and discovery site were relocated to permit mining of the upper part of the Murray orebody.

At the present site, a facies of Contact Sublayer, known as gabbro-peridotite inclusion sulphide, is exposed. It consists of a variety of rounded inclusions of mafic and ultramafic rocks enclosed in a matrix of massive Cu-Ni-Fe sulphide minerals with, or without, a minor amount of igneous silicate material. The Sublayer is overlain by quartz-rich norite of the Main Mass (SIC), which contains patchy, disseminated sulphide minerals whose content decreases rapidly upward. To the northwest, along RR 35, typical facies of the South Range Main Mass are exposed in fresh road cuts. A large olivine diabase dike of the Sudbury swarm (1240 Ma), which cuts the SIC, is exposed a few hundred metres north of the open pit.

Stop 40. McKim Formation (E0499116, N5147437)

Return to Sudbury via RR 35, which becomes Elm Street. Turn right (south) on Big Nickel Road to Lorne Street. Turn left on Lorne Street and go to Edna Street (near Tim Horton's) and park. Walk east, past houses. The stop is on the north side of the street.

The formation consists of thin beds of medium- to fine-grained sandstone and thin, laminated mudstone. The sandstones are typically graded (turbidites) and locally exhibit ripple cross-lamination. Scour surfaces are locally present (Photo S40.1). Note cleavage refraction across units of different grain size. Thin veins of pseudotachylite Sudbury Breccia are present on the glaciated top of the exposure.

Further west along Lorne Street (next to Taylor-Made Auto Body Shop), is a glaciated, oblique surface through the formation, which exposes convolute bedding and sinusoidal ripples (Photo S40.2).

Photo S40.1. Localized scour (at arrow) beneath turbiditic sandstone unit in the McKim Formation.

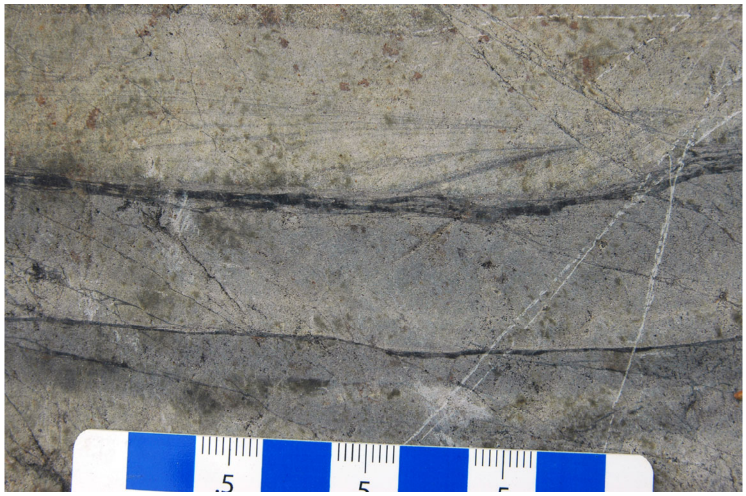


Photo S40.2. Sinusoidal (climbing) ripple cross-lamination in base of sandy turbidite unit in the McKim Formation.

Stop 41. Contact between Creighton pluton and Elsie Mountain Formation (E0494743, N5147449)

Drive west on Lorne Street (RR 55), turn right on Balsam Street and enter the town of Copper Cliff. Turn right on Godfrey Drive and drive north to the intersection with Cliff Street and park.

The contact is obscured by the road. Both the pluton and the formation are deformed. Two granitoid phases of the pluton are present north of the road intersection. Note the shape-preferred orientation of quartz and biotite.

Metabasalt (amphibolite) and local sedimentary layers of the Elsie Mountain Formation outcrop south of the main road. The preferred orientation of amphibole needles, formed under amphibolite facies metamorphism, defines a strong foliation and horizontal lineation. Note the stretched amygdules. The intense deformation may have been induced by ductile flow past the southeastern corner of the quasi-rigid granitoids of the Creighton pluton during regional deformation ("corner effect") or by expansion (ballooning) of the pluton during emplacement.

Stop 42. Aplite dikes in Elsie Mountain Formation and Sudbury Breccia (E0494900, N5147823)

Follow Godfrey Drive uphill, toward Clarabelle Mine, for about 300 m. Park before crossing the overhead line. Godfrey Drive is now gated. Walk to the right and up the hill and beyond the overhead line.

The outcrop displays veins and pods of aplite cutting both metabasalt of the Elsie Mountain Formation and pseudotachylite Sudbury Breccia. The aplite resembles the leucocratic phase of the Creighton pluton. The aplites were emplaced as follows:

1) both concordant and discordant to the foliation in metabasalt;
2) at the contact between metabasalt and Sudbury Breccia (Figure S42); and
3) within the breccia.

Aplite dikes have straight contacts with metabasalt but irregular and lobate contacts where aplite is enclosed by Sudbury Breccia. The lobate contacts suggest that, during emplacement, the viscosity of aplite and breccia were similar. The heat source for aplite mobilization may have been the Copper Cliff offset dike (exposed immediately to the east of the road), the Main Mass of the SIC, or the impact event.

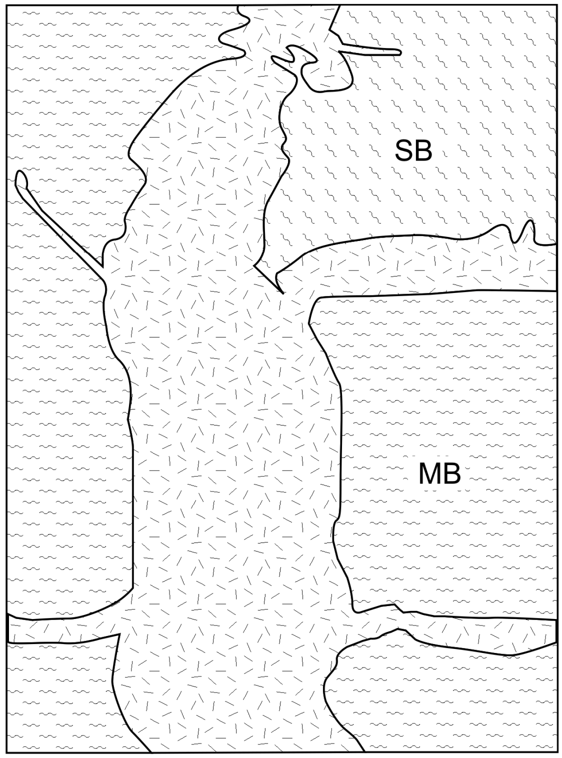


Figure S42. Aplite cuts foliated metabasalt of the Elsie Mountain Formation (MB) and pseudotachylite Sudbury Breccia (SB). Width of figure = 54 cm.

Stop 43. Staurolite and folds in Stobie Formation (E0498277, N5151054)

Return to Lorne Street (RR 55) and turn left. Continue east on RR 55 then turn left on RR 34 (Big Nickel Mine Drive) to RR 35 (Elm Street). Turn left on RR 35 and drive northwest; turn right on Lasalle Boulevard (RR 71) and drive for about 1 km (pass private road on the right). Park on the northwest side of RR 71 in front of a gate and a short fence. Walk to the right and around the fence, over an embankment and across a rusty wet area. Climb the rock face. Continue north-northwest, through the trees, for approximately 300 m to a prominent exposure with an extensive view. Frood Mine is to the north-northeast.

The formation consists of bedded argillite, siltstone and wacke. Note the prominent pseudomorphs after staurolite (Photo S43.1). Plunging folds (Photo S43.2) and patches of Sudbury Breccia are also present. The interested geologist is referred to the detailed map (1 inch to ¼ mile) by Innes (1978b).

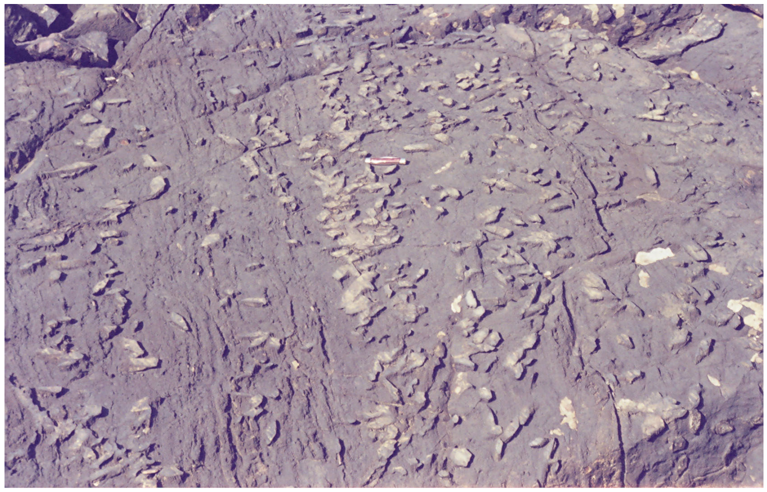


Photo S43.1. Pseudomorphs after staurolite. Stobie Formation. Knife is 6 cm long.

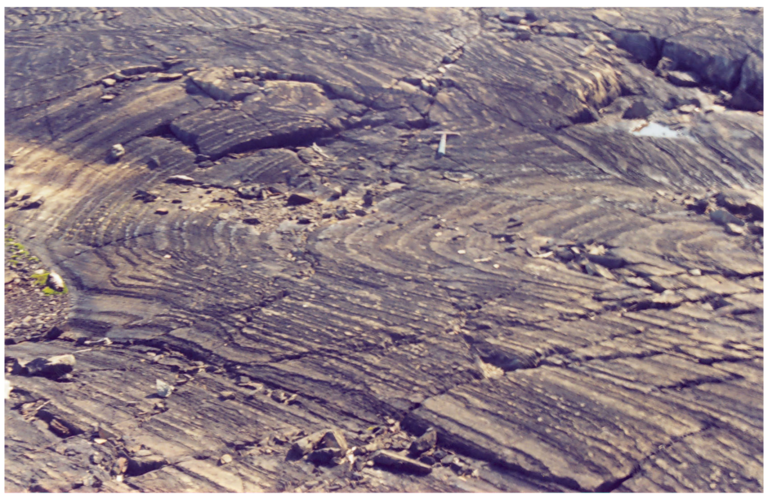


Photo S43.2. Plunging folds. Stobie Formation.

Stop 44. Elsie Mountain Formation–SIC (Sublayer, norite) contact

Drive east on Lasalle Boulevard, past Frood Road, to Notre Dame Avenue (RR 80). Turn left (north) on RR 80. Drive 3.8 km north of the intersection. Park on the right (east) side of RR 80 at a gated road that leads to Whitson Lake (E0501272, N5155928). Note: Do not stop at the OFSC gravel road, located 3.2 km north of the intersection and which continues on the west side of RR 80.

Walk south, on the east side of RR 80, across the contact. The mineralized Sublayer contains scattered mafic clasts, some of which have an enigmatic orbicular texture. Abundant rusty spots of weathered pyrrhotite are rimmed by biotite, as seen in thin section. Continue south for about 0.5 km to the prominent exposure of basalt pyroxene hornfels of the Elsie Mountain Formation.

Return to the Whitson Lake road. South Range norite is well exposed north of the road. Dark grey cumulus norite consists of unaltered plagioclase, orthopyroxene and clinopyroxene with minor biotite, primary hornblende and granophyre. The latter is a cotectic intergrowth of alkali feldspar and quartz. Where the norite appears green, the plagioclase may be saussuritized and the pyroxenes altered to amphiboles. The Sublayer norite differs from the main norite in having inclusions and, in general, more disseminated sulphides.

Stop 45. South Range Norite (E0501270, N5155943)

Proceed north on RR 80 about 300 m to the first prominent roadcut south of the entrance to the Thayer Lindsley Mine.

South Range Norite consists of plagioclase-hypersthene-augite cumulates together with minor quartz, granophyre and primary biotite (Photo S45). It is the main norite member of the South Range SIC. Disseminated sulphides decrease upward in the Main Mass Norite so that they are scarce at the stratigraphic level of this stop. However, at Stop 46, which is higher in the section, there are more sulphides. A possible explanation is as follows. Fractional crystallization of the SIC led to an increasing Fe^{++} concentration in the residual magma. FeS solubility in magma is proportional to Fe concentration and this increase caused the sulphide to become undersaturated. Sulphide became saturated in the overlying transition zone as precipitation of magnetite and ilmenite reduced the Fe^{++} concentration in the liquid.

Photo S45. South Range Norite. Knife is 12 cm long.

Stop 46. South Range Quartz Gabbro (Transition Zone) (E0500674, N5157734)

Continue north on RR 80 approximately 2 km to the top of a hill (McCrea Heights), turn right on Robin Street and park. Walk back to the prominent rock cut and the exposure on top of the cut.

The Quartz Gabbro is a coarse-grained rock consisting of plagioclase, augite, hornblende, granophyre, magnetite and ilmenite. It comprises a narrow transition zone between the thick norite below and the granophyre above. Several aplite dikes, up to 6 m wide, cut the Quartz Gabbro. Heat from the adjacent SIC may have caused melting of the Murray granite. The Quartz Gabbro adjacent to aplite dikes are locally cut by narrow (< 1 m) ductile shears (Photo S46) with the south-side up. Also present are Riedel shears, which dismember narrow aplites, and C/S structures. These elements may be a small-scale representation of the larger-scale elements of the South Range Shear Zone, which begins a few hundred metres to the north.

Photo S46. Narrow shears, with several strike directions, cut Quartz Gabbro. Pen for scale.

Stop 47. Onaping Formation at Joe Lake (North Range) (E0500000, N5174858)

Go north on RR 80 (Desmarais Road) through Val Caron and Val Therese. Do not turn right (east) on RR 80 but continue north on Desmarais Road, which becomes RR 96. Go past Gravel Road and Theriault Road for about 5 km and turn right at Joe Lake Road / Frenchman Lake (green sign), then turn left at the Fatima church and at the sign that says “Joe Lake 2 km”. Continue past Frenchman Lake and go north just past the intersection of the Joe Lake West road. The outcrop is on the west side of the Joe Lake East road at the top of the hill going down to Joe Lake.

The exposure offers a spectacular example of intense fracture-controlled hydrothermal alteration of equant-shard units (interpreted as fall-back breccia) of the Sandcherry Member. The pinkish-white albitization shows ghost textures of equant shards, blocky shards and relict lithic fragments (Photos S47a, b).

Proceed on foot around the corner heading southwest to Joe Lake West road to an outcrop on the south side of the road. This is an exposure of a xenolith-rich quartz diorite intrusion (Onaping) with K-feldspar-epidote-quartz miarolitic cavities (Photo S47c). Similar cavities also occur in granophyre (SIC) and in granophyre patches in the footwall and are indicative of a high volatile content. Note that volatiles are important in mobilizing metals at Sudbury.

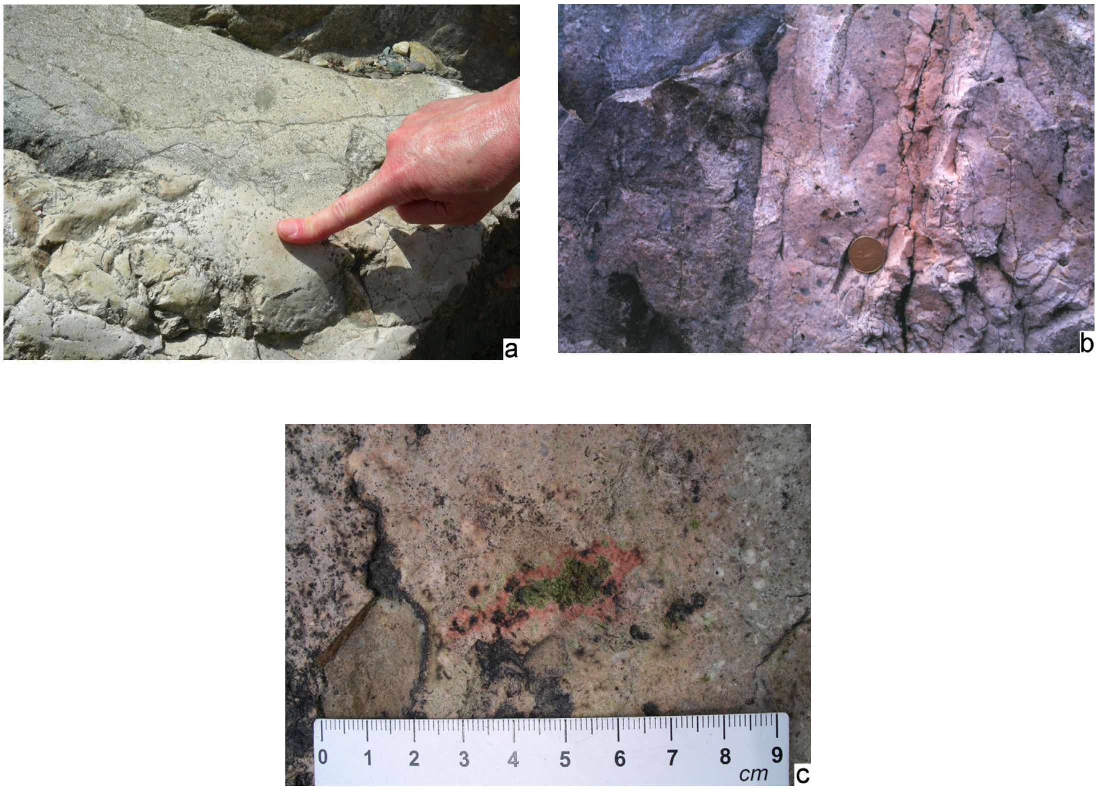


Photo S47. Alteration of the Sandcherry Member, Onaping Formation, near Joe Lake. **a)** Sharp, fracture-controlled albitization alteration front in the equant-shard units. **b)** Albitized equant-shard units; note the orangey, ghost-like remnant shards and quartzose lithic fragments in the greyish matrix. **c)** Miarolitic cavity rimmed with K-feldspar and filled with epidote ± quartz in fine-grained quartz diorite intrusion.

Stop 48. Foliated granophyre (SIC) of the South Range Shear Zone (E0513850, N5161790)

Return south on RR 96. Turn left (i.e., east) on RR 80 through Hanmer to RR 85 (RR 80 takes a jog to the south at Notre Dame Avenue). Go south then east (Radar Road) on RR 85 to the intersection with RR 86 (Bailey Corners). Prominent exposures are located in the vicinity of the intersection.

The stop is located in the bulge at the junction of the East and South ranges. The axis of the bulge trends northwest. The prominent foliation strikes parallel to the axis and dips steeply (80° SW); the stretching lineation pitches steeply southeast. Cowan and Schwerdtner (1994, Photo 5.2) describe an outcrop where the foliation bisects a conjugate set of narrow, steeply dipping shears, which form a lozenge-shaped pattern. The structural elements suggest a subvertical extension and a bulk northeast-southwest shortening. Unfortunately, road work has destroyed this exposure.

Return to Sudbury.

Stop 49. Dynamic Earth   **by Brenda Koziol**   Lorne Street and RR 34

Since opening in 2003, Dynamic Earth has become one of Canada's premier science and educational centres. Dynamic Earth is a geology and mining attraction where interactive exhibits, multi-media shows and an underground mine tour highlight the unique geology and rich mining heritage of the Sudbury area. In a glass-enclosed elevator, visitors descend 20 m into a spectacular rock chasm to embark on an underground mine tour. They walk through drifts to witness the technology and working conditions of different eras of mining in the Sudbury area, from the late 1800s, through the 1950s, to modern day mining. Back on surface, they can explore galleries full of interactive exhibits about rocks and minerals, Sudbury geology, and mining technology. Multi-media theatres show how Sudbury ore is transformed into pure nickel and copper, and how Sudbury has grown and developed over the past 125 years. The building also houses a 125-seat theatre with digital projection for films and live presentations on earth science topics, and a special exhibitions hall for changing exhibits.

The Dynamic Earth building itself was designed to reflect Sudbury's mining heritage. Its shape replicates a mining headframe, its exterior is clad with black bricks made from crushed slag, and artwork made of pure copper adorns its side. However, the Big Nickel is the site's most famous landmark. It is a replica of the 1951 Canadian five cent coin, minted to commemorate the 200th anniversary of the identification of nickel as a new element by A.F. Cronstedt, and to recognize Canada's important role in nickel production. The Big Nickel, 9 metres across and 60 centimetres thick, is sheathed with stainless steel. It has sat atop this hill since 1964, except for a couple of years when the old Big Nickel Mine attraction was closed to make way for construction of Dynamic Earth. At that time, the Big Nickel was entirely refurbished and a new base designed to make it more accessible to visitors.

The site offers great panoramic views of Sudbury, including Vale Inco's Copper Cliff Smelter across the roadway from Dynamic Earth. The smelter's superstack is 381 metres high, making it one of the tallest chimneys in the world. Huge piles of black slag, a waste product from the smelting process, can be seen stretching to the east away from the smelter. Many of the slag slopes are now green with vegetation, the result of a recent regreening project by the company. Another prominent feature on the skyline to the south is a structure that resembles the Eiffel Tower, part of a radio communication system for a unique ready-mix concrete production plant located within the hill below. The headframe for South Mine can also be spotted on the horizon about halfway between the superstack and the tower structure.

The site also bears evidence of the industrial devastation of Sudbury's landscape and the remarkable story of its restoration and recovery. Much of the exposed bedrock has been blackened due to past smelting practices. From the late 1880s to 1929, Sudbury ore was burned over open fires in outdoor roast beds, releasing clouds of sulphur dioxide that rolled across the landscape, killing vegetation and blackening the rock. Today, much of the Dynamic Earth site and the surrounding hills have been revegetated as part of Sudbury's Regreening Program. Since the program began in the 1970s, thousands of hectares of damaged land have been treated with lime and seeded with grass, and more than 12 million trees planted. The large spruce tree about 30 metres down the hill to the south of the Big Nickel was the three millionth tree planted in 1997 to celebrate Sudbury's Regreening Program and the hundreds of volunteers and employees involved in improving the environment.

The bedrock at Dynamic Earth is well exposed, making it a good place to study geological features. The bedrock underlying the site consists of laminated siltstone and argillite of the McKim Formation, part of the Elliot Lake Group of the Huronian Supergroup. Close examination reveals graded bedding, cross lamination and climbing ripples formed during deposition by distal turbidity currents about 2.45 billion years ago. In some of the finer-grained portions of the beds, prominent porphyroblasts up to 1 cm long formed during a metamorphic event of the Penokean Orogeny between 1.9 and 1.7 billion years ago. These were originally staurolite grains, but have since been retrograded to sericite. In places, quartz veins several centimetres wide stand up above the surface of the outcrop due to differential weathering.

Sudbury Breccia can be seen in outcrops near the Big Nickel and other places on site. The breccias are an unusual geological feature, generally considered to have formed during a meteorite impact that created the Sudbury Structure about 1.85 billion years ago. Irregular patches of breccia contain blocks and sub-angular to rounded fragments of McKim Formation siltstone and argillite mixed with large blocks and fragments of Nipissing Diabase and other rock types. In places, the finely ground dark grey matrix shows flow banding around the fragments.

Some of the youngest geological features on site were formed by the erosive power of ice sheets that intermittently covered northern Ontario between 10 000 and 190 000 years ago. These include glacial striations, which mark the surface of many of the outcrops, and a prominent glacial groove that can be seen beside the walkway from the parking lot to the building.

Photo S49. Dynamic Earth building with the Big Nickel on the left.

Stop 50. Science North **by Brenda Koziol** Paris Street and Ramsey Lake Road

Since Science North opened in 1984, it has grown to become Canada's second largest science centre and northern Ontario's largest tourist attraction. Science North is carved out of, and is perched on top of, the Precambrian bedrock of the Huronian Supergroup. Its dramatic architectural design features two snowflakes clad in stainless steel, linked by an underground tunnel with excellent exposures of fractured bedrock. The smaller snowflake houses a multipurpose room, cafeteria, restaurant and administrative offices. The large snowflake, the Exhibit Building, features the 10 m deep Inco Cavern at its base and a ramp to the exhibit floors designed to provide spectacular views of Ramsey Lake and the exposed bedrock within the building. A third building houses the lobby, gift store, IMAX® Theatre, motion simulator ride, and a special exhibitions hall.

Science North is internationally recognized for its style of science communication, innovative programming and interactive exhibits. Visitors to the science centre encounter live tropical butterflies, trade items from nature at Nature Exchange, interact with northern Ontario animals, and test the physical capabilities of their own bodies. They learn about astronomy and space exploration and experiment with robotics, computers and engineering technology. Science North's "Object Theatres" provide immersive multi-media shows about topics such as climate change, neutrinos and genomics.

Visitors to Science North can study and interpret interesting geological features evident right inside the buildings, including shatter cones, sedimentary bedding, joints, faults, glacial striations, and the three types of bedrock underlying the site: conglomerates of the Ramsay Lake Formation, subarkoses of the Mississagi Formation and a younger diabase dike. This allows for a great field trip, even in the dead of winter!

The bedrock on the south side of the site, underlying the small snowflake building and throughout the tunnel connecting the two snowflakes consists of 2.25 billion-year-old metamorphosed subarkose of the Mississagi Formation, part of the Hough Lake Group. The sediments were deposited in braided rivers that flowed southwest across an expansive barren continental landmass very different from present-day North America, transporting vast amounts of sand eroded from higher land to the northeast. Paleogeographic reconstructions indicate that the landmass straddled the equator, with Sudbury located in the Southern Hemisphere at subtropical latitudes.

On the rock face outside the window at the tunnel entrance, you can see well-preserved crossbedding in the Mississagi Formation. The crossbeds formed when ripples and sand bars on the riverbed migrated downstream, and sand cascaded down the front of the ripple or bar to create fine beds at an angle to the main bedding. The broad-scale layering of the sandstone, 30 to 50 cm thick units, is evident in the walls and roof of the tunnel. The beds are now sloping due to folding and deformation during mountain-building episodes. However, in most of the tunnel, detailed features of the orientation of the bedding are difficult to discern because of the low lighting and years of encrustation from water flowing through the fractures.

Watch for a well-developed shatter cone highlighted with a spotlight on the left side of the enclosed passage leading to the tunnel. It is about 15 cm long and looks like half an ice cream cone, with ridges and grooves radiating out from the cone's pointed end. Shatter cones are considered to be strong evidence that the Sudbury Structure was created by a meteorite impact. In the ceiling overhead of the shatter cone, a stained glass panel portrays a meteorite plunging toward the earth; a smaller stained glass panel illustrates an earlier theory of origin, by an immense volcanic explosion.

Joints are evident in the tunnel, particularly where the walkway widens near the end of the tunnel. Here, flat broken surfaces on the rock walls run roughly parallel to the tunnel and can be distinguished from the sandstone beds, which run across the roof of the tunnel. The joints are planes of weakness along which the rock tends to break, developed when the rock was subjected to tectonic forces.

As you leave the tunnel and enter the Exhibit Building, the mosaic of greenish-black polished rock fragments running east-west on the floor marks the fault line of the southern splay of the Creighton Fault. Here, the fault marks the contact between the older Mississagi Formation sandstone of the tunnel and the diabase dike. Movement along the fault has placed these two rock types adjacent to each other. The main splay of the Creighton Fault, seen by the elevator, is within the diabase dike. A closer examination reveals "fault gouge" or ground-up rock flour in the fault, formed when the rocks on both sides ground against each other as they moved along the fault. Towards the east, the fault line forms a dramatic crevasse that can be better viewed from the ramp to the second and third floors.

The Creighton Fault system can be traced for about 70 km, from Wahnapitae to the east through the Science North site to west of Naughton, where it connects with the larger Murray Fault system. The system was active between about 2 billion and 1 billion years ago. In general, it shows right-lateral movement, making it a strike-slip fault, similar in type to the famous San Andreas fault system in California. Right-lateral movement means that when you straddle the fault, the rocks on your right have moved toward you relative to the rocks on your left. Fieldwork carried out in the Sudbury area suggests movement of about 580 to 890 m laterally and 140 to 180 m vertically, with the south side moving up relative to the north side.

The diabase dike, which underlies the south part of the Exhibit Building including the Inco Cavern, formed when magma rose in cracks in the earth's crust from deep-seated chambers. Its age has been estimated as 1.4 billion years. The contact between the diabase dike and Ramsay Lake Formation (named before the spelling of Ramsey Lake was changed), which underlies the northern part of the site, is difficult to see in the building. A good exposure of Ramsay Lake Formation can be seen from the ramp windows on the first landing to the second floor. It is a matrix-supported conglomerate that contains pebbles and cobbles within the dark grey matrix, interpreted by some geologists to have formed by glacial processes more than 2 billion years ago. It lies stratigraphically below the Mississagi Formation; the Pecors Formation normally found between the two is not present at Science North because it has been displaced by movement of the Creighton Fault.

On this outcrop, you can also see a sharp black line two-thirds up the outcrop that marks the location of the soil level prior to construction of Science North. The black coating on the upper part of the rock is the result of pollution from early smelting processes that released clouds of sulphur dioxide gas and particulates at ground level from 1888 to 1929. The natural appearance of the bedrock was exposed when glacial till was removed from the lower part of the rockface during construction of Science North in the early 1980s.

The rock adjacent to the ramp up to the second and third floors shows evidence of erosion by continental ice sheets during the most recent glacial period, which ended about 10 000 years ago. Sculpting and polishing during glaciation smoothed and rounded the surface of the diabase dike. Two sets of striations at about 30° to each other mark two episodes of glacial advance. In the crevasse of the main splay of the Creighton Fault, the north wall is more jagged in appearance, has a steeper face and curves inward compared to the south wall. This is because the ice sheet plucked rock from the north wall of the crevasse as it advanced from northeast to southwest, and smoothed or polished the south wall, which faced the flow of the advancing ice.

The geological features seen within the building can also be observed outside, but the most striking views are of the Creighton Fault near the boat dock (Photo S50). Along the walkway down to the dock, watch for the fault contact of the light beige Mississagi Formation sandstone with the black diabase at the southern splay of the fault. The main splay of the fault is marked by a low area below the diabase cliff, an extension of the crevasse seen inside the building. Good exposures of the Ramsay Lake conglomerate can be seen along the boardwalk from the boat dock to Paris Street.

To represent regional geology, over-sized boulders and paving stones have been transported from other locations and installed on site. Most notable are the large blocks and paving stones from Manitoulin Island, located about 100 km southwest of Sudbury. These creamy-white rocks are dolomitic limestones, formed some 430 million years ago in warm shallow coral seas at the margin on the Michigan Basin, at a time when the Manitoulin Island area was at a tropical paleolatitude. They contain fossilized colonial coral and other fossils.

Photo S50. Aerial photo of Science North. See text for explanation.

REFERENCES

Abel, M.K. 1981. The structure of the Strathcona Mine copper zone; Canadian Institute of Mining and Metallurgy, Bulletin 74, p.89-97.

Abel, M.K., Buchan, R., Coats, C.J.A. and Penstone, M.E. 1979. Copper mineralization in the footwall complex, Strathcona mine, Sudbury, Ontario; Canadian Mineralogist, v.17, p.275-286.

Addison, W.D., Brumpton, G.R., Vallini, D.A., McNaughton, N.J., Davis, D.W., Kissin, S.A., Fralick, P.W. and Hammond, A.L. 2005. Discovery of distal ejecta from the 1850 Ma Sudbury impact event; Geology, v.33, no.3, p.193-196.

Ames, D.E. 1999. Geology and regional hydrothermal alteration of the crater-fill Onaping Formation: Association with Zn-Pb-Cu mineralization, Sudbury Structure Canada; unpublished PhD thesis, Carleton University, Ottawa, Ontario, 460p.

Ames, D.E., Bleeker, W., Heather, K.B. and Wodicka, N. 1997. Timmins to Sudbury transect: new insights into the regional geology and setting of mineral deposits; Geological Association of Canada–Mineral Association of Canada, Joint Annual Meeting, Ottawa, 1997, Field Trip Guidebook B6, 133p.

Ames, D.E., Davidson, A., Buckle, J.L. and Card, K.D. 2005. Geology, Sudbury bedrock compilation, Ontario; Geological Survey of Canada, Open File 4570, scale 1:50 000.

Ames, D.E. and Farrow, C.E.G. 2007. Metallogeny of the Sudbury mining camp; *in* Mineral Deposits of Canada, Geological Association of Canada, Mineral Deposits Division, Special Publication 5, p.329-350.

Ames, D.E. and Gibson, H.L. 2004a. Geology, alteration and mineralization of the Onaping Formation, Morgan Township, Sudbury Structure; Geological Survey of Canada, Open File 3717.

——— 2004b. Geology, alteration and mineralization of the Onaping Formation, Rockcut Lake area, Norman Township, Sudbury, Ontario; Geological Survey of Canada, Open File 4565.

——— 2004c. Geology, alteration and mineralization of the Onaping Formation, Joe Lake area, Wisner Township, Sudbury, Ontario; Geological Survey of Canada, Open File 4566.

——— 2004d. Geology, alteration and mineralization of the Onaping Formation, Simmons Lake area, Dowling Township, Sudbury, Ontario; Geological Survey of Canada, Open File 4567.

——— 2004e. Geology, alteration and mineralization of the Onaping Formation, Cow Lake area, Dowling Township, Sudbury, Ontario; Geological Survey of Canada, Open File 4568.

Ames, D.E., Golightly, J.P., Lightfoot, P.C. and Gibson, H.L. 2002. Vitric compositions in the Onaping Formation and their relationship to the Sudbury Igneous Complex, Sudbury Structure; *in* Geology and Ore Deposits of the Sudbury Structure, Economic Geology, v.97, p.1541-1562.

Ames, D.E., Jonasson, I.R., Gibson, H.L. and Pope, K.O. 2005. Impact-generated hydrothermal system – constraints from the large Paleoproterozoic Sudbury crater, Canada; *in* Biological Processes Associated with Impact Events, Impact Studies, Springer-Verlag, Berlin-Heidelberg, p.55-100.

Ames, D.E., Watkinson, D.H. and Parrish, R.R. 1998. Dating of a regional hydrothermal system induced by the 1850 Ma Subury impact event; Geology, v.26, p.447-450.

Amsden, A.A., Ruppel, H.M. and Hirt, C.W. 1980. SALE: A simplified ALE computer program for fluid flow at all speeds; Los Alamos National Laboratory Report LA-8095, Los Alamos, New Mexico, 101p.

Angelier, J. 1994. Fault slip analysis and paleostress reconstruction; *in* Continental Deformation, Pergamon Press, Oxford, p.53-100.

Arengi, J.T. 1977. Sedimentary evolution of the Sudbury Basin; unpublished MSc thesis, University of Toronto, Toronto, Ontario, 141p.

Ariskin, A.A. 1997. COMAGMAT-3.5: A renewed programme for the calculation of magma differentiation processes; Vernadsky Institute, Moscow, Russia.

Ariskin, A.A., Deutsch, A. and Ostermann, M. 1999. Sudbury Igneous Complex: Simulating phase equilibria and *in situ* differentiation for two proposed parental magmas; *in* Large Meteorite Impacts and Planetary Evolution II, Geological Society of America, Special Paper 339, p.373-388.

Aspler, L.B. and Chiarenzelli, J.R. 1998. Two Neoarchean supercontinents? Evidence from the Paleoproterozoic; Sedimentary Geology, v.120, p.75-104.

Avermann, M.E. 1999. The green member of the Onaping Formation, the collapsed fireball layer of the Sudbury impact structure, Ontario, Canada; *in* Large Meteorite Impacts and Planetary Evolution II, Geological Society of America, Special Paper 339, p.323-330.

Avermann, M.E. and Brockmeyer, P. 1992. The Onaping Formation of the Sudbury Structure, Canada: an example of allochthonous impact breccias; Tectonophysics, v.216, p.227-234.

Bailey, J., Lafrance, B., McDonald, A.M., Fedorowich, J.S., Kamo, S. and Archibald, D.A. 2004. Mazatzal–Labradorian-age (1.7-1.6 Ga) ductile deformation of the South Range Sudbury impact structure at the Thayer Lindsley Mine, Ontario; Canadian Journal of Earth Sciences, v.41, p.1491-1505.

Barlow, A.E. 1907. Origin, geological relations and compositions of the nickel and copper deposits of the Sudbury Mining District, Ontario; *in* Geological Survey of Canada, Annual Report 14 (for 1901), p.1-236 (No. 873). Revised and reprinted 1907.

Barnes, F.Q. and Lalonde, E.J. 1973. Lower Huronian stratigraphy, Hyman and Drury townships, Sudbury District; *in* Huronian Stratigraphy and Sedimentation, Geological Association of Canada, Special Paper Number 12, p.147-156.

Beales, F.W. and Losej, G.P. 1975. Sudbury Basin sediments and meteorite impact theory of origin for the Sudbury Structure; Canadian Journal of Earth Sciences, v.12, p.629-635.

Becker, L., Poreda, R.J. and Bada, J.L. 1996. Extraterrestrial helium trapped in fullerenes in the Sudbury impact structure; Science, v.272, no.5259, p.249-252.

Bell, R. 1891a. On the Sudbury Mining District; *in* Geological Survey of Canada, Annual Report, New Series, 5, Part 1, p.5F-95F

——— 1891b. The nickel and copper deposits of Sudbury District, Canada; Geological Society of America Bulletin, v.2, p.125-140.

Bennett, G., Dressler, B.O. and Robertson, J.A. 1991. The Huronian Supergroup and associated intrusive rocks; *in* Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 1, p.549-591.

Bennett, G. and Meyer, W. 1995. Stratigraphic relationships in the lower Huronian of the Sault Ste. Marie–Elliot Lake areas, and possible implications regarding early Huronian tectonics; The Canadian Mineralogist, v.33, part 4, p.924-925.

Bhattacharji, S. and Smith, C.H. 1964. Flowage differentiation; Science, v.145, p.150-153.

Binney, W.P., Poulin, R.Y., Sweeny, J.M. and Halladay, S.H. 1994. The Lindsley Ni-Cu-PGE deposit and its geological setting; *in* Proceedings of the Sudbury – Noril'sk Symposium, Ontario Geological Survey, Special Volume 5, p.91-103.

Blue, A. 1894. The Creighton gold mine; *in* Ontario Bureau of Mines, Annual Report for 1993, v.3, p.45-47.

Boast, M. and Spray, J.G. 2006. Superimposition of a thrust-transfer fault system on a large impact structure: implications for Ni-Cu-PGE exploration at Sudbury; Economic Geology, v.101, p.1583-1594.

Bonney, T.G. 1888. Notes on a part of the Huronian series in the neighbourhood of Sudbury, Canada; Quarterly Journal of the Geological Society of London, v.44, p.32-45.

Bouma, A.H. 1962. Sedimentology of some Flysch Deposits; Elsevier, Amsterdam, London, New York, 168p.

Brocum, S.J. and Dalziel, I.W. 1974. The Sudbury basin, the Southern Province, the Grenville Front, and the Penokean Orogeny; Geological Society of America Bulletin, v.85, p.1571-1580.

Buchan, K.L., Card, K.D. and Chandler, F.W. 1989. Paleomagnetism of Nipissing diabase and associated rocks in the Englehart area, Ontario; Canadian Journal of Earth Sciences, v.26, p.427-445.

Burrows, A.G. and Rickaby, H.C. 1930. Sudbury Basin area; 38th Annual Report of the Ontario Department of Mines, v.38, pt.3, 55p.

Bussolaro, N., Rousell, D.H. and Beswick, A.E. 1984. Mineralization in the Onaping Formation, Sudbury Basin, Ontario; Geological Survey of Canada, Open File Report 1090, 52p.

Butler, H.R. 1994. Lineament analysis of the Sudbury multi-ring impact structure; *in* Large Meteorite Impacts and Planetary Evolution, Geological Society of America, Special Paper 293, p.319-329.

Cantin, R. and Walker, W.G. 1972. Was the Sudbury Basin circular during deposition of the Chelmsford Formation?; *in* New Developments in Sudbury Geology, Geological Association of Canada, Special Paper Number 10, p.93-101.

Capes, C. 2001. A petrological investigation of the Copper Cliff embayment structure, Sudbury, Canada; unpublished MSc thesis, University of Toronto, Toronto, Ontario, 162p.

Card, K.D. 1964. Metamorphism in the Agnew Lake area, Sudbury District, Ontario, Canada; Geological Society of America Bulletin, v.75, p.1011-1030.

——— 1965. Geology of Hyman and Drury townships; Ontario Department of Mines, Geological Report 34, 38p.

——— 1967. Sudbury Mining area; Ontario Department of Mines, Map 2170, scale 1:63 360.

——— 1968. Geology of Denison–Waters area; Ontario Department of Mines, Geological Report 60, 63p.

——— 1978a. Metamorphism of the middle Precambrian (Aphebian) rocks of the eastern Southern Province; *in* Metamorphism in the Canadian Shield, Geological Survey of Canada, Paper 78-10, p.269-282.

——— 1978b. Geology of the Sudbury–Manitoulin area; Ontario Geological Survey, Report 166, 238p.

——— 1979. Regional geological synthesis, central Superior Province; *in* Current Research, Geological Survey of Canada, Paper 79-1A, p.87-90.

——— 1994. Geology of the Levack Gneiss Complex, the northern footwall of the Sudbury Structure, Ontario; *in* Current Research, Geological Survey of Canada, Paper 1994-C, p.269-278.

Card, K.D., Church, W.R., Franklin, J.M., Frarey, M.J. and Robertson, J.A. 1972. The Southern Province; *in* Variations in Tectonic Styles in Canada, Geological Association of Canada, Special Paper Number 11, p.335-380.

Card, K.D., Gupta, V.K., McGrath, P.H. and Grant, F.S. 1984. The Sudbury Structure: its regional geological and geophysical setting; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.25-43.

Card, K.D. and Hutchison, R.W. 1972. The Sudbury structure: its regional geological setting; *in* New Developments in Sudbury Geology, Geological Association of Canada, Special Paper Number 10, p.67-78.

Card, K.D. and Innes, D.G. 1981. Geology of the Benny area, District of Sudbury; Ontario Geological Survey, Report 206, 117p.

Card, K.D., Innes, D.G. and Debicki, R.L. 1977. Stratigraphy, sedimentology and petrology of the Huronian Supergroup in the Sudbury–Espanola area; Ontario Division of Mines, Geoscience Study 16, 99p.

Card, K.D. and Jackson, S.L. 1995. Tectonics and metallogeny of the Early Proterozoic Huronian fold belt and the Sudbury Structure of the Canadian Shield; Geological Survey of Canada, Open File 3139, Field Trip Guidebook, 55p.

Card, K.D. and Meyn, H.D. 1969. Geology of the Leinster–Bowell area, District of Sudbury; Ontario Department of Mines, Geological Report 65, 40p.

Card, K.D. and Pattison, E.F. 1973. Nipissing diabase of the Southern Province, Ontario; *in* Huronian Stratigraphy and Sedimentation, Geological Association of Canada, Special Paper Number 12, p.7-30.

Card, K.D. and Poulsen, K.H. 1998. Geology and mineral deposits of the Canadian Shield; *in* Geology of the Precambrian Superior and Grenville Provinces and Precambrian Fossils in North America, Geological Survey of Canada, Geology of Canada, v.7, p.205-270.

Card, K.D. and Wodicka, N. in prep. Geology, geochronology, and tectonic history of the Levack Gneiss Complex, the northern footwall of the Sudbury Igneous Complex, Ontario; Geological Survey of Canada, Open File Report with Map 4266.

Carmichael, D.M. 1978. Metamorphic bathozones and bathograds: a measure of the depth of uplift and erosion on a regional scale; American Journal of Science, v.278, p.769-797.

Chai, G. and Eckstrand, O.R. 1994. Rare earth element characteristics of the Sudbury Igneous Complex and its country rocks: new constraints on genesis; Chemical Geology, v.113, no.3-4, p.221-244.

Choudhry, A.G. 1984. Geology of Hart, Ermatinger and Totten townships; Ontario Geological Survey, Open File Report 5525, 75p.

Chyi, L.L. and Crocket, J.H. 1976. Partition of platinum, palladium, iridium and gold among coexisting minerals from the deep ore zone, Strathcona Mine, Sudbury, Ontario; Economic Geology, v.71, p.1196-1205.

Coats, C.J.A. and Snajdr, P. 1984. Ore deposits of the North Range, Onaping–Levack area, Sudbury; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.327-346.

Cochrane, L.B. 1983. The Creighton Fault; unpublished report prepared for Science North (Sudbury), 51p.

——— 1984. Ore deposits of the Copper Cliff Offset; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.347-359.

——— 1991. Analysis of the structural and tectonic environments associated with rock-mass failures in the mines of the Sudbury District; unpublished PhD thesis, Queen’s University, Kingston, Ontario, 264p.

Coleman, A.P. 1905. The Sudbury Nickel Region; Report of the Ontario Bureau of Mines, v.14, pt.3, 183p.

——— 1913. The nickel industry, with special reference to the Sudbury Region, Ontario; Canada Department of Mines, Mines Branch, no.170, 206p.

Coleman, A.P., Moore, E.S. and Walker, T.L. 1929. The Sudbury nickel intrusive; University of Toronto Studies, Geology Section 28, p.1-54.

Collins, W.H. 1934. Life history of the Sudbury nickel irruptive (1): Petrogenesis; Royal Society of Canada, Transactions, Third Series, v.28, section 4, p.123-177.

——— 1935. Life history of the Sudbury nickel irruptive (2): Intrusion and Deformation; Royal Society of Canada, Transactions, Third Series, v.29, section 4, p.27-47.

——— 1936. Life history of the Sudbury nickel irruptive (3): Environment; Royal Society of Canada, Transactions, Third Series, v.30, section 4, p.29-53.

——— 1937. Life history of the Sudbury nickel irruptive (4): Mineralization; Royal Society of Canada, Transactions, Third Series, v.31, section 4, p.15-43.

Conrod, D.M. 1988. Petrology, geochemistry, and PGE potential of the Nipissing intrusions; unpublished MSc thesis, University of Toronto, Toronto, Ontario, 494p.

——— 1989. The petrology and geochemistry of the Ducan Lake, Beaton Bay, Milner Lake and Miller Lake Nipissing intrusions within the Gowganda area, District of Timiskiming; Ontario Geological Survey, Open File Report 5701, 210p.

Cooke, H.C. 1946. Problems of Sudbury geology, Ontario; Canada Department of Mines, Geological Survey Bulletin 3, 77p.

——— 1948. Regional structure of the Lake Huron–Sudbury area; *in* Structural Geology of Canadian Ore Deposits, Canadian Institute of Mining and Metallurgy, Jubilee Volume, p.580-589.

Corfu, F. and Andrews, A. 1986. A U-Pb age for mineralized Nipissing diabase, Gowganda, Ontario; Canadian Journal of Earth Sciences, v.23. p.107-112.

Cowan, E.J. 1996. Deformation of the eastern Sudbury Basin; unpublished PhD thesis, University of Toronto, Toronto, Ontario, 332p.

Cowan, E.J., Riller, U. and Schwerdtner, W.M. 1999. Emplacement geometry of the Sudbury Igneous Complex: structural examination of a proposed impact melt sheet; *in* Large Meteorite Impacts and Planetary Evolution II, Geological Society of America, Special Paper 339, p.399-418.

Cowan, E.J. and Schwerdtner, W.M. 1994. Fold origin of the Sudbury Basin; *in* Procedings of the Sudbury – Noril'sk Symposium, Ontario Geological Survey, Special Volume 5, p.45-55.

Davidson, S. 1948. Falconbridge Mine; *in* Structural Geology of Canadian Ore Deposits, Canadian Institute of Mining and Metallurgy, Jubilee Volume, p.618-626.

Davis, G.C. 1984. Little Stobie Mine: A South Range contact deposit; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.361-369.

Debicki, R.L. 1990. Stratigraphy, paleoenvironment and economic potential of the Huronian Supergroup in the southern Cobalt Embayment; Ontario Geological Survey, Miscellaneous Paper 148, 154p.

Dence, M.R. 1972. Meteorite impact craters and the origin of the Sudbury Basin; *in* New Developments in Sudbury Geology, Geological Association of Canada, Special Paper Number 10, p.7-18.

——— 2004. Structural evidence for shock metamorphism in simple and complex craters: Linking observations to theory; Meteoritics & Planetary Science 39, no.2, p.267-286.

Desborough, G.A. and Larsen, R.R. 1970. Nickel-bearing iron sulphides in the Onaping Formation, Sudbury Basin, Ontario; Economic Geology, v.75, p.728-730.

Deutsch, A., Grieve, R.A.F., Avermann, M., Bischoff, L., Brockmeyer, B., Buhl, D., Lakomy, R., Müller-Mohr, R., Ostermann, M. and Stoffer, D. 1995. The Sudbury Structure (Ontario, Canada): a tectonically deformed multi-ring impact basin; Geologische Rundschau 84, p.697-709.

Dickin, A.P., Nguyen, T. and Crocket, J.H. 1999. Isotope evidence for a single impact melting origin of the Sudbury Igneous Complex; *in* Large Meteorite Impacts and Planetary Evolution II, Geological Society of America, Special Paper 339, p.361-371.

Dickin, A.P., Richardson, J.M., Crocket, J.H., McNutt, R.H. and Peredery, W.V. 1992. Osmium isotope evidence for a crustal origin of platinum group elements in the Sudbury nickel ore, Ontario, Canada; Geochimica et Cosmochimica Acta, v.56, p.3531-3537.

Dickson, C.W. 1903. The ore deposits of Sudbury; Transactions of the American Institute of Mining Engineers, v.34, p.1-67.

Dietz, R.S. 1962. Sudbury Structure as an astrobleme; Transactions of the American Geophysical Union, v.43, p.445-446.

——— 1964. Sudbury Structure as an astrobleme; Journal of Geology, v.72, p.412-434.

——— 1972. Sudbury astrobleme, splash-emplaced sub-layer and possible cosmic ores; *in* New Developments in Sudbury Geology, Geological Association of Canada, Special Paper Number 10, p.29-40.

Dressler, B.O. 1979. Geology of McNish and James townships, District of Sudbury; Ontario Geological Survey, Report 191, 91p.

——— 1981. Precambrian geology of Levack Township, District of Sudbury; Ontario Geological Survey, Preliminary Map P.2428, scale 1:15 840.

——— 1982. Geology of the Wanapitei Lake area, District of Sudbury; Ontario Geological Survey, Report 213, 131p.

——— 1984a. General geology of the Sudbury area; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.57-82.

——— 1984b. The effects of the Sudbury Event and the intrusion of the Sudbury Igneous Complex on the footwall rocks of the Sudbury Structure; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.97-136.

——— 1984c. Sudbury geological compilation; Ontario Geological Survey, Map 2491, scale 1:50 000.

Dressler, B.O., Gupta, V.K. and Muir, T.L. 1991. The Sudbury Structure; *in* Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 1, p.593-625.

Dressler, B.O., Peredery, W.V. and Muir, T.L. 1992. Geology and mineral deposits of the Sudbury Structure; Ontario Geological Survey, Guidebook 8, 33p.

Dressler, B.O. and Reimold, W.U. 2001. Terrestrial impact melt rocks and glasses; Earth Science Reviews, v.56, p.205-284.

Duke, M.J. and Naldrett, A.J. 1976. Sulfide mineralogy of the Main Irruptive, Sudbury, Ontario; Canadian Mineralogist, v.14, p.450-461.

Dutch, S.I. 1979. The Creighton Pluton, Ontario: An unusual example of a forcefully emplaced intrusion; Canadian Journal of Earth Sciences, v.16, p.333-349.

Easton, R.M. 2000. Metamorphism of the Canadian Shield, Ontario, Canada; *in* Proterozoic Metamorphic History, The Canadian Mineralogist, p.319-344.

——— 2006. Geology and mineral potential of the Southern Province in Baldwin Township; *in* Summary of Field Work and Other Activities 2006, Ontario Geological Survey, Open File Report 6192, p.14-1 to 14-22.

Easton, R.M. and Berman, R.G. 2004. Tectonometamorphic map of Ontario, Canada and parts of the United States of America; Geological Survey of Canada, Open File 1810, Ontario Geological Survey, Preliminary Map P.3533, scale 1:500 000.

Elsila, J.E., DeLeon, N.P., Plows, F.L., Buseck, P.R. and Zare, R.N. 2005. Extracts of impact breccia samples from Sudbury, Gardnas and Ries impact craters and the effects of aggragation on C60 detection; Geochemica et Cosmochemica Acta, v.67, p.2891-2899.

Faggart, B.E., Basu, A.B. and Tatsumoto, M.M. 1985. Origin of the Sudbury Complex by meteorite impact: neodymium isotopic evidence; Science, v.230, p.436-439.

Fahrig, W.F. and West, T.D. 1986. Diabase dike swarms of the Canadian Shield; Geological Survey of Canada, Map 1627A, scale 1:4 873 900.

Fairbairn, H.W., Faure, G., Pinson, W.H. and Hurley, P.M. 1968. Rb-Sr whole-rock age of the Sudbury Lopolith and basin sediments; Canadian Journal of Earth Sciences, v.5, p.707-714.

Fairbairn, H.W. and Robson, G.M. 1942. Breccia at Sudbury, Ontario; Journal of Geology, v.50, p.1-33.

Farrow, C.E.G. and Lightfoot, P.C. 2002. Sudury PGE revisited: Toward an integrated model; *in* The Geology, Geochemistry, Mineralogy and Mineral Beneficiation of Platinum Group Elements, Canadian Institute of Mining, Metallurgy and Petroleum, Special Volume 54, p.273-297.

Farrow, C.E.G. and Watkinson, D.H. 1992. Alteration and the role of fluids in Ni, Cu and platinum-group element deposition, Sudbury Igneous Complex contact, Onaping–Levack area, Ontario; Mineralogy and Petrology, v.46, p.67-83.

——— 1997. Diversity of precious metal mineralization in footwall Cu-Ni-PGE deposits, Sudbury, Ontario: Implications for hydrothermal models of formation; Canadian Mineralogist, v.35, p.817-839.

——— 1999. An evaluation of the role of fluids in Ni-Cu-PGE-bearing, mafic-ultramafic systems; *in* Dynamic Processes in Magmatic Ore Deposits and their Application in Mineral Exploration, Geological Association of Canada, Short Course Notes, v.13, p.31-67.

Fedorowich, J.S. and Nickolic, B. 1999. Field trip B1 Sudbury Ni-Cu-PGE deposits – North Range; *in* Sudbury Ni-Cu-PGE deposits – South Range (A1) and North Range (B1), Geological Association of Canada–Mineralogical Association of Canada, Joint Annual Meeting, Sudbury, Ontario, Field Trips A1 and B1 Guidebook, p.25-44.

Fedorowich, J.S., Rousell, D.H. and Peredery, W.V. 1999. Sudbury breccia distribution and orientation in an embayment environment; *in* Large Meteorite Impacts and Planetary Evolution II, Geological Society of America, Special Paper 339, p.305-315.

Finn, G.C., Edgar, A.D. and Rowell, W.F. 1982. Petrology, geochemistry, and economic potential of the Nipissing diabase; *in* Geoscience Research Grant Program, Summary of Research 1981-1982, Ontario Geological Survey, Miscellaneous Paper 103, p.43-57.

Fleet, M.E., Barnett, R.L. and Morris, W.A. 1987. Prograde metamorphism of the Sudbury Igneous Complex; Canadian Mineralogist, v.25, p.499-514.

Fox, J.S. 1971. Coexisting chloritoid and staurolite and the staurolite-chlorite isograd from the Agnew Lake area, Ontario, Canada; Geological Magazine, 108, p.205-219.

Frarey, M.J., Loveridge, W.D. and Sullivan, R.W. 1982. A U-Pb age for the Creighton granite, Ontario; *in* Rb-Sr and U-Pb Isotope Age Studies, Report 5, Current Research, Part C, Geological Survey of Canada, Paper 82-C, p.129-132.

French, B.M. 1967. Sudbury Structure, Ontario: some petrographic evidence for origin by meteorite impact; Science, v.156, p.1094-1098.

——— 1968. Sudbury Structure, Ontario: some petrographic evidence for origin by meteorite impact; *in* Shock Metamorphism of Natural Materials, Mono Book Corporation, Baltimore, p.383-412.

——— 1972. Shock-metamorphic features in the Sudbury Structure, Ontario; *in* New Developments in Sudbury Geology, Geological Association of Canada, Special Paper Number 10, p.19-28.

Fueton, F. and Redmond, D.J. 1992. Structural studies in the Southern Province, south of Sudbury, Ontario; *in* Current Research, Part C, Geological Survey of Canada, Paper 92-1C, p.179-187.

——— 1997. Documentation of a 1450 Ma contractional orogeny preserved between the 1850 Ma Sudbury impact structure and the 1 Ga Grenville orogenic front, Ontario; Geological Society of America Bulletin, v.109, p.268-279.

Gates, B.I. 1991. Sudbury mineral occurrence study; Ontario Geological Survey, Open File Report 5771, 235p.

Gibbins, S.F.M. 1994. Geology, geochemistry, stratigraphy and mechanisms of emplacement of the Onaping Formation, Dowling area, Sudbury Structure, Ontario, Canada; unpublished MSc thesis, Laurentian University, Sudbury, Ontario, 315p.

——— 1997. The Sudbury Structure with emphasis on the Whitewater Group; Institute on Lake Superior Geology, 43rd Annual Meeting, Sudbury, Field Trip Guidebook, v.43, part 4, 53p.

Giblin, P.E. 1984. History of exploration and development, of geological studies and development of geological concepts; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.3-23.

Golightly, J.P. 1994. The Sudbury Igneous Complex as an impact melt: evolution and ore genesis; *in* Proceedings of the Sudbury – Noril'sk Symposium, Ontario Geological Survey, Special Volume 5, p.105-117.

Golightly, J.P. and Rousell, D.H. 2000. Nature and origin of the Sudbury Structure and the Ni-Cu-PGE ores; *in* Geology and Ore Deposits 2000: the Great Basin and Beyond, Proceedings, Geological Society of Nevada, p.37-60.

Grant, R.W. and Bite, A. 1984. Sudbury quartz diorite offset dikes; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.275-300.

Gray, M.J. 1995. The geological setting of the Vermilion Zn-Cu-Pb-Ag-Au massive sulphide deposit; unpublished MSc thesis, Laurentian University, Sudbury, Ontario, 244p.

Greenman, L.C. 1970. The petrology of the footwall breccias in the vicinity of the Strathcona mine, Levack, Ontario; unpublished PhD thesis, University of Toronto, Toronto, Ontario, 153p.

Gregory, S.K. 2005. Geology, mineralogy and geochemistry of transitional contact / footwall mineralization in the McCreedy East Ni-Cu-PGE deposit, Sudbury Igneous Complex; unpublished MSc thesis, Laurentian University, Sudbury, Ontario, 138p.

Grieve, R.A.F. 1994. An impact model of the Sudbury Structure; *in* Proceedings of the Sudbury – Noril'sk Symposium, Ontario Geological Survey, Special Volume 5, p.119-132.

Grieve, R.A.F., Stoffler, D. and Deutsch, A. 1991. The Sudbury Structure: controversial or misunderstood?; Journal of Geophysical Research, v.96, p.22753-22764.

Guy-Bray, J.G. 1966. Shatter cones at Sudbury; Journal of Geology, v.74, p.243-245.

——— 1972. Introduction; *in* New Developments in Sudbury Geology, Geological Association of Canada, Special Paper Number 10, p.1-5.

Hambrey, M.J. 1994. Glacial Environments; University of British Columbia Press, Vancouver, 296p.

Hanley, J.J., Mungall, J.E., Bray, C.J. and Gordon, P. 2004. The origin of bulk and water-soluble Cl and Br enrichments in ore-hosting Sudbury breccia in the Fraser copper zone, Strathcona embayment, Sudbury, Ontario, Canada; Canadian Mineralogist, v.42, p.1297-1318.

Hawley, J.E. 1962. The Sudbury ores: their mineralogy and origin; The Canadian Mineralogist, v.7, pt.1, 207p.

——— 1965. Upside-down zoning at Frood, Sudbury, Ontario; Economic Geology, v.60, p.529-575.

Heaman, L.M. 1989. U-Pb dating of mafic dyke swarms: what are the options?; abstract *in* International Association of Volcanology and Chemistry of the Earth's Interior, Continental Magmatism, New Mexico Bureau of Mines and Mineral Resources, Bulletin 131, p.125.

——— 1995. U-Pb dating of mafic rocks: past, present and future; Geological Association of Canada–Mineralogical Association of Canada, Program with Abstracts, v.20, p.A43.

Herget, G. 1977. Structural Geology; *in* Pit Slope Manual, Chapter 2, Canada Centre for Mineral and Energy Technology, Canmet Report 77-41, 123p.

Hocking, M.W.A. 2003. The geology and geochemistry of the Elsie Mountain Formation: implications for Paleoproterozoic magmatism in the Southern Province; unpublished BSc thesis, Laurentian University, Sudbury, Ontario, 79p.

Hoffman, E.L., Naldrett, A.J., Alcock, R.A. and Hancock, R.G.V. 1979. The noble metal content of ore in the Levack West and Little Stobie mines, Ontario; Canadian Mineralogist, v.17, p.437-452.

Holloway, G.T., Miller, W.G., Young, M. and Gibson, T.W. 1917. Report of the Royal Ontario Nickel Commission, Toronto, Ontario, 584p.

Hriskevich, M.E. 1968. Petrology of the Nipissing Diabase sill of the Cobalt area, Ontario, Canada; Geological Society of America Bulletin, v.79, p.1387-1404.

Innes, D.G. 1978a. Proterozoic volcanism in the Southern Province of the Canadian Shield; unpublished MSc thesis, Laurentian University, Sudbury, Ontario, 161p.

——— 1978b. McKim Township, District of Sudbury; Ontario Geological Survey, Preliminary Map P.1978, scale 1:15 840.

Innes, D.G. and Colvine, A.C. 1979. Metallogenic development of the eastern part of the Southern Province in Ontario; *in* Summary of Field Work 1979, Ontario Geological Survey, Miscellaneous Paper 90, p.184-189.

——— 1984. The regional metallogenic setting of Sudbury; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.45-56.

Ivanov, B.A. and Deutsch, A. 1999. Sudbury impact event: cratering mechanism and thermal history; *in* Large Meteorite Impacts and Planetary Evolution II, Geological Society of America, Special Paper 339, p.389-398.

Jackson, S.L. 2001. On the structural geology of the Southern Province between Sault Ste. Marie and Espanola, Ontario; Ontario Geological Survey, Open File Report 5995, 55p.

Jackson, S.L. and Fyon, J.A. 1991. The western Abitibi Subprovince in Ontario; *in* Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 1, p.405-482.

Jambor, J.L. 1971. The Nipissing Diabase; Canadian Mineralogist, v.42, p.743-770.

James, R.S. and Dressler, B.O. 1992. Nature and significance of the Levack Gneiss Complex – footwall rocks of the North and East Ranges of the Sudbury Igneous Complex; *in* The Sudbury – Noril'sk Symposium, Program and Abstracts, p.487.

James, R.S., Easton, R.M., Peck, D.C. and Hrominchuk, J.L. 2002. The East Bull Lake intrusive suite: remnants of an approximately 2.48 Ga large igneous and metallogenic province in the Sudbury area of the Canadian Shield; Economic Geology, v.97, p.1577-1606.

James, R.S., Jobin-Bevans, S., Easton, R.M., Wood, P., Hrominchuk, J.L., Keays, R.R. and Peck, D.C. 2002. Platinum-group element mineralization in Paleoproterozoic basic intrusions in central and northeastern Ontario, Canada; *in* The Geology, Geochemistry, Mineralogy and Mineral Beneficiation of the Platinium Group Elements, Canadian Institute of Mining, Metallurgy and Petroleum, Special Volume 54, p.339-365.

James, R.S., Sweeny, J.M. and Peredery, W.V. 1992. Thermobarometry of the Levack gneisses – footwall rocks to the Sudbury Igneous Complex; *in* Lithoprobe, Abitibi–Grenville Project, Abitibi–Grenville Transect, Report 32, p.179-182.

Jobin-Bevans, L.S. 2004. Platinum-group mineralization in Nipissing Gabbro intrusions and the River Valley intrusion, Sudbury Region, Ontario; unpublished PhD thesis, University of Western Ontario, London, Ontario, 573p.

Jobin-Bevans, L.S., Keays, R.R. and MacRae, N.D. 1999. Cu-Ni-PGE in Nipissing Diabase: Results from surface and core samples; *in* Summary of Field Work and Other Activities 1999, Ontario Geological Survey, Open File Report 6000, p.33-1 to 33-5.

Jobin-Bevans, L.S., MacRae, N.D. and Keays, R.R. 1998. Cu-Ni-PGE potential of the Nipissing Diabase; *in* Summary of Field Work and Other Activities 1998, Ontario Geological Survey, Miscellaneous Paper 169, p.220-223.

Jolly, W.T., Dickin, A.P. and Wu, T.W. 1992. Geochemical stratigraphy of the Huronian continental volcanics at Thessalon, Ontario: contributions of two-stage crustal fusion; Contributions to Mineralogy and Petrology, v.110, p.411-428.

Keays, R.R. and Crocket, J.H. 1970. A study of precious metals in the Sudbury Nickel Irruptive ores; Economic Geology, v.65, p.438-450.

Keays, R.R. and Lightfoot, P.C. 2004. Formation of Ni-Cu-Platinum Group sulfide mineralization in the Sudbury impact melt sheet; Mineralogy and Petrology, v.82, p.217-258.

Kellet, R.L. and Rivard, B. 1996. Characterization of the Benny deformation zone, Sudbury, Ontario; Canadian Journal of Earth Sciences, v.33, p.1256-1267.

Killick, A.M. and Reimold, W.U. 1990. Review of the pseudotachylites in and around the Vredefort "Dome", South Africa; South African Journal of Geology, v.93, p.350-365.

Knight, C.W. 1917. Geology of the Sudbury area and description of Sudbury ore bodies; *in* Report of the Royal Ontario Nickel Commission, 1917, p.104-211.

——— 1923. The chemical composition of the norite-micropegmatite, Sudbury, Ontario; Economic Geology, v.18, p.529-594.

Krauskopf, K.B. 1979. Source rocks for metal-bearing fluids; *in* Geochemistry of Hydrothermal Ore Deposits, Holt, Rinehart and Winston, p.1-33.

Kretz, R. 1983. Symbols for rock-forming minerals; American Mineralogist, v.68, p.277-279.

Krogh, T.E., Corfu, F., Davis, D.W., Dunning, G.R., Heaman, L.M., Kamo, S.L., Machado, N., Greenough, J.D. and Nakamura, E. 1987. Precise U-Pb isotope ages of diabase dikes and mafic to ultramafic rocks using trace amounts of baddeleyite and zircon; *in* Mafic Dike Swarms, Geological Association of Canada, Special Paper 34, p.147-152.

Krogh, T.E., Davis, D.W. and Corfu, F. 1984. Precise U-Pb zircon and baddeleyite ages for the Sudbury area; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.431-446.

Krogh, T.E., Kamo, S.L. and Bohor, B.F. 1996. Shocked metamorphosed zircons with correlated U-Pb discordance and melt rocks with concordant protolith ages indicate an impact origin for the Sudbury Structure; *in* Earth Processes: Reading the Isotope Code, American Geophysical Union, Monograph 95, p.343-352.

Kuo, H.-Y. and Crocket, J.H. 1979. Rare earth elements in the Sudbury Nickel Irruptive: comparison with layered gabbros and implications for Nickel Irruptive petrogenesis; Economic Geology, v.79, p.590-605.

Kusky, T.M. 1993. Collapse of Archean orogens and the generation of late- to post-kinematic granitoids; Geology, v.21, p.925-928.

Lafleur, J. 1981. Cascaden, Dowling and Trill townships, District of Sudbury; *in* Summary of Field Work 1981, Ontario Geological Survey, Miscellaneous Paper 100, p.80-83.

Lafleur, J. and Dressler, B.O. 1985. Geology of Cascaden, Dowling, Levack and Trill townships, District of Sudbury; Ontario Geological Survey, Open File Report 5533, 135p.

Lakomy, R. 1990. Implications for cratering mechanics from a study of the Footwall Breccia of the Sudbury Impact Structure, Canada; Meteoritics, v.25, p.195-207.

Langford, F.F. 1960. Geology of Levack Township and the northern part of Dowling Township, District of Sudbury; Ontario Department of Mines, Preliminary Report 1960-5, 78p.

Legault, D., Lafrance, B. and Ames, D.E. 2003. Structural study of the Sudbury breccia and sulphide veins, Levack embayment, North Range of the Sudbury Structure, Ontario; *in* Current Research, Geological Survey of Canada, Paper 2003-C1, p.1-9.

Lesher, C.M. and Keays. R.R. 2002. Komatiite-associated Ni-Cu (PGE) deposits: mineralogy, geochemistry and genesis; *in* The Geology, Geochemistry, Mineralogy and Mineral Beneficiation of the Platinium Group Elements, Canadian Institute of Mining, Metallurgy and Petroleum, Special Volume 54, p.579-617.

Li, C. and Naldrett, A.J. 1993. High chlorine alteration minerals and calcium-rich brines in fluid inclusions from Strathcona Deep Copper Zone, Sudbury, Ontario; Economic Geology, v.88, p.1780-1796.

Li, C. and Ripley, E.M. 2005. Empirical equations to predict the sulfur content of mafic magmas at sulfide saturation and applications to magmatic sulfide deposits; Mineralium Deposita, v.40, p.218-230.

Lightfoot, P.C., Conrod, D., Naldrett, A.J. and Evensen, N.M. 1986. Petrologic, chemical, isotopic, and economic-potential studies of the Nipissing diabase; *in* Geoscience Research Grant Program, Summary of Research 1985-1986, Ontario Geological Survey, Miscellaneous Paper 130, p.87-106.

——— 1987. Petrologic, chemical, isotopic, and economic-potential studies of the Nipissing diabase; *in* Geoscience Research Grant Program, Summary of Research 1986-1987, Ontario Geological Survey, Miscellaneous Paper 136, p.4-26.

Lightfoot, P.C., De Souza, H. and Doherty, W. 1993. Differentiation and source of the Nipissing Diabase intrusions, Ontario, Canada; Canadian Journal of Earth Sciences, v.30, p.1123-1140.

Lightfoot, P.C., Doherty, W., Farrell, K.P., Keays, R.R., Moore, M.L. and Pekeski, D. 1997. Geochemistry of the Main Mass, Sublayer, Offsets and inclusions from the Sudbury Igneous Complex; Ontario Geological Survey, Open File Report 5959, 231p.

Lightfoot, P.C. and Farrow, C.E.G. 2002. Geology, geochemistry and mineralogy of the Worthington Offset Dike: a genetic model for offset dike mineralization in the Sudbury Igneous Complex; Economic Geology, v.97, p.1419-1446.

Lightfoot, P.C., Keays, R.R. and Doherty, W. 2001. Chemical evolution and origin of nickel sulphide mineralization in the Sudbury Igneous Complex, Ontario, Canada; Economic Geology, v.96, p.1855-1875.

Lightfoot, P.C., Keays, R.R., Morrison, G.G., Bite, A. and Farrell, K.P. 1997a. Geochemical relationships in the Sudbury Igneous Complex: origin of the Main Mass and Offset dikes; Economic Geology, v.92, p.289-307.

——— 1997b. Geological and geochemical relationships between the Contact Sublayer, inclusions, and Main Mass of the Sudbury Igneous Complex: A case study of the Whistle Mine Embayment; Economic Geology, v.92, p.647-673.

Lightfoot, P.C. and Naldrett, A.J. 1989. Assimilation and fractionation in basic magma chambers: significance of trace-element and Nd-isotopic data from the Kerns sill, Nipissing diabase, Ontario; Canadian Journal of Earth Sciences, v.26, p.737-754.

——— 1996. Petrology and geochemistry of the Nipissing Gabbro: exploration strategies for nickel, copper, and platinum group elements in a large igneous province; Ontario Geological Survey, Study 58, 80p.

Lloyd, T.R. 1984. Stratigraphy and sedimentology of basal Huronian (Lower Aphebian) strata near Sudbury, Ontario; unpublished BSc thesis, Laurentian University, Sudbury, Ontario, 70p.

Lochhead, D.R. 1955. The Falconbridge ore deposit, Canada; Economic Geology, v.50, p.42-50.

Long, D.G.F. 1974. Glacial and paraglacial genesis of conglomeratic rocks of the Chibougamau Formation (Aphebian), Chibougamau, Quebec; Canadian Journal of Earth Sciences, v.11, p.1236-1252.

——— 1976. The stratigraphy and sedimentology of the Huronian (Lower Aphebian) Mississagi and Serpent Formations; unpublished PhD thesis, University of Western Ontario, London, Ontario, 291p.

——— 1978. Depositional environments of a thick Proterozoic sandstone, the (Huronian) Mississagi Formation of Ontario, Canada; Canadian Journal of Earth Sciences, v.15, p.190-206.

——— 1987. Sedimentary framework of uranium deposits in the southern Cobalt Embayment, Ontario, Canada; *in* Uranium Deposits in Proterozoic Quartz-Pebble Conglomerates, AEA-TECDOC-27, International Atomic Energy Agency, Vienna, Austria, p.155-188.

——— 1995. Huronian sandstone thickness and paleocurrent trends as a clue to the tectonic evolution of the Southern Province; Canadian Mineralogist, v.33, part 4, p.922-923.

——— 2004. The tectonostratigraphic evolution of the Huronian basement and the subsequent basin fill: geological constraints on impact models of the Sudbury event; Precambrian Research, v.129, p.203-223.

Long, D.G.F. and Leslie, C.A. 1986. The placer gold potential of the early Aphebian Gowganda Formation along the northern margin of the Cobalt Embayment, with comments on associated concentrations of silver and copper; Ontario Geological Survey, Open File Report 5608, 54p.

Long, D.G.F. and Lloyd, T.R. 1983. Placer gold potential of basal Huronian strata of the Elliot Lake Group in the Sudbury area, Ontario; *in* Summary of Field Work 1983, Ontario Geological Survey, Miscellaneous Paper 116, p.256-258.

Long, D.G.F., Young, G.M., Rainbird, R.H. and Fedo, C.M. 1999. Actualistic and non-actualistic Precambrian sedimentary styles: examples from the Proterozoic, north shore of Lake Huron; Field Trip B5, Geological Association of Canada–Mineralogical Association of Canada, Joint Annual Meeting, Sudbury, Ontario, Guidebook, 50p.

Lovell, H.L. and Caine, T.W. 1970. Lake Timiskaming Rift Valley; Ontario Department of Mines, Miscellaneous Paper 39, 16p.

Lowman, P.D., Jr. 1997. Original size and shape of the Sudbury Structure; *in* Conference on Large Meteorite Impacts and Planetary Evolution (Sudbury), Contribution, v.922, Lunar and Planetary Institute, Houston, p.31-32.

——— 2002. Exploring Space, Exploring Earth. New Understanding of the Earth from Space Research; Cambridge University Press, Cambridge, New York, Melbourne, Madrid, Cape Town, 362p.

Mamen, C. 1955. Nickel Rim Mines Limited. A property evaluation; Canadian Mining Journal, v.76, no.6, p.56-61.

Marsh, B.D. 2006. Dynamics of magma systems; Elements, v.2, p.287-292.

Martin, W.C. 1957. Errington and Vermilion Lake mines; *in* Structural Geology of Canadian Ore Deposits, Canadian Institute of Mining and Metallurgy, Congress Volume, p.363-376.

Masaitis, V.L., Shafranovsky, G.I., Grieve, R.A.F., Langenhorst, F., Peredery, W.V., Therriault, A.M., Balmasov, E.L. and Fedorova, I.G. 1999. Impact diamonds in the suevitic breccias of the Black member of the Onaping Formation, Sudbury Structure, Ontario, Canada; *in* Large Meteorite Impacts and Planetary Evolution II, Geological Society of America, Special Paper 339, p.317-321.

McBirney, A.R. 1985. Further consideration of double-diffusive stratification and layering in the Skaergaard intrusion; Journal of Petrology, v.26, p.993-1001.

——— 1987. Constitutional zone refining of layered intrusions; *in* Origin of Igneous Layering, NATO ASI Series C, 196, p.437-452.

McBirney, A.R. and Noyes, R.M. 1979. Crystallization and layering of the Skaergaard intrusion; Journal of Petrology, v.20, p.487-554.

McCormick, K.A., Lesher, C.M., McDonald, A.M., Fedorowich, J.S. and James, R.S. 2002. Chlorine and alkali geochemical haloes in the Footwall Breccia and Sublayer Norite at the margin of the Strathcona Embayment, Sudbury Structure, Canada; Economic Geology, v.97, p.1509-1520.

McGrath, P.H. and Broome, H.J. 1994. A gravity model for the Sudbury Structure along the lithoprobe seismic line; Geophysical Research Letters, v.21, no.10, p.955-958.

Meldrum, A., Abel-Rahman, A.M., Martin, R.F. and Wodicka, N. 1997. The nature, age and petrogenesis of the Cartier Batholith, northern flank of the Sudbury Structure, Ontario; Precambrian Research, v.82, p.265-285.

Melosh, H.L. 1989. Impact Cratering. A Geological Process; Oxford University Press, 245p.

Meyer, W., Campbell, R.W., Adlington, R. and Toews, F.H. 1986. Sudbury Resident Geologist's area, Northeastern Region; *in* Report of Activities 1985, Regional and Resident Geologists, Ontario Geological Survey, Miscellaneous Paper 128, p.256-272.

Meyer, W., Campbell, R.W. and Toews, F.H. 1987. Sudbury Resident Geologist's area, Northeastern Region; *in* Report of Activities 1986, Regional and Resident Geologists, Ontario Geological Survey, Miscellaneous Paper 134, p.256-273.

Meyn, H.D. 1970. Geology of Hutton and Parkin townships; Ontario Department of Mines, Geological Report 80, 78p.

Milkereit, B., Green, A. and Sudbury Working Group 1992. Deep geometry of the Sudbury Structure from seismic reflection profiling; Geology, v.20, p.807-811.

Mitchell, G.P. and Mutch, A.D. 1957. Geology of the Hardy mine; *in* Structural Geology of Canadian Ore Deposits, Canadian Institute of Mining and Metallurgy, Congress Volume, p.350-363.

Miyashiro, A. 1961. Evolution of metamorphic belts; Journal of Petrology, v.2, p.277-311.

Molinar, F., Watkinson, D.H. and Jones, P.C. 2001. Multiple hydrothermal processes in footwall units of the North Range, Sudbury Igneous Complex, Canada, and implications for the genesis of vein-type Cu-Ni-PGE deposits; Economic Geology, v.96, p.1645-1670.

Moncrieff, A.C.M. 1989. Classification of poorly sorted sedimentary rocks; Sedimentary Geology, v.65, p.95-144.

Moore, M., Lightfoot, P.C. and Keays, R.R. 1993. Geology and geochemistry of footwall ultramafic rocks, Sudbury Igneous Complex, Fraser Mine, Sudbury, Ontario; *in* Summary of Field Work and Other Activities 1993, Ontario Geological Survey, Miscellaneous Paper 162, p.85-86.

Morgan, J.W., Walker, R.J., Horan, M.F., Beary, E.S. and Naldrett, A.J. 2002. ^{190}Pt-^{186}Os and ^{187}Re-^{187}Os systematics of the Sudbury Igneous Complex, Ontario; Geochimica et Cosmochimica Acta, v.66, p.273-290.

Morrison, G.G. 1984. Morphological features of the Sudbury Structure in relation to an impact origin; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.513-520.

——— 1999. Magmatic sulphide ore deposits of the Sudbury Structure: an overview; Geological Association of Canada–Mineralogical Association of Canada, Abstracts Volume 24, p.86-87.

Morrison, G.G., Jago, B.C. and White, T.L. 1994. Footwall mineralization of the Sudbury Igneous Complex; *in* Proceedings of the Sudbury – Noril'sk Symposium, Ontario Geological Survey, Special Volume 5, p.57-64.

Mourre, G.A. 2000. Geological relationships at discontinuities in the Copper Cliff quartz diorite offset: an investigation into offset dikes and their relationship to the Sudbury Igneous Complex, Ontario; unpublished MSc thesis, Laurentian University, Sudbury, Ontario, 255p.

Muir, T.L. 1981. Geology of the Capreol Lake area, District of Sudbury; Ontario Geological Survey, Open File Report 5344, 168p.

——— 1983. Geology of the Morgan Lake–Nelson Lake area, District of Sudbury; Ontario Geological Survey, Open File Report 5426, 203p.

——— 1984. The Sudbury Structure: considerations and models for an endogenic origin; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.449-489.

Muir, T.L. and Peredery, W.V. 1984. The Onaping Formation; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.139-210.

Muir, T.L. and Taylor, A.B. 1981. Precambrian geology of the Morgan Lake–Nelson Lake area, Sudbury District; Ontario Geological Survey, Preliminary Map P.2419, scale 1:15 840.

Müller-Mohr, V. 1992a. Breccias in the basement of a deeply eroded impact structure, Sudbury, Canada; Tectonophysics, v.216, p.219-226.

——— 1992b. Gang breccien der Sudbury-Struktur: Geologie, Petrographie und Geochemie der Sudbury Breccie, Ontario, Kanada; unpublished PhD thesis, University of Munster, Nordhein Westfalen, Germany.

Mungall, J.E., Ames, D.E. and Hanley, J.J. 2004. Geochemical evidence from the Sudbury Structure for crustal redistribution by large bolide impacts; Nature, v.429, p.546-548.

Mungall, J.E. and Brenan, J.M. 2003. Experimental evidence for the chalcophile behaviour of the halogens; Canadian Mineralogist, v.41, p.207-220.

Murphy, A.J. and Spray, J.S. 2002. Geology, mineralization and emplacement of the Whistle–Parkin Offset dike, Sudbury; Economic Geology, v.97, 1399-1418.

Murray, A. 1857. Report for the year 1856; *in* Geological Survey of Canada, Report of Activities, 1853-1856, p.180-181.

Naldrett, A.J. 1969. A portion of the system Fe-S-O and its application to sulfide ore magmas; Journal of Petrology, v.10, p.171-201.

——— 1984a. Introduction to the Sudbury Igneous Complex; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.234.

——— 1984b. Ni-Cu ores of the Sudbury Igneous Complex; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.302-307

——— 1984c. Mineralogy and composition of the Sudbury ores; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.309- 325

——— 1999. Summary of developments of ideas on Sudbury geology 1992-1998; *in* Large Meteorite Impacts and Planetary Evolution II, Geological Society of America, Special Paper 339, p.431-442.

——— 2004. Magmatic Sulfide Deposits: Geology, Geochemistry and Exploration; Springer-Verlag, Berlin, Heidelberg, New York, 727p.

Naldrett, A.J., Asif, M., Schandl, E., Searcy, T., Morrison, G., Binney, P. and Moore, C. 1999. PGE in the Sudbury ores: Significance with respect to the origin of different ore zones and the exploration for footwall ore bodies; Economic Geology, v.94, p.185-210.

Naldrett, A.J., Bray, J.G., Gasparrini, E.L., Podolsky, T. and Rucklidge, J.C. 1970. Cryptic variations and the petrology of the Sudbury Nickel Irruptive; Economic Geology, v.65, p.122-155.

Naldrett, A.J., Greenman, L. and Hewins, R.H. 1972. The Main Irruptive and Sublayer at Sudbury, Ontario; *in* 24th International Geological Congress, Proceedings, Section 4, p.206-214.

Naldrett, A.J. and Hewins, R.H. 1984. The Main Mass of the Sudbury Igneous Complex; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.235-251.

Naldrett, A.J., Hewins, R.H., Dressler, B.O. and Rao, B.V. 1984. The Contact Sublayer of the Sudbury Igneous Complex; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.253-274.

Naldrett, A.J., Innes, D.G., Sowa, J. and Gorton, M. 1982. Compositional variation within and between five Sudbury ore deposits; Economic Geology, v.77, p.1519-1534.

Naldrett, A.J. and Kullerud, G. 1967. A study of the Strathcona Mine and its bearing on the origin of the nickel-copper ores of the Sudbury district, Ontario; Journal of Petrology, v.8, p.453-531.

Naldrett, A.J., Pessaran, A., Asif, M. and Li, C. 1994. Compositional variation in the Sudbury ores and prediction of the proximity of footwall copper-PGE ore deposits; *in* Proceedings of the Sudbury – Noril'sk Symposium, Ontario Geological Survey, Special Volume 5, p.133-146.

Naldrett, A.J., Rao, B.V. and Evensen, N.M. 1986. Contamination at Sudbury and its role in ore formation; *in* Proceedings of the Conference 'Metallogeny of Basic and Ultrabasic Rocks', The Institution of Mining and Metallurgy, p.75-91.

Noble, S.R. and Lightfoot, P.C. 1992. U-Pb baddeleyite ages for the Kerns and Triangle Mountain intrusions, Nipissing diabase, Ontario; Canadian Journal of Earth Sciences, v.29, p.1124-1129.

Norman, M.D. 1994. Sudbury Igneous Complex: impact melt or endogenous magma? Implications for lunar crustal evolution; Geological Society of America, Special Paper 187, p.331-341.

Ostermann, M., Scharer, U. and Deutsch, A. 1996. Impact melt dikes in the Sudbury multi-ring basin (Canada): implications from uranium-lead geochronology of the Foy Offset Dike; Meteoritics & Planetary Science, v.31, p.494-501.

Owen, D.L. and Coats, C.J.L. 1984. Falconbridge and East mines; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.371-378.

Paakki, J.J. 1992. The Errington Zn-Cu-Pb massive sulphide deposit, Sudbury, Ontario: its structural and stratigraphic setting and footwall alteration; unpublished MSc thesis, Laurentian University, Sudbury, Ontario, 140p.

Park, R.G. 1997. Foundations of Structural Geology, 3rd ed.; Chapman & Hall, London, 202p.

Pattison, E.F. 1979. The Sudbury Sublayer; Canadian Mineralogist, v.17, p.257-274.

Peck, D.C., James, R.S. and Chubb, P.T. 1993. Geological environments for PGE-Cu-Ni mineralization in the East Bull Lake gabbro-anorthosite intrusion, Ontario; Exploration and Mining Geology, v.2, p.85-104.

Peredery, W.V. 1972. Chemistry of fluidal glasses and melt bodies in the Onaping Formation; *in* New Developments in Sudbury Geology, Geological Association of Canada, Special Paper Number 10, p.49-59.

Peredery, W.V. and Morrison, G.G. 1984. Discussion of the origin of the Sudbury Structure; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.491-511.

Peredery, W.V and Naldrett, A.J. 1975. Petrology of the upper irruptive rocks, Sudbury, Ontario; Economic Geology, v.70, p.164-175.

Phemister, T.C. 1926. Igneous rocks of Sudbury and their relation to the ore deposits; *in* Ontario Department of Mines, Annual Report for 1925, Volume 34, Part 8, p.1-61.

——— 1937. A review of the problems of the Sudbury irruptive; Journal of Geology, v.45, p.1-47.

Piercey, P., Schneider, D.A. and Holm, D.K. 2007. Geochronology of Proterozoic metamorphism in the deformed Southern Province, northern Lake Huron region, Canada; Precambrian Research, v.157, p.27-143.

Prevec, S.A. 1993. An isotopic, geochemical and petrographic investigation of the genesis of early Proterozoic mafic intrusions and associated volcanism near Sudbury, Ontario; unpublished PhD thesis, University of Alberta, Edmonton, Alberta, 223p.

Prevec, S.A. and Cawthorn, R.G. 2002. Thermal evolution and interaction between impact melt sheet and footwall: a genetic model for the contact sublayer of the Sudbury Igneous Complex, Canada; Journal of Geophysical Research, v.107, p.14.

Pye, E.G., Naldrett, A.J. and Giblin, P.G. 1984. The Geology and Ore Deposits of the Sudbury Structure; Ontario Geological Survey, Special Volume 1, 603p.

Ramsay, J.G. 1967. Folding and Fracturing of Rocks; McGraw-Hill, New York, 568p.

Reimold, W.U. 1995. Pseudotachylite in impact structures - generation by frictional melt and shock brecciation?: A review and discussion; Earth Science Reviews, v.39, p.247-265.

Reynolds, D.L. 1954. Fluidization as a geological process, and its bearing on the problem of intrusive granites; American Journal of Science, v.252, p.577-614.

Riller, U. 2005. Structural characteristics of the Sudbury Impact Structure, Canada: impact induced and orogenic deformation; Meteoritics & Planetary Science, v.40, p.1723-1740.

Riller, U., Cruden, A.R. and Schwerdtner, W.M. 1996. Magnetic fabric and microstructural evidence for a tectono-thermal overprint of the Murray pluton, central Ontario, Canada; Journal of Geology, v.18, p.1005-1016.

Riller, U. and Schwerdtner, W.M. 1997. Mid-crustal deformation of the southern flank of the Sudbury Basin, central Ontario, Canada; Geological Society of America Bulletin, v.109, p.841-854.

Riller, U., Schwerdtner, W.M., Halls, H.C. and Card, K.D. 1999. Transpression tectonism in the eastern Penokean orogen, Canada: consequences for Proterozoic crustal kinematics and continental fragmentation; Precambrian Research, v.93, p.51-70.

Robertson, P.B. and Grieve, R.A.F. 1977. Shock attenuation at terrestrial impact sites; *in* Impact and Explosion Cratering, Proceedings of the Symposium on Planetary Cratering Mechanisms, Pergamon Press, New York, N.Y., p.687-702

Roscoe, S.M. 1969. Huronian rocks and uraniferous conglomerates in the Canadian Shield; Geological Survey of Canada, Paper 68-40, 205p.

——— 1996. Paleoplacer uranium, gold; *in* Geology of Canadian Mineral Deposit Types, Geological Survey of Canada, Geology of Canada, v.8, p.10-23.

Roscoe, S.M. and Card, K.D. 1992. Early Proterozoic tectonics and metallogeny of the Lake Huron region of the Canadian Shield; Precambrian Research, v.48, p.99-119.

——— 1993. The reappearance of the Huronian in Wyoming: rifting and drifting of ancient mountains; Canadian Journal of Earth Sciences, v.30, p.2475-2480.

Rosenberg, C.L. and Riller, U. 2000. Partial-melt topology in statically and dynamically recrystallized granite; Geology, v.29, p.7-10.

Rousell, D.H. 1972. The Chelmsford Formation of the Sudbury Basin - a Precambrian turbidite; *in* New Developments in Sudbury Geology, Geological Association of Canada, Special Paper Number 10, p.79-91.

——— 1975. The origin of foliation and lineation in the Onaping Formation and the deformation of the Sudbury Basin; Canadian Journal of Earth Sciences, v.12, p.1379-1395.

——— 1980. Kink bands in the Onaping Formation, Sudbury Basin, Ontario; Tectonophysics, v.66, p.83-97.

——— 1981. Sudbury and the meteorite theory; Geoscience Canada, v.8, p.167-169.

——— 1984a. Structural geology of the Sudbury Basin; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.83-95.

——— 1984b. Onwatin and Chelmsford formations; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.211-218.

——— 1984c. Mineralization in the Whitewater Group; *in* The Geology and Ore Deposits of the Sudbury Structure, Ontario Geological Survey, Special Volume 1, p.219-232.

——— 1984d. Nature and origin of mineral occurrences inside the Sudbury Basin; Canadian Institute of Mining and Metallurgy Bulletin, December, p.63-75.

——— 2006. Unsolved problems and the evolution of Sudbury geology; *in* 52nd Annual Meeting, Institute on Lake Superior Geology, Part 1 - Proceedings and Abstracts, p.57-58.

Rousell, D.H. and Everitt, R.A. 1981. Jointing in the Sudbury Basin, Ontario; *in* Proceedings of the Third International Conference on Basement Tectonics, Basement Tectonics Committee Incorporated, Denver, Colorado, p.381-391.

Rousell, D.H., Fedorowich, J.S. and Dressler, B.O. 2003. Sudbury Breccia (Canada): a product of the 1850 Ma Sudbury Event and host to footwall Cu-Ni-PGE deposits; Earth Science Reviews, v.60, p.147-174.

Rousell, D.H., Gibson, H.L. and Jonasson, I.R. 1997. The tectonic, magmatic and mineralization history of the Sudbury Structure; Exploration and Mining Geology, v.6, p.1-22.

Rousell, D.H., Jansons, K.J. and Richards, P.A. 1998. Urban geology of the City of Sudbury; *in* Urban Geology of Canadian Cities, Geological Association of Canada, Special Paper 42, p.207-224.

Rousell, D.H. and Long, D.G.F. 1998. Are outliers of the Huronian Supergroup preserved in structures associated with the collapse of the Sudbury impact crater?; The Journal of Geology, v.106, p.407-419.

Rousell, D.H., Meyer, W. and Prevec, S.A. 2002. Bedrock geology and mineral deposits; *in* The Physical Environment of the City of Greater Sudbury, Ontario Geological Survey, Special Volume 6, p.21-55.

Rowell, W.F. and Edgar, A.D. 1986. Platinum-group element mineralization in a hydrothermal Cu-Ni sulphide occurrence, Rathbun Lake, northeastern Ontario; Economic Geology, v.81, p.1272-1277.

Sadler, J.F. 1958. A detailed study of the Onwatin Formation; unpublished MSc thesis, Queen's University, Kingston, Ontario, 184p.

Sangster, D.F. 1970. Metallogenesis of some Canadian lead-zinc deposits in carbonate rocks; Proceedings of the Geological Association of Canada, v.22, p.27-36.

Schandl, E.S., Gorton, M.P. and Davis, D.W. 1994. Albitization at 1700 ±2 Ma in the Sudbury–Wanapitei Lake area, Ontario: implications for deep-seated alkalic magmatism in the Southern Province; Canadian Journal of Earth Sciences, v.32, p.597-601.

Schandl, E.S., Gorton, M.P., Davis, D.W., Wasteneys, H.A. and Martin, R.F. 1992. Paragenesis and geochronology of sodium metasomatism associated with gold mineralization in the Espanola–Sudbury–Wanapitei Lake area; *in* Geoscience Research Grant Program, Summary of Research 1991-1992, Ontario Geological Survey, Miscellaneous Paper 159, p.63-79.

Scott, R.G. and Benn, K. 2001. Peak-ring collapse accommodated by impact melt-filled transfer faults, Sudbury impact structure, Canada; Geology, v.29, p.747-750.

Scott, R.G. and Spray, J.G. 1999. Magnetic fabric constraints on friction melt flow regimes and ore emplacement direction within the South Range Breccia Belt, Sudbury Impact Structure; Tectonophysics, v.307, p.163-189.

——— 2000. The South Range Breccia Belt of the Sudbury Impact Structure: A possible terrace collapse feature; Meteoritics & Planetary Science, v.35, p.505-520.

Scribbins, B.T., Rae, D.R. and Naldrett, A.J. 1984. Mafic and ultramafic inclusions in the Sublayer of the Sudbury Igneous Complex; Canadian Mineralogist, v.22, p.67-75.

Severin, P.W.A. and Gates, B.I. 1981. Giant Yellowknife Mines Limited Ore Reserve Calculation Errington/Vermilion Deposits; Falconbridge Limited, Internal Report, 17p.

Shand, S.J. 1916. The pseudotachylyte of Parijs (Orange Free State) and its relation to “trap-shotten gneiss” and “flinty crush-rock”; Quarterly Journal of the Geological Society of London, v.72, p.198-221.

Shanks, W.S. and Schwerdtner, W.M. 1991. Structural analysis of the central and southwestern Sudbury Structure, Southern Province, Canadian Shield; Canadian Journal of Earth Sciences, v.28, p.411-430.

Shaw, C.S.J., Young, G.M. and Fedo, C.M. 1999. Sudbury-type breccias in the Huronian Gowganda Formation near Whitefish Falls, Ontario: products of diabase intrusion into completely consolidated sediments?; Canadian Journal of Earth Sciences, v.36, p.1435-1448.

Simony, P.S. 1964. Northwestern Temagami area; Ontario Department of Mines, Geological Report 28, 30p.

Sims, P.K. and 15 coauthors. 1993. The Lake Superior region and the Trans-Hudson Orogen; *in* Precambrian Conterminous, U.S. Geological Society of America, v.C2, p.11-120.

Sims, P.K., Card, K.D., Morey, G.B. and Peterman, Z.E. 1980. The Great Lakes tectonic zone - a major crustal structure in central North America; Geological Society of America Bulletin, v.91, p.690-698.

Slaught, W.H. 1951. A petrographic study of the Copper Cliff Offset in the Sudbury District; unpublished MSc thesis, McGill University, Montreal, Quebec, 68p.

Smith, M.D., Heaman, L.M. and Beakhouse, G.P. 1999. Constraints on the timing of felsic magmatism associated with Matachewan igneous events: Preliminary U-Pb results for the Creighton granite, Ontario; Geological Association of Canada–Mineralogical Association of Canada, Abstract Volume 24, p.119.

Souch, B.E., Podolsky, T. and Geological Staff, The International Nickel Company of Canada Limited 1969. The suphide ores at Sudbury: their particular relationship to a distinctive inclusion-bearing facies of the Nickel Irruptive; *in* Magmatic Ore Deposits, Economic Geology, Monograph 4, p.252-261.

Spear, F.S. 1993. Metamorphic phase equilibria and pressure-temperature-time paths; Mineralogical Society of America, Monograph 1, 799p.

Speers, E.C. 1957. The age relations and origin of common Sudbury Breccia; Journal of Geology, v.65, p.497-514.

Spray, J.G. 1997. Superfaults; Geology, v.25, p.579-582.

Spray, J.G. and Thompson, L.M. 1995. Friction melt distribution in a multi-impact basin; Nature, v.373, p.130-132.

Stevenson, J.S. 1961. Recognition of the Quartzite Breccia in the Whitewater Series, Sudbury Basin, Ontario; *in* Transactions of The Royal Society of Canada, v.55, Series 3, Section 4, p.57-66.

——— 1963. The upper contact phase of the Sudbury micropegmatite; The Canadian Mineralogist, v.7, pt.3, p.413-419.

——— 1972. The Onaping ash-flow sheet, Sudbury, Ontario; *in* New Developments in Sudbury Geology, Geological Association of Canada, Special Paper Number 10, p.41-48.

——— 1990. The volcanic origin of the Onaping Formation, Sudbury, Canada; Tectonophysics, v.171, p.249-257.

Stockwell, C.H. 1982. Proposal for the time classification and correlation of Precambrian rocks and events in Canada and adjacent areas of the Canadian Shield; Geological Survey of Canada, Paper 80-19, 135p.

Stoness, J.A. 1994. The stratigraphy, geochemistry and depositional environment of the Paleoproterozoic Vermilion and Onwatin formations, and their relationship to the Zn-Cu-Pb massive sulphide deposits in the Sudbury Basin; unpublished MSc thesis, Laurentian University, Sudbury, Ontario, 205p.

Sun, S.S. and McDonough, W.F. 1989. Chemical and isotopic systematics of ocean basalts: Implications of mantle composition and processes; *in* Magmatism of the Ocean Basins, Geological Society of London, v.42, p.313-345.

Therriault, A.M., Fowler, A.D. and Grieve, R.A.F. 2002. Sudbury Igneous Complex: a differentiated impact melt sheet; Economic Geology, v.97, p.1521-1540.

Thompson, L.M. 1996. A study of pseudotachylyte associated with Sudbury Structure, Ontario; unpublished PhD thesis, University of New Brunswick, Fredericton, New Brunswick, 202p.

Thompson, L.M. and Spray, J.G. 1994. Pseudotachylytic rock distribution and genesis within the Sudbury impact structure; *in* Large Meteorite Impacts and Planetary Evolution, Geological Society of America, Special Paper 293, p.275-287.

——— 1996. Pseudotachylyte petrogenesis: constraints from the Sudbury impact structure; Contributions to Mineralogy and Petrology, v.125, p.359-374.

Thomson, J.E. 1957. Geology of the Sudbury Basin; *in* 65th Annual Report of the Ontario Department of Mines, v.65, pt.3, p.1-56.

——— 1969. A discussion of Sudbury geology and sulphide deposits; Ontario Department of Mines, Miscellaneous Report 30, 22p.

Thomson, J.E. and Williams, H. 1959. The myth of the Sudbury Lopolith; Canadian Mining Journal, v.80, p.57-62.

Tolman, C.F., Jr. and Rogers, A.F. 1916. A study of the magmatic sulfide ores; Leland Stanford Junior University Publications, University Series 1916, Engineering and Mining Journal, v.103, p.226-229.

Tomlinson, K.Y. 1996. The geochemistry and tectonic setting of Early Precambrian greenstone belts; unpublished PhD thesis, University of Portsmouth, UK, 287p.

Tomlinson, K.Y., Bennett, G., Keays, R.R. and Thurston, P.C. 1999. The Thessalon volcanic formation: multiple pulses of magmatism in a continental rift environment; *in* Abstract Volume 24, Joint Annual Meeting, Geological Association of Canada–Mineralogical Association of Canada, p.130.

Tuchscherer, M.G. and Spray, J.G. 2002. Geology, mineralization and emplacement of the Foy Offset Dike, Sudbury impact structure; Economic Geology, v.97, p.1377-1397.

Vogel, D.C., James, R.S. and Keays, R.R. 1998. The early tectono-magmatic evolution of the Southern Province: implications from the Agnew intrusion, central Ontario, Canada; Canadian Journal of Earth Sciences, v.35, p.854-8790.

Vogel, D.C., Vuollo, J.I., Alapieti, T.T. and James, R.S. 1998. Tectonic, stratigraphic and geochemical comparisons between ca. 2500-2440 Ma mafic igneous events in the Canadian and Fennoscandian shields; Precambrian Research, v.92, p.89-116.

Walker, R.J., Morgan, J.W., Naldrett, A.J., Li, C. and Fassett, J.D. 1991. Re-Os isotopic systematics of Ni-Cu sulfide ores, Sudbury Igneous Complex, Ontario: Evidence for a major crustal component; Earth and Planetary Science Letters, v.105, p.416-429.

Wandke, A. and Hoffman, R. 1924. A study of the Sudbury ore deposits; Economic Geology, v.19, p.169-204.

Watkinson, D.H. 1999. Platium-group element enrichment in Cu-Ni-rich sulfides from footwall deposits, Sudbury Igneous Complex, Canada; Chronique de la Recherche Miniere, No.535, p.29-43.

Williams, H.G. 1957. Glowing avalanche deposits of the Sudbury Basin; *in* 65th Annual Report of the Ontario Department of Mines, v.65, pt.3, p.57-89.

Williams, H.R., Stott, G.M., Thurston, P.C., Sutcliffe, R.H., Bennett, R.M., Easton, R.M. and Fyon, J.A. 1992. Tectonic evolution of Ontario; *in* Geology of Ontario, Ontario Geological Survey, Special Volume 4, Part 2, p.1255-1332.

Wilson, H.D.B. 1956. Structure of lopoliths; Bulletin of the Geological Society of America, v.67, p.289-300.

Wood, C.R. and Spray, J.G. 1998. Origin and emplacement of offset dikes in the Sudbury impact structure: constraints from Hess; Meteoritics & Planetary Science, v.33, p.337-347.

Wu, J., Milkereit, B. and Boerner, D. 1994. Time constraints on deformational history of the Sudbury Impact Structure; Canadian Journal of Earth Sciences, v.31, p.1654-1660.

Wykes, J.L. and Mavrogenes, J.A. 2005. Hydrous sulfide melting: Experimental evidence for the solubility of H_2O in sulfide melts; Economic Geology, v.100, p.157-164.

Yates, A.B. 1938. The Sudbury Intrusive; Royal Society of Canada, Transactions, Section 4, p.151-172.

——— 1948. Properties of the International Nickel Company of Canada; *in* Structural Geology of Canadian Ore Deposits, Canadian Institute of Mining and Metallurgy, Jubilee Volume, p.596-617.

Young, G.M. 1970. An extensive early Proterozoic glaciation in North America?; Paleogeography, Paleoclimatology and Paleoecology, v.7, p.85-101.

——— 1973a. Tillites and aluminum quartzites as possible time markers for Middle Precambrian (Aphebian) rocks of North America; *in* Huronian Stratigraphy and Sedimentation, Geological Association of Canada, Special Paper Number 12, p.97-127.

——— 1973b. Origin of carbonate-rich early Proterozoic Espanola Formation, Ontario; Geological Society of America Bulletin, v.84, p.135-160.

——— 1981. Diamictites of the early Proterozoic Ramsay Lake and Bruce formations, north shore of Lake Huron, Ontario, Canada; *in* Earth's Pre-Pleistocene Glacial Record, Cambridge University Press, Cambridge, UK, p.813-816.

——— 1983. Tectono-sedimentary history of early Proterozoic rocks of the northern Great Lakes region; *in* Early Proterozoic Geology of the Great Lakes Region, Geological Society of America, Memoir 160, p.15-32.

——— 1995. The Huronian Supergroup in the context of a Paleoproterozoic Wilson Cycle in the Great Lakes region; The Canadian Mineralogist, v.33, part 4, p.917-944.

Young, G.M., Long, D.G.F., Fedo, C.M. and Nesbitt, H.W. 2001. The Paleoproterozoic Huronian Basin: product of a Wilson Cycle accompanied by glaciation and meteorite impact; Sedimentary Geology, v.141-142, p.233-254.

Young, G.M. and Nesbitt, H.W. 1985. The Gowganda Formation in the southern part of the Huronian outcrop belt, Ontario, Canada: stratigraphy, depositional environments and regional tectonic significance; Precambrian Research, v.29, p.265-301.

Zieg, M.J. and Marsh, B.D. 2005. The Sudbury Igneous Complex: Viscous emulsion differentiation of a superheated impact melt sheet; Geological Society of America Bulletin, v.117, p.1427-1450.

Zolnai, A.I., Price, R.A. and Helmstaedt, H. 1984. Regional cross section of the Southern Province adjacent to Lake Huron, Ontario: implications for the tectonic significance of the Murray fault zone; Canadian Journal of Earth Sciences, v.21, p.447-456.

Zurbrigg, H.F. 1957. Frood–Stobie Mine; *in* Structural Geology of Canadian Ore Deposits, Canadian Institute of Mining and Metallurgy, Jubilee Volume, p.341-349.

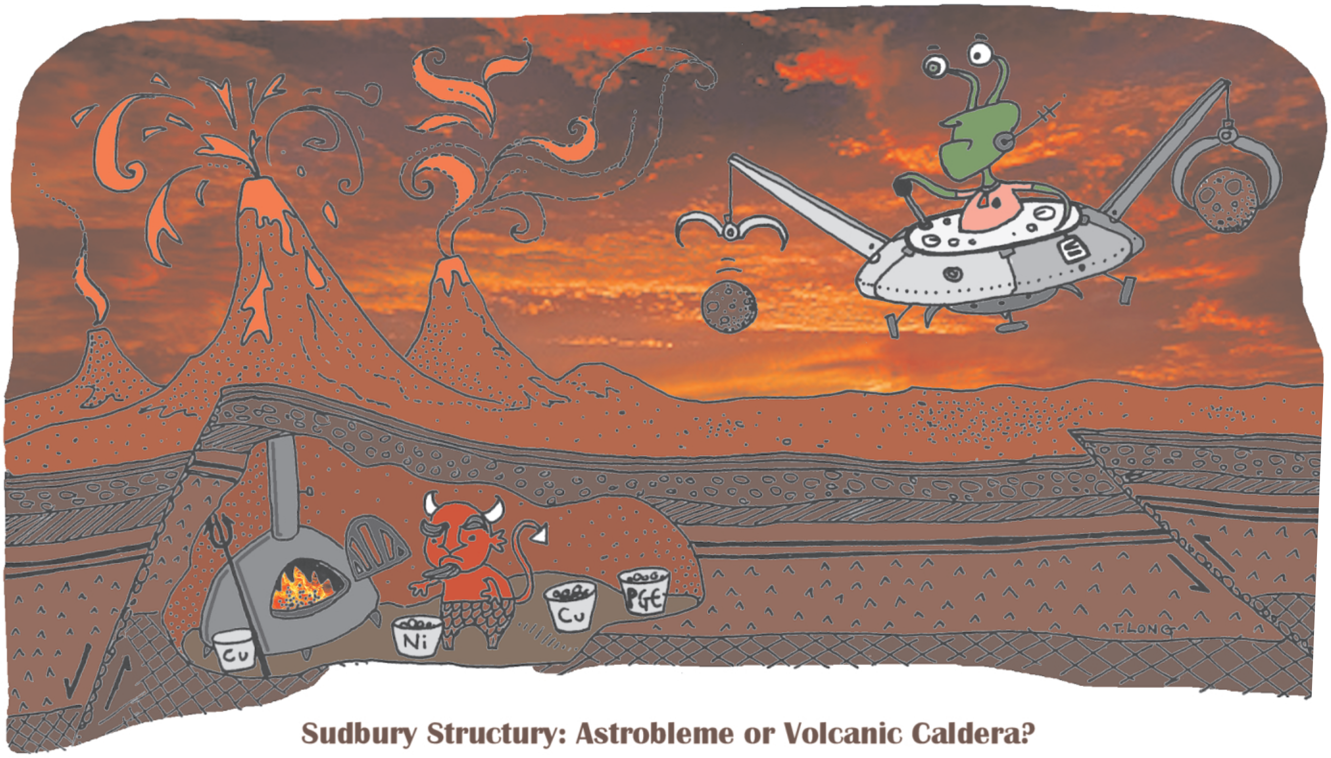
Metric Conversion Table

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

OTHER USEFUL CONVERSION FACTORS

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

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



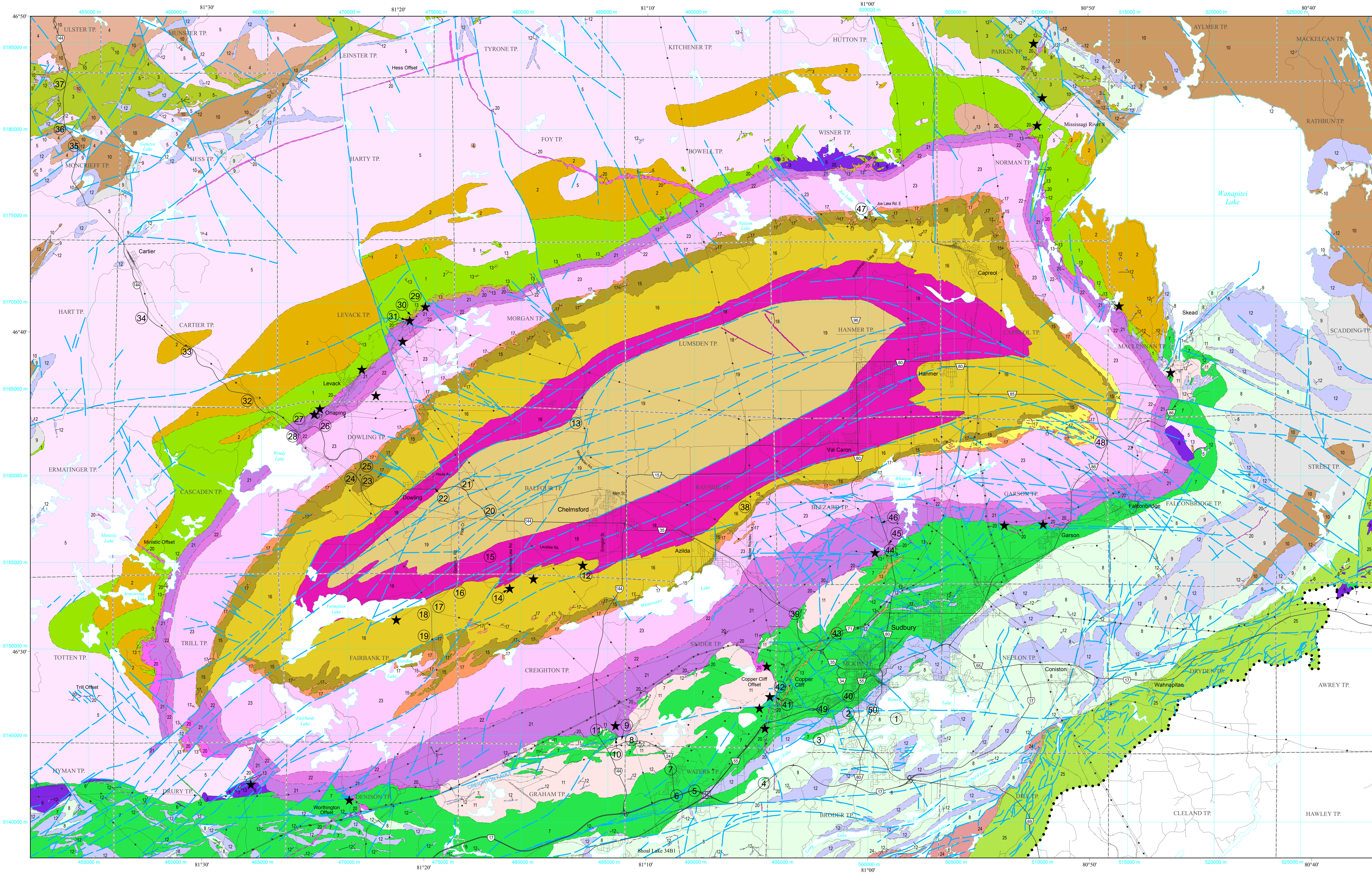
Sudbury Structury: Astrobleme or Volcanic Caldera?

By Tristan Long

LOCKERBY MINE

ISSN 0826-9580 [print]
ISBN 978-1-4435-1574-0 [print]

ISSN 1916-6117 [online]
ISBN 978-1-4435-1575-7 [PDF]

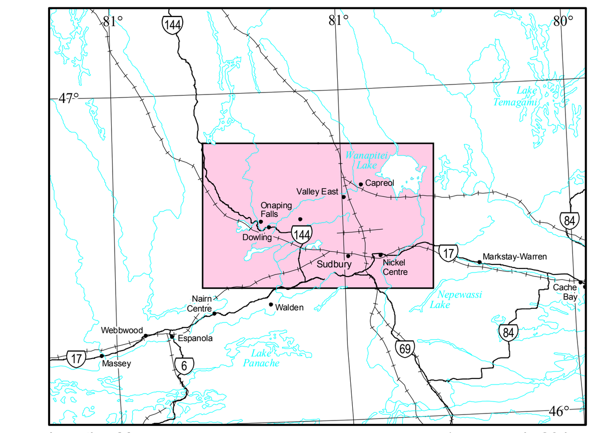


PRECAMBRIAN GEOLOGY OF THE SUDBURY AREA

Scale 1:100 000

2000 m 0 2 4 km

NTS References: 41 I/5,6,7,10,11,12,13,14,15



Location Map 1 cm equals 20 km

LEGEND

GRENVILLE PROVINCE

25	Gneisses and intrusive rocks

SOUTHERN PROVINCE

CHIEF LAKE IGNEOUS COMPLEX

24	Massive to foliated felsic intrusive rocks, minor gabbro

SUDBURY IGNEOUS COMPLEX (1850 Ma)

23	Upper Zone: granophyre, granite
22	Middle Zone: quartz gabbro
21	Lower Zone: felsic and mafic norite, gabbro
20	Contact and Offset Sublayer

WHITEWATER GROUP

19	Chelmsford Formation
18	Onwatin Formation
17	Syn-Onaping intrusive rocks

Onaping Formation

16	Dowling Member
15	Sandcherry Member
14	Garson Member

13	Footwall Breccia
12	Nipissing gabbro
11	Creighton, Murray and Skead felsic plutons

HURONIAN SUPERGROUP (< 2480 to > 2220 Ma)

10	Cobalt Group
9	Quirke Lake Group
8	Hough Lake Group
7	Elliot Lake Group

6	Mafic intrusive rocks

SUPERIOR PROVINCE

5	Cartier batholith
4	Metatexite and diatexite with >30% granitic leucosome; local monzonite
3	Felsic, intermediate and mafic volcanic rocks; minor metasedimentary rocks

LEVACK GNEISS COMPLEX (~ 2700 to 2640 Ma)

2	Massive to foliated felsic rocks; paragneiss, locally migmatite
1	Tonalite-granodiorite gneiss; mafic-ultramafic gneiss

SYMBOLS

Geologic boundary

Limit of study area

Faults

Road (major, secondary)

Mines: Present/Past

Railway

Powerline, pipeline

Township boundary

Field Trip stops (see text of report)

SOURCES OF INFORMATION

Base map information is derived from the Ontario Land Information Warehouse, Land Information Ontario, Ontario Ministry of Natural Resources, scale 1:20 000, with modifications by staff of the Ministry of Northern Development, Mines and Forestry.

Universal Transverse Mercator (UTM) co-ordinates are in North American Datum 1983 (NAD83), Zone 17.

Geology modified from: Ames, D.E., Davidson, A., Buckle, J.L. and Card, K.D. 2005. Geology, Sudbury bedrock compilation, Ontario; Geological Survey of Canada, Open File 4570, scale 1:50 000.

Magnetic declination approximately 10°15′W at the centre of the map area in 2009.

This map accompanies Open File Report 6243 “*A Field Guide to the Geology of Sudbury, Ontario*”.