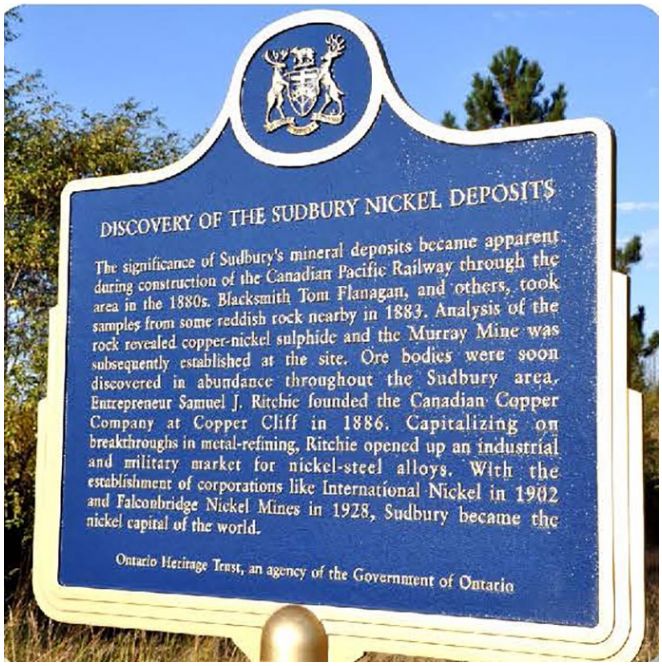
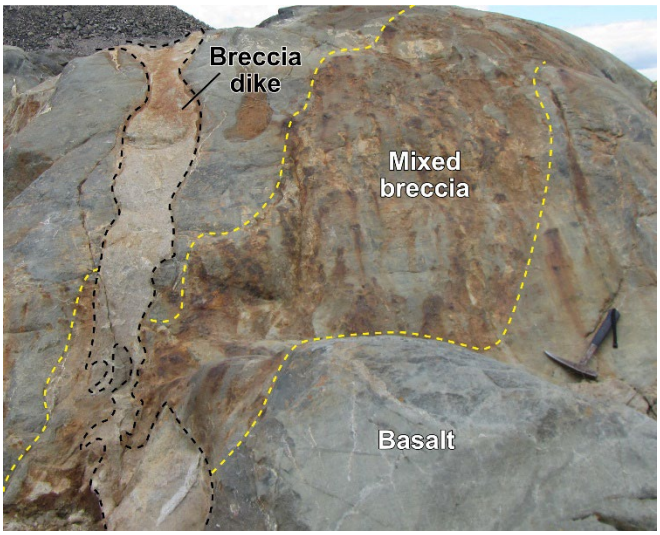
68th Annual Meeting

Sudbury, Ontario — May 10-11, 2022

Institute on Lake Superior Geology

Part 2 — Field Trip Guidebook











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68th ANNUAL MEETING

INSTITUTE ON LAKE SUPERIOR GEOLOGY

May 10-11, 2022
Sudbury, Ontario

HOSTED BY

Michael Easton and Wouter Bleeker

Co-Chairs

Ontario Geological Survey and Geological Survey of Canada

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Part 2 – Field Trip Guidebook

Compiled and edited by Michael Easton

Cover Photos. Upper Left — Signage at the Sudbury ore discovery site (Trips 1 and 4). Upper Right— Footwall Breccia in the Crean Hill Mine area (Trip 3). Lower Left — Arkosic sandstone overlain by matrix- to clast-supported conglomerate, Gowganda Formation. On Highway 108 north of Elliot Lake (Trip 5). Lower Right — Deformed, migmatitic gneiss in the Grenville Front tectonic zone showing garnet dispersed throughout the rock. Scale card is 9 cm long. On Highway 537 southeast of Sudbury (Trip 2).

68th Institute on Lake Superior Geology

Volume 68 consists of:

Part 1: Program and Abstracts

Part 2: Field Trip Guidebook

Trip 1: A Traverse Across the Sudbury Impact Structure

Trip 2: Geology of the Grenville Front and the Grenville Front Tectonic Zone in the Sudbury Area

Trip 3: Magmatism and Brecciation in the Footwall Rocks in the Southwestern Sudbury Structure

Trip 4: An Overview of the Geology of the Sudbury Structure

Trip 5: A Cross-Section through the Huronian Supergroup at Elliot Lake, Ontario

Reference to material in Part 2 should follow the example below:

Gordon, C., Généraux, C-A. and Clarke, B. 2022. Magmatism and Brecciation in the Footwall rocks of the southwestern Sudbury Structure; *in* Easton, R.M. (Ed.), Institute on Lake Superior Geology Proceedings, 68th Annual Meeting, Sudbury, Ontario, Part 2 – Field trip guidebook. v.68, part 2, p.147-180.

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Part 2: Field Trip Guidebook

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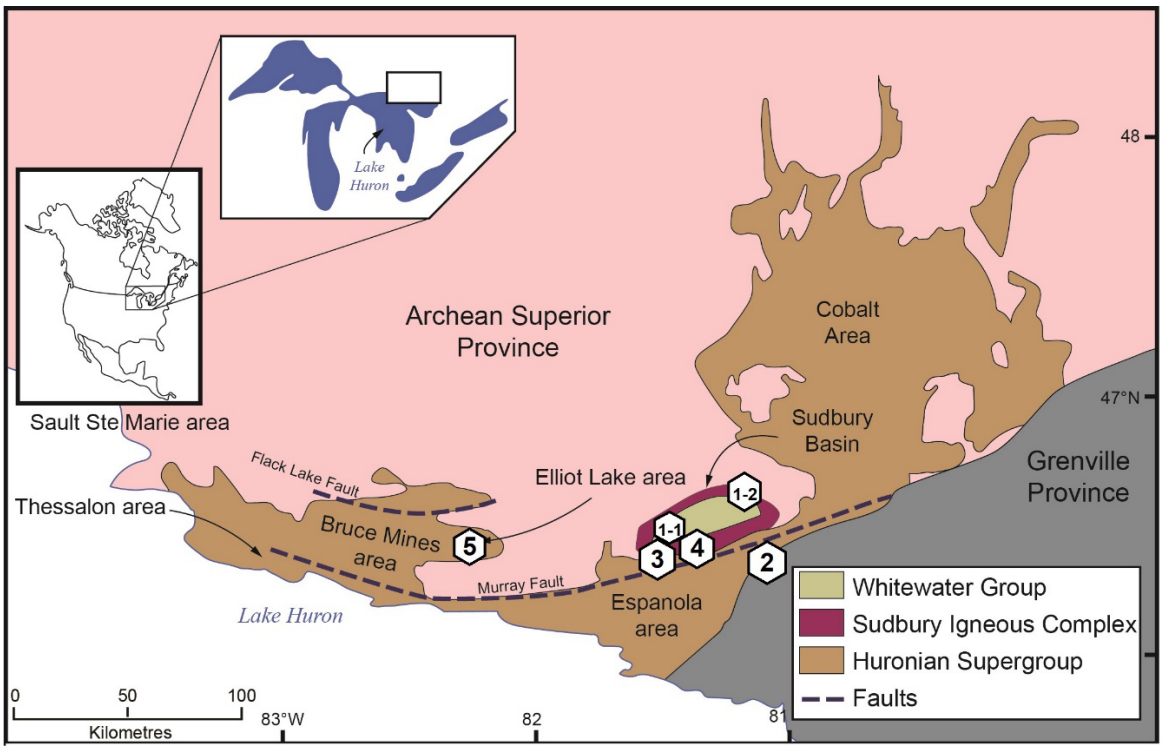


Figure 1. Map showing the location of the five field trips offered in 2022.

Introduction, safety considerations and acknowledgements

Michael Easton

Ontario Geological Survey, 933 Ramsey Lake Road, Sudbury, Ontario P3E 6B5

and

Wouter Bleeker

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Sudbury is located near the boundary between 3 major geological provinces (the Archean Superior Province, the Paleoproterozoic Southern Province, and the Mesoproterozoic Grenville Province) and the largest preserved ancient impact crater on Earth. Consequently, it is an ideal setting for geological field trips. Despite the impact of Covid-19 related health-measures on meeting planning and organization, 3 pre-meeting and 2 post-meeting field trips were available for delegates to the 68th Annual Institute on Lake Superior Geology (ILSG) meeting in 2022 (*see* Figure 1, opposite).

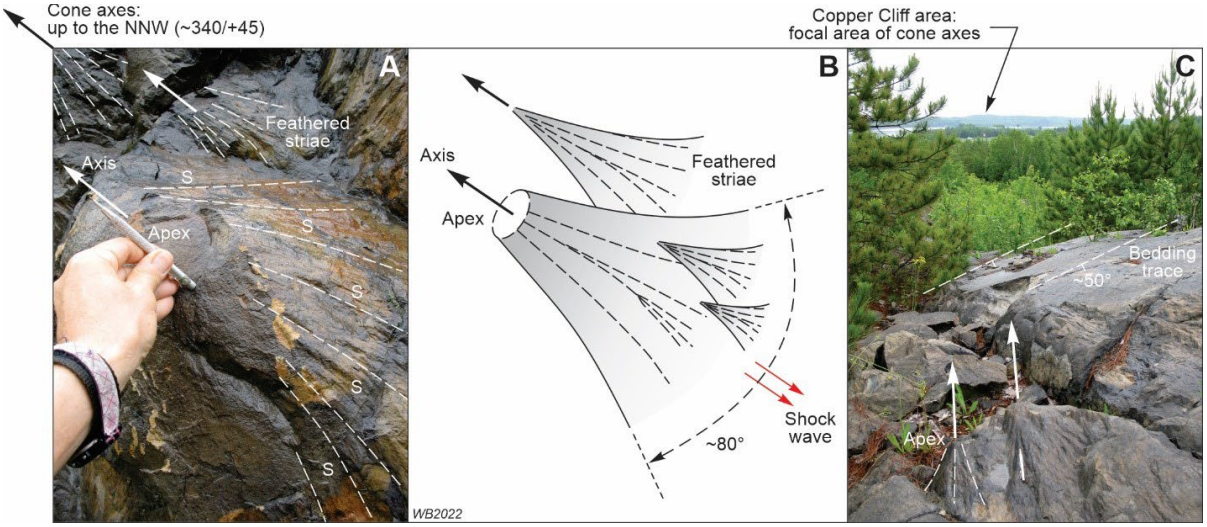
This volume is intended to serve not only as a guide for 68th ILSG field trip participants but also as a reference for those planning to revisit these areas at a later date. Consequently, we have included UTM coordinates in the NAD 83 datum for stops, as well as instructions on how to reach them. As some of the stops are on private and/or staked land, please be sure to obtain the land owners' permission before entering their land. Contact the staff of the Resident Geologist Program of the Ontario Geological Survey in Sudbury for current ownership information.

The field trips, for the most part, will be visiting stops along either major highways or municipal roads. Please take care when crossing or parking along these roads.

In the case of Trips 1 and 3, some stops are on property owned by mining companies, who granted special permission to the ILSG trip leaders and participants to access these properties. It will not be possible for the average guidebook user to revisit these stops.

We would like to thank all the other authors who contributed to this field guide, all those who provided comments and/or assisted with the running of the field trips themselves (Manuel Duguet, Peter MacDonald, Alinda Aubin, Matthew Eles, and Monica Easton). In addition, the efforts of Johanne Roux and Carlo Casrechino in producing the guidebooks in a timely fashion is greatly appreciated.

We also appreciate the assistance and co-operation of the exploration and mining companies in providing access and information concerning their properties, particularly Vale Canada, Limited Crean Hill Mine, (Whistle Mine), Lonmin PLC, Wallbridge Mining Company Limited (Parkin, Trill, Hess), SPC Nickel Corporation (Worthington), KGHM International (Podolosky), Sudbury Integrated Nickel Operations (a Glencore Company) and North American Nickel (distal Whistle). We also thank the City of Greater Sudbury for providing access to a trip stop on field trip 4.



Frontispiece: Classical "shatter cones" in the shocked footwall and target rocks of the 1850 Ma Sudbury impact crater, well developed in quartzites of the *circa* 2.4 Ga Mississagi Formation. With the newly recognized knowledge, in the mid- to late-1950s, that these conical, radiating fracture surfaces represent unique "trace fossils" for the high-velocity, extremely high-pressure shock waves associated with meteorite impacts (Dietz, 1959), Sudbury was quickly recognized, in 1962, as an astrobleme—the scar of an ancient impact crater (see Dietz, 1964; and Dietz and Butler, 1964).

Field Trip 1 – A Traverse Across the Sudbury Structure—Earth's Largest Preserved Impact Crater

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This two-day field trip involves a traverse across the deformed 1850 Ma Sudbury impact structure, from the older country rocks and brecciated target rocks to the south of the preserved melt sheet, across the entire folded impact structure, the melt sheet, and crater fill, and onto its northern rim. Day 1 will concentrate on the regional cross-section/traverse and will introduce many of the key aspects—and controversies—associated with the structure. Day 2 will finish the regional traverse and allow time to examine some of the ore environments in more detail, along the basal contact of the differentiated melt sheet and into the brecciated footwall.

Note 1: Ore environments to be examined are dependent on company approval to access properties, as well as the evolving situation regarding COVID-19.

Note 2: This trip is a more detailed examination of the Sudbury impact structure than that offered by post-meeting Trip #4. Trip #4 provides more of an overview of the structure.

Day 1

Wouter Bleeker[1] and Sandra Kamo[2]

[1]*Geological Survey of Canada, Ottawa*
[2]*Jack Satterly Geochronology Laboratory, University of Toronto*

1. Introduction

This two-day fieldtrip will involve a traverse across the deformed Sudbury structure, one of the world's largest and oldest preserved meteorite impact structures: from deformed and brecciated older country rocks in the south, across the folded impact structure, the melt sheet and crater fill, onto the northern rim, and into the footwall rocks below (Figure 1). Day 1 will focus on the regional cross-section and traverse and will introduce many of the key aspects—and controversies—associated with this unique structure. Day 2 will finish the regional traverse and allow time to examine some of the ore environments in more detail, along the basal contact of the differentiated melt sheet and into the brecciated footwall. Due to the on-going but hopefully waning Covid-19 situation, the issue of access to company properties remains somewhat fluid and some last-minute changes may have to be made in terms of field trip stops and localities. Nevertheless, this two-day trip will allow most major points of interest regarding the Sudbury impact structure to be addressed and discussed.

With well over a century of geological research in the area, the literature on the Sudbury structure is voluminous (e.g., Coleman, 1905; Thomson, 1956; Hawley, 1962; Dietz, 1964; Souch et al., 1969; Naldrett et al., 1970; Brocoum and Dalziel, 1974; Krogh et al., 1982; Pye et al., 1984 and all contributions therein; Faggart et al., 1985; Grieve et al., 1991; Dickin et al., 1992; Butler, 1994; Wu et al., 1995; Spray et al., 2004; Lightfoot and Zotov, 2005; Zieg and Marsh, 2005; Ames et al., 2002, 2008a,b; Bleeker et al., 2015; Papapavlou et al., 2018; and numerous other papers listed in the reference list). Yet many key questions remain. For instance, how big was the original impact structure? And where was "Ground Zero", i.e. the centre of the impact? Is there a global Sudbury fall-out layer and, if so, where is it? How much of the melted footwall geological heterogeneity has been inherited in the differentiated melt sheet? Given the size of the structure and its transient crater, did it trigger any mantle melting? And, with respect to the complex spectrum of observations on the ores and the footwall rocks: where do impact processes and crater-modification processes stop, and where do regional deformation and hydrothermal-metamorphic processes take over?

In the context of an Institute of Lake Superior Geology (ILSG) fieldtrip, one of the questions raised above is particularly pertinent: is the accretionary lapilli layer recognized at the disturbed top of the Gunflint Formation in the Lake Superior area indeed the Sudbury event and fall-out layer (Figure 2), as is permissible and as has been argued based on current evidence (e.g., Addison et al., 2005, 2010; Cannon et al., 2010), or does that horizon represent a different event layer? From an ECREE perspective ("extraordinary claims require extraordinary evidence", Carl Sagan), a unique link of this prominent event layer to the Sudbury area still remains to be documented, and other localities of an 1850 Ma global event layer need to be demonstrated (Figure 2).

Among the largest preserved impact craters on Earth (Dietz, 1964; Dietz and Butler, 1964), Sudbury's very thick (~3–5 km) and differentiated melt sheet (the "Sudbury Igneous Complex", hereafter SIC; Pye et al., 1984 and contributions therein) is unique. Why? Perhaps this can be reasoned away by other comparable melt sheets not being preserved (e.g., Vredefort) or not yet fully explored (Chicxulub), or perhaps was Sudbury bigger than the current consensus (~200 km final crater diameter; *see* Grieve et al., 1991; Grieve, 1994; Butler, 1994; Spray et al., 2004), and is it in a class of its own?

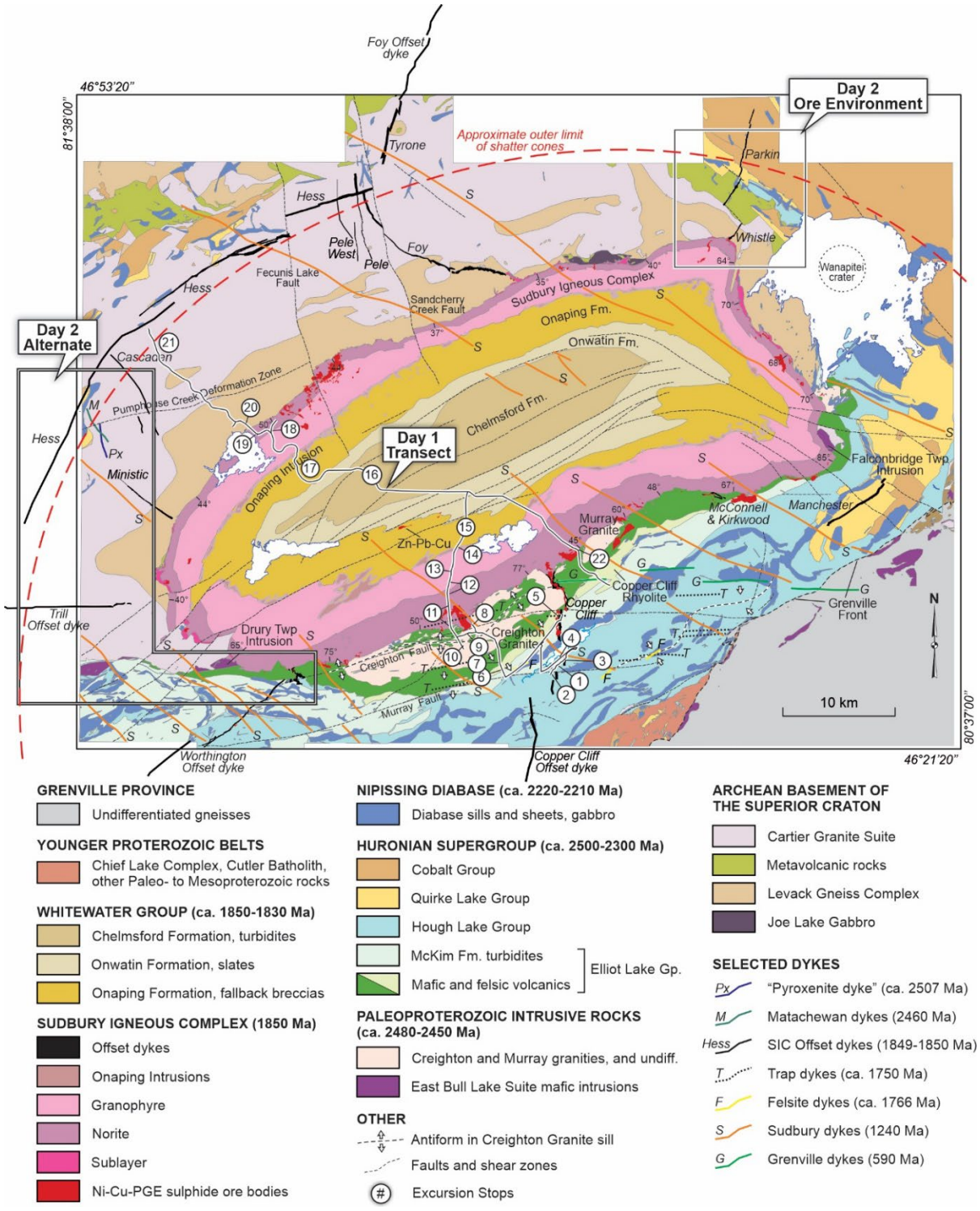


Figure 1: Geological map of the Sudbury area centered on the erosional remnants of the deformed Sudbury impact structure, the Sudbury Igneous Complex (SIC); *after* Bleeker et al. (2015), and adapted from Dressler (1984) and Ames et al. (2005). Day 1 fieldtrip stops are indicated, as is the area we will visit on Day 2 (and a possible alternate area for Day 2). The arcuate red dashed line marks the outer limit of observed shatter cones in the footwall. The *circa* 1.0 Ga Grenville Front truncates the area in the southeast.

Finally, Sudbury is fascinating from a science history perspective: how it was discovered and how its interpretation was slow to change from, originally, a largely mafic igneous complex or lopolith — "the Sudbury Irruptive" — in part or wholly derived from mantle melting (e.g., Wilson, 1956; prior to the work by R. Dietz in the early 1960s); to a hybrid igneous and impact-generated structure (*see*, for instance, the various contributions in Pye et al., 1984; *see also* Dietz, 1964[1]); and, finally, to an impact-only crater, melt sheet, and crater fill complex with little to no mantle input and merely modified by deformation (e.g., Stöffler et al., 1989; Grieve et al., 1991; Grieve, 1994). With respect to the latter perspective, the pendulum only changed in the late 1980s and early 1990s when detailed isotopic data became available showing that most if not all of the melt sheet was derived from melting of the crust (e.g., Faggart et al., 1985; Stöffler et al., 1989; Walker et al., 1991; Dickin et al., 1992; Deutsch, 1994 and references therein). From this slowly evolving historical perspective, the brilliant early 1960s papers by Robert Dietz, who in 1962 quickly confirmed "shatter cones" in the footwall rocks around the Sudbury complex (*see* frontispiece of this guidebook), and from there confidently posited a meteorite impact origin, stand out as even more remarkable.[2] This field trip will allow participants to become familiar with this amazing structure and to discuss and debate all these first-order questions. The current fieldtrip guidebook should be viewed as a preliminary offering, as many details remain to be expanded on.

2. The Sudbury Structure: Geological Setting

The deformed Sudbury impact structure, now preserved as a broadly doubly plunging, synclinal, erosional remnant ~60 km long by ~30 km wide, is situated in the southern Canadian Shield, approximately on the boundary between two main structural provinces, the Archean Superior Province to the north and the Paleoproterozoic Southern Province to the south (Figures 1 and 2). The former is represented by the *circa* 2.85-2.64 Ga granite-greenstone terrain of the southern Superior craton, whereas the latter is dominated by moderately to tightly folded and faulted Paleoproterozoic strata of the *circa* 2.50-2.30 Ga Huronian Supergroup (Young, 1973 and contributions therein; Bennett et al., 1991; Young et al., 2001; Rasmussen et al., 2013) overlying Superior craton basement (Figures 3 and 4). The Southern Province trends roughly E-W and, to the west, extends into the Lake Superior area where it is known as the Penokean fold belt, part of the larger 1.87-1.83 Ga Penokean orogen (e.g., Brocoum and Dalziel, 1974; Schulz and Cannon, 2007).

Just to the south of the city of Sudbury, the Paleoproterozoic Southern Province fold belt is truncated by the relatively sharply defined deformation front of the *circa* 1.1-1.0 Ga Grenville Front *sensu stricto* (see front #4, Figure 2). This front trends from northeast to southwest and represents the northern boundary of the complex, multi-cyclic and very extensive Grenville orogen that rims Proterozoic Laurentia to the southeast and played major role in building the late Proterozoic supercontinent Rodinia. In more detail, the Grenville Front *sensu stricto* represents, in the Sudbury area, the relatively well-defined deformation front of the terminal collisional phase of the Grenville orogen, an orogen that also involves older Paleo- to Mesorproterozoic events, some of which are represented by rocks units right along the Grenville Front (e.g., see front #3 in Figure 2).

[1] Although quickly recognizing the fundamental impact origin of the structure in 1962, based on his identification of shatter cones in the footwall rocks, in his seminal 1964 paper Dietz maintains an intrusive origin for much of the Sudbury Igneous Complex.

[2] A bibliography of relevant papers by Dietz is included at the beginning of the reference list and they make for an interesting read. After convincing himself, in the mid-1950s, that shatter cones were a key indicator for high-energy meteorite impacts, he quickly clarifies a large number of impact structures, such as Sudbury, which were previously seen as enigmatic and attributed to "crypto-volcanic" (i.e. endogenic) processes. In 1961 he coined the term "astroblemes" for these structures (i.e., "star wounds" or impact scars; Dietz, 1961) to highlight the fundamental role of impact processes here on Earth.

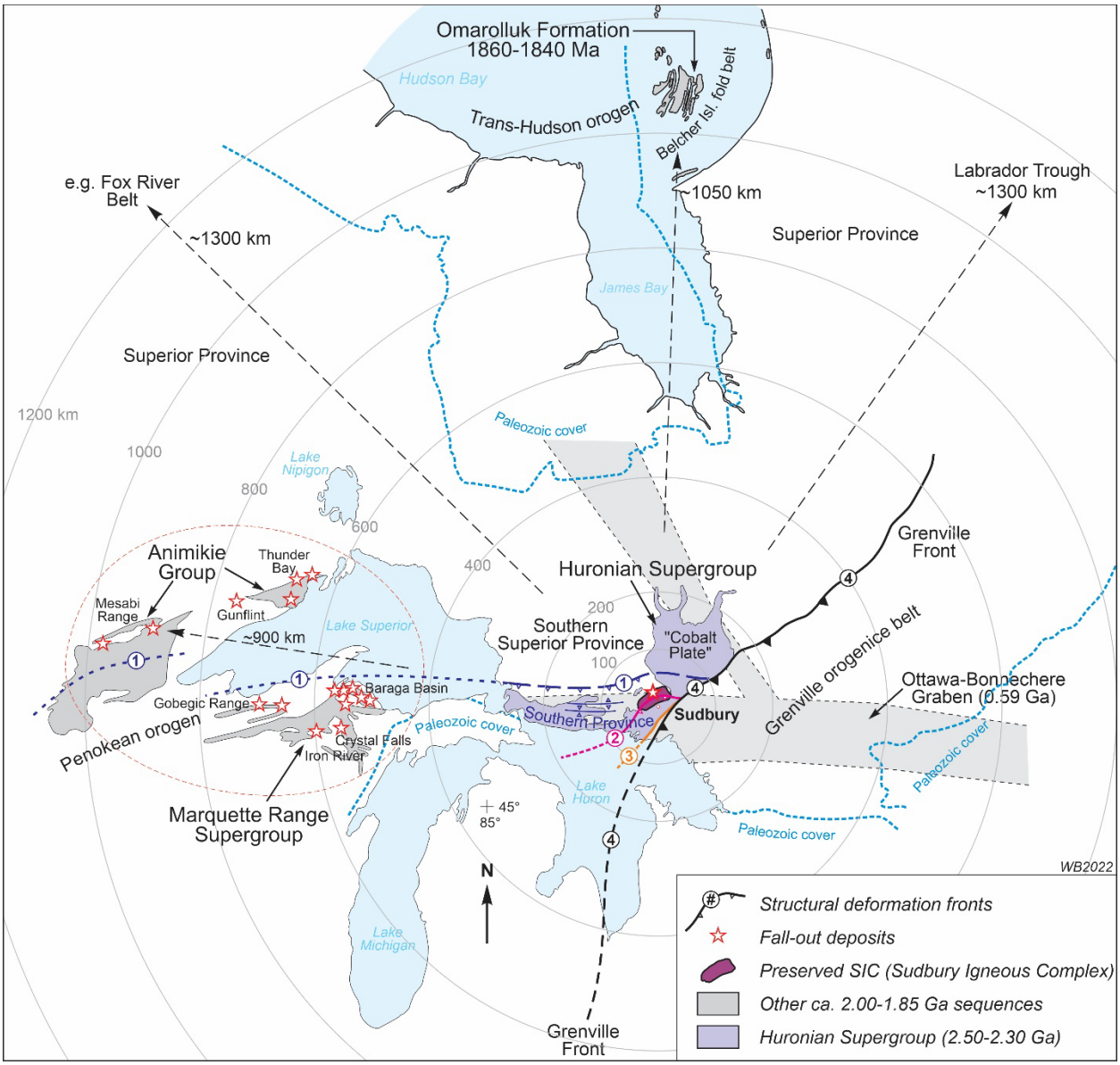


Figure 2: General location of the deformed Sudbury impact structure on the approximate boundary of the Archean Superior Province and the Paleoproterozoic Southern Province. The latter is broadly part of the "Penokean fold belt" and the associated orogen that rims the southern margin of the Superior craton (e.g., Card et al., 1972; Brocoum and Dalziel, 1974; Card et al., 1984; Holm et al., 2007; Schulz and Cannon, 2007). Structural fronts younger than the *circa* 1.87-1.84 Ga Penokean *sensu stricto* (front #1) also contributed to the deformation of the Southern Province (i.e. fronts #2 and 3). South of Sudbury, the Southern Province is truncated by the main *circa* 1.1-1.0 Ga Grenville Front (front #4; Davidson, 1997). Also shown is the approximate footprint of the *circa* 0.6–0.5 Ga Ottawa-Bonnechere Graben (in light grey), which is manifested in the Sudbury area by E-W-trending olivine-bearing diabase dykes. Red star symbols indicate the localities where the accretionary lapilli event layer at the top of the Gunflint Formation and correlative strata has been identified (e.g., Cannon et al., 2010). The figure also draws attention to other *circa* 1870-1840 Ma sequences preserved in the Canadian Shield, only at marginally larger distances than the Mesabi Range locality, where a comparable Sudbury fall-out layer remains to be identified. Map *adapted from* Wheeler et al. (1996), Young (1983), and Cannon et al. (2010).

A simplified NNW-SSE cross-section through the area illustrates the main pre-impact, litho-tectonic elements that define the wider Sudbury area (Figure 3). A summary stratigraphic column, highlighting the main features of the Paleoproterozoic Huronian Supergroup in more detail, is shown in Figure 4. The age range of the Huronian Supergroup is constrained by the onset of rifting and associated large-scale mafic magmatism at *circa* 2500-2480 Ma (Krogh et al., 1984), defining its base, and final emplacement of extensive mafic sill complexes at *circa* 2250 Ma (May Township sills, Bleeker et al., in prep.) and *circa* 2217 Ma (Nipissing sills; Corfu and Andrews, 1986; Noble and Lightfoot, 1992; Bleeker et al., 2015; Davey et al., 2019). The latter intrusive episodes provide a minimum age for the supergroup. Thin felsic ash layers in the fine-grained sedimentary rocks of the Gordon Lake Formation, in the upper part of the supergroup, have an approximate age of *circa* 2308 Ma, based on SHRIMP dating of zircons (Rasmussen et al., 2013) on a sample obtained from drill core through the formation.

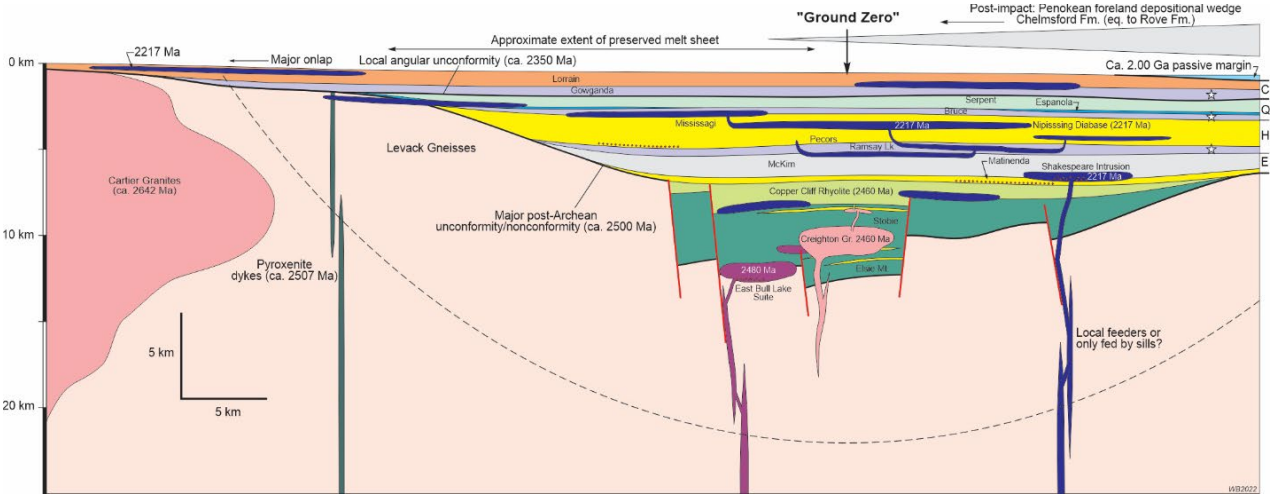


Figure 3: A simplified cross-section through the Sudbury area from NNW to SSE, showing the main litho-tectonic elements that define the geology of the area, prior to Penokean and younger deformation and prior to the 1850 Ma impact. Note the lower Huronian rift structure and associated rift fill (the Elliot Lake Group, E), which is well developed and exposed in the Sudbury area, in part due to impact-induced uplift and exhumation. The estimated position of "Ground Zero" is indicated above the section, based on both ring-like structures in the foreland (Butler, 1994) and a statistical intersection of shatter cone axes (Bleeker, unpublished). A *circa* 2.0 Ga passive margin sequence, comparable to the Gunflint Formation in the Lake Superior area, is indicated for general comparison only but is not preserved in the Sudbury area. Post-impact, the area was covered by the depositional wedge of Penokean foreland basin deposits, the upper shale dominated Onwatin Formation and the turbiditic Chelmsford Formation, which are preserved only within the doubly plunging syncline of the "Sudbury Basin". These turbiditic deposits are comparable and likely directly correlative to the Rove Formation of the Lake Superior area. The three stars on the right side of the section indicate glacially-influenced formations of diamictite, from bottom to top: the Ramsay Lake, Bruce, and Gowganda Formation diamictites. Based on global correlations and U-Pb age dating in the Transvaal Basin of South Africa, we know that the Ramsay Lake Formation glacial episode ended at *circa* 2426 Ma (Gumsley et al., 2017).

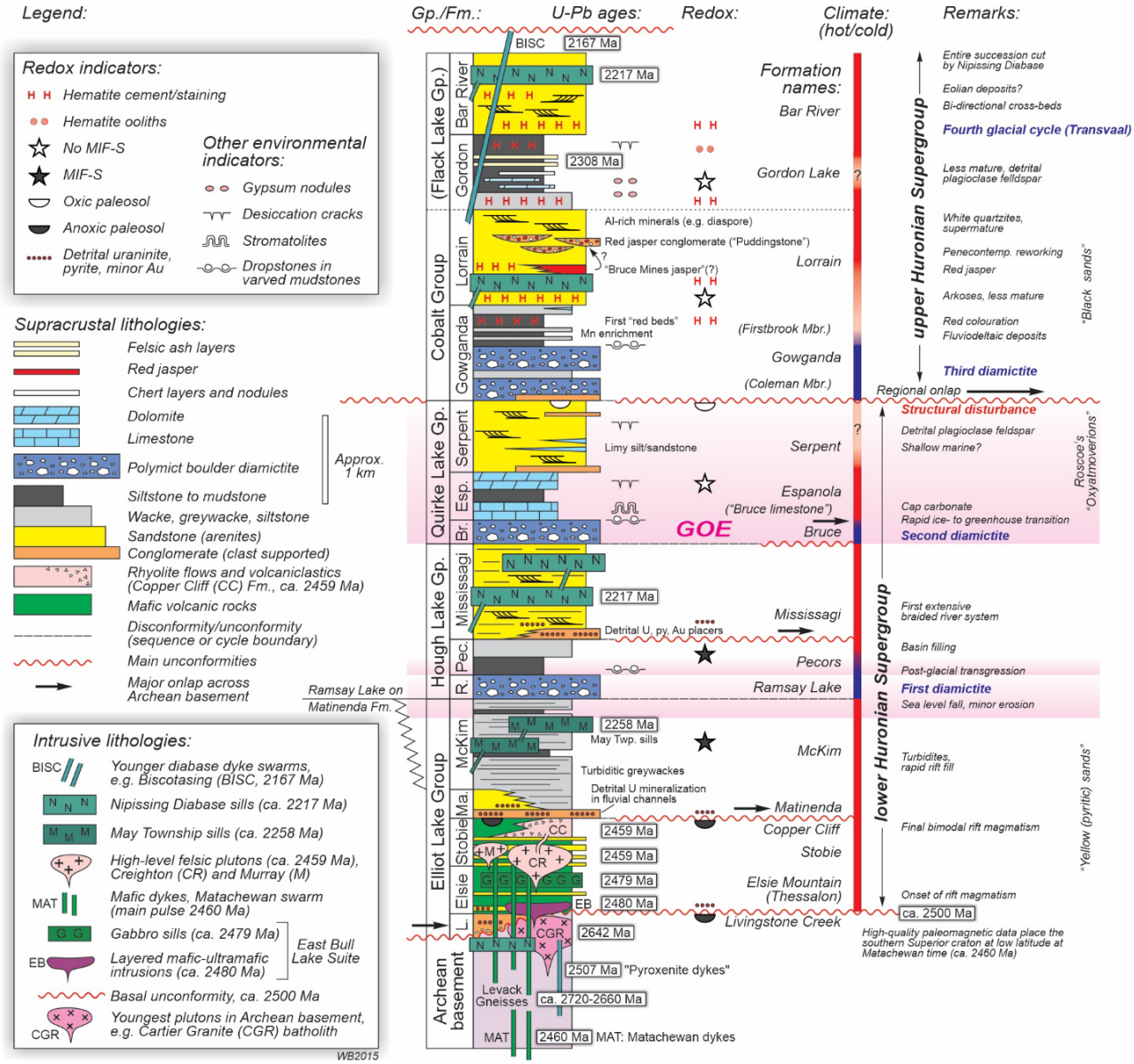


Figure 4: Summary stratigraphic column of the 2.50-2.30 Ga Paleoproterozoic Huronian Supergroup, unconformably overlying Archean basement of the southern Superior craton. Compiled from numerous sources: Young (1973) and contributions therein; Bennett et al. (1991); Young et al. (2001); Gumsley et al. (2017). Age data from: Krogh et al. (1982) and Krogh et al. (1984); Corfu and Andrews (1986); Noble and Lightfoot (1992); Heaman (1997); Rasmussen et al. (2013); Bleeker et al. (2015); Davey et al. (2019).

3. More on the 2.50-2.30 Huronian Supergroup

Age range, thickness, Archean-Proterozoic boundary

The Huronian Supergroup (Figure 4) is of more than just local interest. With its extended age range and ~6–12 km total stratigraphic thickness, it represents one of the principal, and best-preserved Paleoproterozoic stratigraphic records in the world, perhaps only rivalled by the Transvaal Supergroup of southern Africa, and correlative units of the Hamersley Basin overlying the Pilbara craton. Its basal unconformity and low-grade metamorphic state, overlying Archean granite-greenstone terrain, played a major role in the debate and final definition of the Archean–Proterozoic boundary, one of the most fundamental boundaries of the terrestrial geological time scale.

GOE: the Great Oxidation Event, pyritic placer deposits, continental red beds

Equally important from a historical point of view, the Huronian Supergroup straddles the *circa* 2.4–2.3 Ga "Great Oxidation Event" (GOE, see Figure 4; e.g., Roscoe, 1973[3]; Holland, 1978, 1984, 2002; Prasad and Roscoe, 1996; Bekker et al., 2004; Gumsley et al., 2017), which marks the initial rise of atmospheric oxygen across the critical threshold above which oxidized surface environments promoted the deposition of hematite-stained red sandstones ("red beds"). The world's oldest genuine terrestrial red beds occur in the Cobalt Group of the upper Huronian, specifically the Gowganda and overlying Lorrain formations. The elevated oxygen levels no longer allowed prolonged preservation of detrital pyrite and associated heavy minerals, in terrestrial sedimentary environments, thus resulting in pyrite placer deposits largely disappearing from the geological record. Hence, no such deposits are known from the upper Huronian Supergroup, but, with associated gold and uraninite, they form minor but important components of the lower part of the supergroup that was deposited prior to the GOE transition. Pyrite-uraninite placers form important placer deposits in the basal Matinenda Formation (Figures 3 and 4) and were mined extensively for uranium in the Elliot Lake area west of Sudbury. Gold-bearing pyrite placers are currently the subject of active exploration northeast of Sudbury and are thought to be part of the conglomerates and cross-bedded sandstones at the base of the Mississagi Formation (Long et al., 2011; Whymark and Frimmel, 2018).

The emergence of these important insights involving atmospheric evolution, increasing oxygen levels, detrital pyrite, and the first red beds were largely based on work in the Huronian Supergroup and correlated sequences, such as the Snowy Pass Supergroup overlying the Wyoming craton (e.g., Roscoe and Card, 1992, 1993; Prasad and Roscoe, 1996).

GOE: disappearance of the mass-independent sulphur isotope fractionation signature

Correlated with the observable transition "detrital pyrite out, red bed sandstones in", more recent research has shown that there is also a fundamental shift in global S-isotopic signatures, particularly the disappearance of anomalous "mass-independent fractionation" (MIF) of the ^{33}S isotope relative to ^{32}S and ^{34}S isotopes (Farquhar et al., 2000; Farquhar and Wing, 2003). This MIF signature of ^{33}S is understood to result from UV-induced processes in the upper atmosphere and this MIF signal can only survive and be transmitted to the sedimentary record at very low total oxygen atmospheric levels, estimated at $<10^{-5}$ times present atmospheric level (PAL) of oxygen (Farquhar and Wing, 2003). These insights have revolutionized the understanding of both the shallow and deep sulphur cycles in the last two decades and shed new light

[3] Although now widely referred to as the "Great Oxidation Event" (GOE), following Holland and others, Stew Roscoe through his work in the Elliot Lake area, on the pyrite-uraninite placer deposits there, was one of the first to draw attention to this important transition and called it the "oxyatmoversion". In his 1973 paper he also draws attention to the obvious potential of this transition in terms of a first-order time scale boundary.

on a host of fundamental recycling processes. As shown in Figure 4, detrital pyrite and the MIF-S signature are absent above the Mississagi Formation.

Glacial episodes, Snowball Earth events, cap carbonates

Stratigraphic and sedimentological studies of the Huronian Supergroup also played a key role in the recognition of early Precambrian glacial deposits, and possibly global-scale glacial events. Unsorted diamictite deposits, almost certainly tillites or reworked tillites, are recognized at three stratigraphic levels within the Huronian Supergroup (Figure 3 and 4): in the Ramsay Lake Formation, the Bruce Formation, and in the post-GOE Gowganda Formation. Some of these diamictite deposits have other associated features that confirm a glacial origin, such as dropstones in overlying varve-like siltstones, and glacially polished clasts with striae. This is certainly the case for the Gowganda Formation (Young, 1983; Young and Nesbitt, 1985; Young et al., 2001).

Paleomagnetic evidence places the Superior craton at low latitudes in the earliest Paleoproterozoic (e.g., Evans and Hall, 2010; Salminen et al., 2014). This, together with glacial deposits in what was at least in part a marine basin, thus places glacial deposits at sea level, and close to the equator and far from the poles. In turn, observations such as these argue for glacial events of global significance, and events that can be correlated to other cratons and similar sedimentary successions such as the Transvaal basin of southern Africa (e.g., Gumsley et al., 2017). This is particularly relevant for the second of the three glacial events represented by the Bruce Formation diamictites. This formation is overlain by the only carbonates in the Huronian succession, the Espanola Formation (*see* Figure 4). These carbonates may represent deposits that are thought to have formed in response to rapid deglaciation of global ice cover and, in this context, are referred to as "cap carbonates" (e.g., Kirschvink, 1992; Hoffman et al., 1998; Hoffman and Shrag, 2000; Kirschvink et al., 2000). If so, these carbonates would allow global correlations and constitute an ideal, globally synchronized time scale boundary, as they do in the Neoproterozoic.

Epicontinental rift and sag succession or passive margin sequence?

Various authors have, somewhat uncritically, referred to the Huronian Supergroup as a rift and passive margin sequence, associated with the breakup of the Superior craton. That implies that the Superior craton broke up along its southern margin sometime around 2.4 Ga following the formation of the lower Huronian rift succession. There is, however, little evidence for this early breakup. Clearly the Superior craton, with its present outline, is just a fragment of a much larger ancestral Archean supercraton, which Bleeker (2003, 2004) has referred to as "Superia". Breakup of Superia was clearly progressive but may have only started with the very extensive *circa* 2.22 Ga Nipissing and related mafic magmatic events or large igneous provinces (LIPs), and likely even later along the southern margin of the craton.

The specific LIP event that most likely represents initiation of breakup along the southern margin of the Superior is the 2125-2100 Ma Marathon event (Halls et al., 2008; Davey et al., 2020, 2022), with very extensive mafic dyke swarms projecting north into the craton and bimodal magmatism along the southern margin of the craton. The Wyoming and Karelia-Kola cratons are among the cratonic fragments that broke away at that time, as they can be matched to the southern Superior, based on multiple lines of evidence, prior to 2.1 Ga breakup (e.g., Roscoe and Card, 1993; Bleeker and Ernst, 2006; Kilian et al., 2016a, b). This alternative scenario argues for the entire Huronian Supergroup to represent a long-lived intra- and epi-continental rift and sag basin, without a proximal passive margin prism. Consequently, this puts the Huronian Supergroup in a different perspective and has implications for global reconstructions and compilations, such as those of successions thought to represent passive margin sequences through time (Bradley, 2008). Similar arguments very likely apply to other Proterozoic sequences that have been too easily characterized as passive margins.

4. Large Igneous Provinces (LIPs) and Mafic Magmatic Events of the Wider Sudbury Area

Mafic magmatic events

Here we briefly list and describe some of the large mafic magmatic events in the wider Sudbury area. These events are important as temporal and structural markers, either pre-dating certain structural events, or the impact event itself, or post-dating such events and thus providing minimum age constraints. From old to young, a brief summary of these magmatic events would include (e.g., Krogh et al., 1984; Kamo et al., 1995; Heaman, 1997; Noble and Lightfoot, 1992; Ernst and Bleeker, 2010; Bleeker et al., 2015):

- Major pyroxenite dykes: *circa* 2507 Ma, perhaps part of the broader Mistassini event? These dykes, which occur just north of the SIC, may indicate the onset of the Huronian rifting event (Bleeker et al., 2015).
- Matachewan-I: *circa* 2480 Ma layered intrusions and sills, dykes; the East Bull Lake Suite.
- Matachewan-II: *circa* 2460 Ma major diabase swarm, main pulse of Matachewan dykes.
- May Township sill complex: *circa* 2250 Ma, large sills to the west of the Sudbury area.
- Nipissing sills and Senneterre dykes: *circa* 2217 Ma, volumetrically important in the immediate Sudbury area.
- Biscotasing dykes: 2167 Ma dykes, mostly north of Sudbury.
- Marathon dykes: *circa* 2110 Ma dykes, mostly west of Sudbury.
- Lauzon Lake dykes: *circa* 1950 Ma, major NW-trending dykes west of Sudbury.
- Alkaline dykes, lamprophyres and carbonatite intrusions, *circa* 1880 Ma; e.g. the Spanish River Complex north of Sudbury.
- Thin mafic sills/subhorizontal sheets: undated and of yet unknown significance.
- Lamprophyre dykes: undated and of yet unknown significance; probably more than one event.

- Sudbury impact event and associated dykes: 1850-1849 Ma.

- Trap dykes: *circa* 1750 Ma, numerous E-W trending diabase dykes cutting across the Sudbury South Range, still affected by metamorphism.
- Alkaline intrusions in the wider area, e.g. the Croker Island Intrusion, southwest of Sudbury, Mesoproterozoic in age.
- Larder Lake dykes: *circa* 1270 Ma, undeformed olivine diabase dykes similar to Sudbury dykes.
- Sudbury dykes: *circa* 1235 Ma, abundant NW-trending undeformed olivine diabase dykes cutting across the area and the Sudbury structure, but themselves cut and truncated by the Grenville Front.
- Grenville dykes: *circa* 590 Ma, E-W-trending dykes cutting across the Sudbury structure and Grenville Front, and a manifestation of the extensive Ottawa-Bonnechere Graben system.
- Kimberlites and associated intrusions in the wider area, Jurassic to Cretaceous.

Some of the pre-impact mafic magmatic events are rather voluminous in the Sudbury area, and thus a significant component of the overall target rocks. This is particularly true for: 1) the early Matachewan (I) mafic rocks, which form mid-sized layered intrusions in the general area (e.g., James et al., 2002), at or close to the Archean–Paleoproterozoic (i.e. basal Huronian) unconformity (Figure 3), and less voluminous dykes and sills; and 2) the very extensive Nipissing Diabase sill complex, which invaded large parts of the Huronian Supergroup stratigraphy as well as the basement immediately below the unconformity (*see* Figures 3 and 4).

Some of these units have minor, to locally significant, Ni-Cu-PGE magmatic sulphide mineralization; for instance, the Shakespeare Intrusion that is part of the Nipissing magmatic event (e.g., Sproule et al.,

2007; Davey et al., 2019). For these reasons (overall volume, sulphur (S), metals (Ni, Cu, PGEs)), they have featured and continue to feature in the debate on metal sources and elemental mass balances in the differentiated Sudbury melt sheet and the associated magmatic sulphide ores. In other words, did certain geological units in the target rocks play an important role in the sulphur and metal budgets of the impact melt sheet, and thus in early sulphur saturation and in the overall metal endowment in the orebodies, or are such roles insignificant in the overall melt sheet evolution and the formation and segregation of the ores? Clearly, immediately following the impact, and the generation of a superheated melt sheet, early sulphide saturation was critical and can be clearly demonstrated (Figure 5); but overall metal budgets may not require enriched source rocks if sulphur saturation was indeed early and given the enormous volume of very hot impact melt available. In the latter scenario, it is simply all about the efficiency of early sulphide saturation, exsolution and segregation, sulphide droplet formation, and the "rain out" of sulphide globules enriched in chalcophile metals.

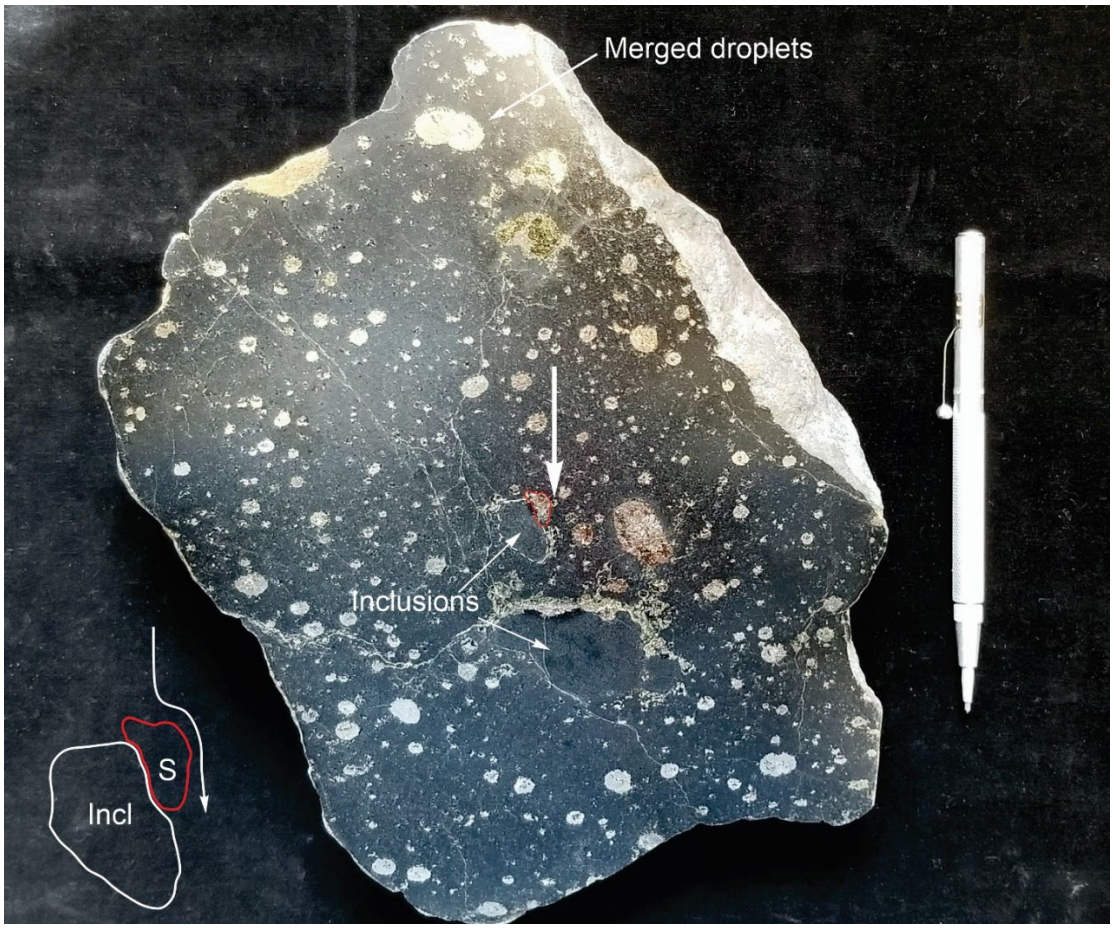


Figure 5: A polished slab of Fe-Cu-Ni magmatic sulphide droplets, millimetres to centimetres in size, in ore from the proximal part of the Copper Cliff "offset dyke". This dyke was injected into the footwall of the impact crater, during an early stage of the melt sheet evolution. The host rock consists of relatively unfractionated early mafic melt sheet material formed near the base of the melt sheet. Sulphide droplets were actively "raining out" (down in the picture) and physically interacting with mafic inclusions on their way down. Textures such as these provide clear evidence for early sulphide saturation in the melt sheet and that the "rain out" of exsolved sulphide melt droplets was the first-order process collecting magmatic sulphide ores at or near the basal contact of the melt sheet. Note the dumbbell structure of two merging sulphide globules near the top of the sample.

Several of the intrusive events are also of key importance as structural markers. For instance, the extensive Nipissing Diabase sills within the Huronian succession, and locally in basement just below the Huronian unconformity, are folded with the strata of the Huronian Supergroup and, where well-exposed, sill contacts are largely concordant with the sedimentary layering in the immediate host rocks; although locally the sills are cross-cutting and dyke-like due to a "saucer-shaped" sill geometry (*see* Figure 3). However, the largely concordant nature of the observed contacts of many of the sills, together with a notable absence of clearly documented examples of such sills cutting across both limbs of previously folded Huronian strata, strongly favours an interpretation of the Huronian being largely unfolded, or perhaps only very weakly folded and/or locally tilted, at the time of widespread Nipissing sill emplacement at *circa* 2217 Ma (Bleeker et al., 2015). This important observation has major implications for the concept of the "Blezardian orogeny" and, ultimately, how to interpret the Sudbury structure (*see below*), and is fully supported by observations on the orientation of shatter cones.

5. The "Blezardian Orogeny", Fact or Fiction?

The lowermost Huronian Supergroup is intruded by a number of granite plutons in the Sudbury area, one of the major plutons being the Creighton Granite on the South Range, just west of Sudbury (Figure 1). As part of early attempts to systematize the geology of the Canadian Shield and divide this enormous territory into "structural provinces" based on regional deformation patterns and tectono-magmatic events (e.g., Wilson, 1949; Stockwell, 1982), late-stage granites were generally seen, and indeed correctly in many places, as the terminal phase of a major orogenic episode. As it had been clear since the early mapping of the southern Canadian Shield that the folded Huronian Supergroup unconformably overlies Archean basement that had previously been affected by major late Archean deformation and metamorphism, this grew into the concept of the two structural provinces: the Archean Superior Province to the north, affected and shaped by a terminal "Kenoran orogeny", and a younger Proterozoic Southern Province and fold belt to the south, affected by deformation attributed to a "Blezardian orogeny" dated by the intrusion of the Creighton Granite and similar granite plutons. Early zircon dating attempts of these (shock-deformed!) granite bodies, based on large, multigrain, highly discordant zircon fractions (unabraded), suggested interpreted upper intercept ages of *circa* 2.35 Ga (Frarey et al., 1982); i.e. a *circa* 2.35 Ga Blezardian orogeny associated with granite magmatism that terminated the Huronian cycle of sedimentation and associated deformation.

Perhaps reasonable at the time, we now know this interpretation is no longer tenable for the following reasons: 1) the granite plutons are lower Huronian rift-related A-type granites, dated at 2455–2460 Ma (Bleeker et al., 2015), not collisional granites as part of a terminal orogenic phase; 2) they only intrude the lowermost Huronian rift volcanics and are deformed with the lower Huronian strata they intrude; and 3) as explained above, the Huronian strata are extensively intruded by Nipissing Diabase sills at 2217 Ma, with most or all of the deformation post-dating the emplacement of the sills.

From Bleeker et al. (2015): "A tightly folded Nipissing Diabase sill has been dated at 2215 ±1 Ma. It is fully conformable with surrounding Huronian strata on the South Range, inconsistent with the concept of a pre-Nipissing "Blezardian orogeny". The main rationale for the Blezardian orogeny was the idea that deformation and intrusion of granite plutons, such as the Creighton Granite, thought to be *circa* 2.35 Ga in age, terminated the depositional history of the Huronian succession (Frarey et al., 1982; Stockwell, 1982). None of these ideas are supported by present evidence. The Creighton Granite is an early Huronian 2455–2460 Ma rift-related granite, not an orogenic granite pluton; folding of the Huronian succession did not commence until well after emplacement of Nipissing Diabase sills and sheets with the onset of Penokean accretion and collision events at *circa* 1860 Ma. Other observations that have contributed to the concept of

a Blezardian orogeny can all be explained without a significant pre-Nipissing deformation event. For instance, saucer-shaped Nipissing sills locally may appear to crosscut Huronian strata and, after superimposed Penokean deformation, could easily lead to confusing field relationships."

With these caveats, does this leave any room for pre-Nipissing deformation of the Huronian? It seems reasonable that there was some local deformation and/or tilting related to movement on rift faults, perhaps local inversion. As shown in Figures 3 and 4, the Huronian succession consists of several major groups, and some of the boundaries between these groups may represent second order sequence boundaries and local, relatively low-angle unconformities. In this context, the basal contact of the aerially extensive Cobalt Group, overlain by coarse conglomerate and diamictite, is particularly relevant. The Cobalt Group far oversteps the lower groups of the Huronian and onlaps onto the Superior craton far to the north. In some areas northeast of Sudbury, careful examination of township geological maps suggests that a low-angle unconformity separates the Cobalt Group from the lower Huronian groups (Hough Lake and Quirke Lake Groups), as depicted in the sections of Figures 3 and 4 (*see also*, Meyn, 1973). These interesting relationships do not equate with a Blezardian orogeny, however.

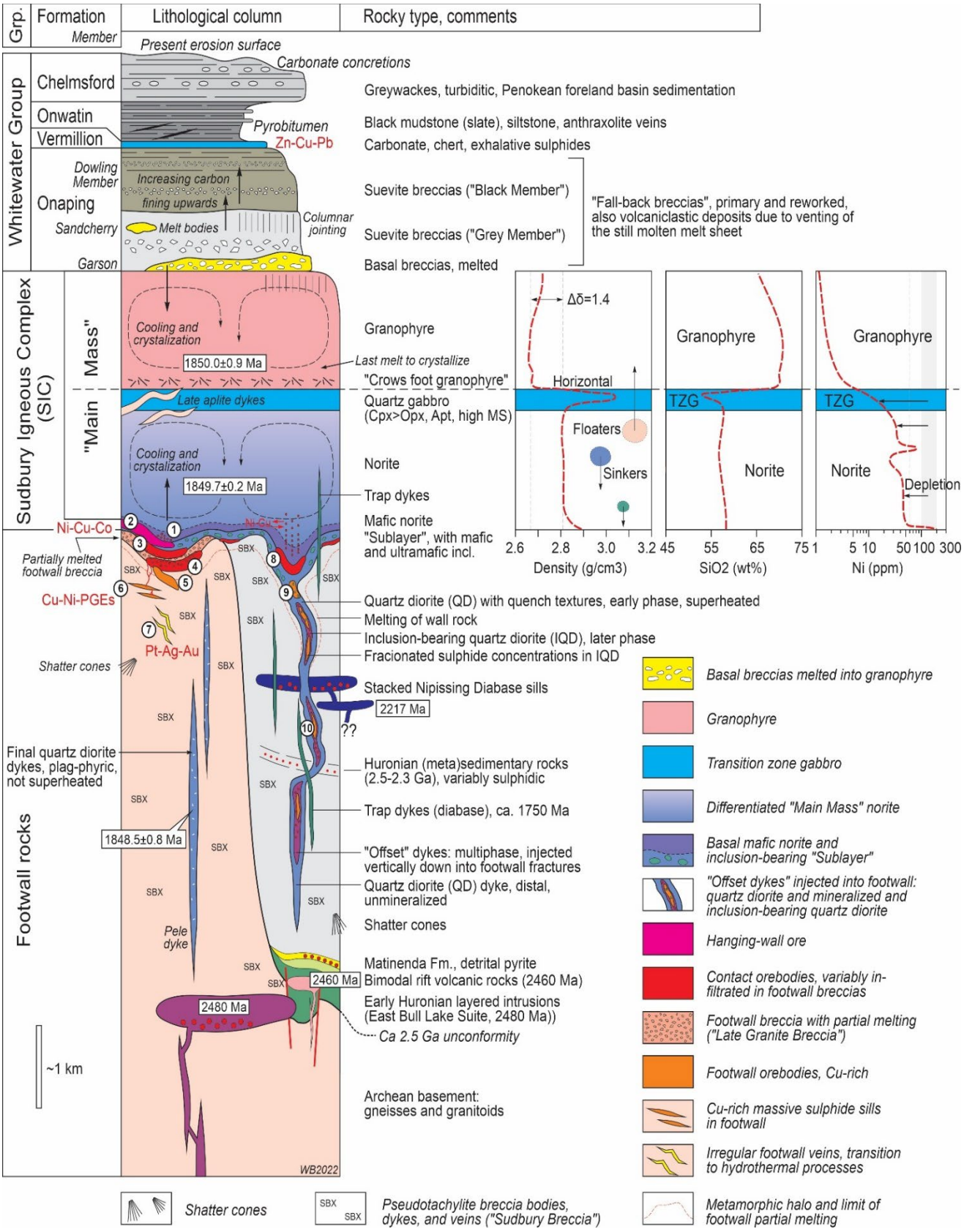
6. Post-impact Deformation

Beyond the observations discussed above, it is clear that most of the deformation and metamorphism of the Southern Province and the Sudbury area, locally intense, post-dates all of the Huronian Supergroup, the Nipissing Diabase sill emplacement, and also the final settling of the differentiated Sudbury melt sheet, as well as the deposition of the overlying Whitewater Group (Figure 6; *see also* Figure 3). The major deformation that folds the Sudbury structure, melt sheet, crater fill, and overlying Whitewater Group into a regional-scale, doubly plunging synclinal structure can be largely attributed to the Penokean orogeny (e.g., Card et al., 1972; Brocoum and Dalziel, 1974), amplified and overprinted to varying degrees, particularly on the South Range and toward the Grenville Front, by younger Proterozoic events (Shanks and Schwerdtner, 1991; Bailey et al., 2004; Papapavlou et al., 2017).

Figure 6 (next page): Simplified lithologic-stratigraphic column for the Sudbury impact structure, its differentiated melt sheet, and the overlying Whitewater Group. Adapted and compiled from various sources, including Naldrett and Hewins (1984), Grieve et al. (1991), Zieg and Marsh (2005), Ames et al. (2005), Bleeker et al. (2015), and Lightfoot (2016). In the context of the present discussion, note the uppermost units of the Whitewater Group, which represent the foreland depositional wedge of the Penokean foreland basin. These sedimentary formations are only preserved in the regional scale, doubly plunging, Sudbury Basin syncline. The turbiditic Chelmsford Formation is similar in age and character to the Rove (Virginia) Formation of the Lake Superior area.

Note the very tight age control on the differentiated melt sheet, including a new age of last crystallizing basal granophyres at 1850.0±0.9 Ma based on chemically abraded, fully concordant zircon data.

Numbers highlight the different orebody settings and types: 1) disseminated sulphides in mafic norite near the base of the SIC; 2 and 3) disseminated to semi-massive sulphides in the Sublayer and along the footwall contact; 4) sulphides infiltrated in the footwall breccia; 5) semi-massive to massive sulphides in footwall rocks, variably fractionated and enriched in Cu; 6) massive sulphide sills deeper within the footwall, (sub)parallel to the footwall contact, very Cu-rich; 7) deeper remobilized veins, with a transition to hydrothermal processes; 8) major sulphide concentrations in funnel-like embayments; 9) sulphide ore within inclusion-bearing diorite (IQD) injected into footwall; and 10) globules to semi-massive sulphides, fractionated and Cu-rich, injected deep into footwall dykes.


Grp.
Formation
Member
Lithological column
Rocky type, comments
Present erosion surface
Carbonate concretions
Whitewater Group
Chelmsford
Greywackes, turbiditic, Penokean foreland basin sedimentation
Onwatin
Vermillion
Pyrobitumen
Zn-Cu-Pb
Black mudstone (slate), siltstone, anthraxolite veins
Carbonate, chert, exhalative sulphides
Dowling Member
Onaping
Increasing carbon
fining upwards
Sandcherry
Melt bodies
Columnar jointing
Suevite breccias ("Black Member")
Suevite breccias ("Grey Member")
"Fall-back breccias", primary and reworked, also volcaniclastic deposits due to venting of the still molten melt sheet
Garson
Basal breccias, melted
Sudbury Igneous Complex (SIC)
"Main Mass"
Cooling and crystalization
1850.0±0.9 Ma
Granophyre
Last melt to crystallize
"Crows foot granophyre"
Late aplite dykes
Quartz gabbro (Cpx>Opx, Apt, high MS)
"Main"
Cooling and crystalization
1849.7±0.2 Ma
Norite
Trap dykes
Mafic norite "Sublayer", with mafic and ultramafic incl.
Ni-Cu-Co
Ni-Cu
Partially melted footwall breccia
Cu-Ni-PGEs
Pt-Ag-Au
Shatter cones
SBX
Δδ=1.4
Granophyre
Horizontal
TZG
Floaters
Sinkers
Norite
Depletion
2.6
2.8
3.0
3.2
Density (g/cm3)
45
55
65
75
SiO2 (wt%)
1
5
10
50
100
300
Ni (ppm)
Quartz diorite (QD) with quench textures, early phase, superheated
Melting of wall rock
Inclusion-bearing quartz diorite (IQD), later phase
Fracionated sulphide concentrations in IQD
Stacked Nipissing Diabase sills
2217 Ma
??
Final quartz diorite dykes, plag-phyric, not superheated
Huronian (meta)sedimentary rocks (2.5-2.3 Ga), variably sulphidic
Trap dykes (diabase), ca. 1750 Ma
1848.5±0.8 Ma
"Offset" dykes: multiphase, injected vertically down into footwall fractures
Quartz diorite (QD) dyke, distal, unmineralized
Shatter cones
Pele dyke
Footwall rocks
Matinenda Fm., detrital pyrite
2460 Ma
Bimodal rift volcanic rocks (2460 Ma)
2480 Ma
Early Huronian layered intrusions (East Bull Lake Suite, 2480 Ma))
Ca 2.5 Ga unconformity
~1 km
Archean basement: gneisses and granitoids
WB2022
Basal breccias melted into granophyre
Granophyre
Transition zone gabbro
Differentiated "Main Mass" norite
Basal mafic norite and inclusion-bearing "Sublayer"
"Offset dykes" injected into footwall: quartz diorite and mineralized and inclusion-bearing quartz diorite
Hanging-wall ore
Contact orebodies, variably in-filtrated in footwall breccias
Footwall breccia with partial melting ("Late Granite Breccia")
Footwall orebodies, Cu-rich
Cu-rich massive sulphide sills in footwall
Irregular footwall veins, transition to hydrothermal processes
Metamorphic halo and limit of footwall partial melting
Shatter cones
Pseudotachylite breccia bodies, dykes, and veins ("Sudbury Breccia")

Early publications, at a time when the Grenville orogenic front was already well recognized, attributed the folding of the Sudbury structure to Grenvillian deformation (e.g., Dietz, 1964). More recent mapping and geochronological data suggest, however, that the actual Grenville Front *sensu stricto.* is rather sharply defined and that *circa* 1.0 Ga deformation did not significantly extend far into the immediate foreland and had little effect on the Sudbury structure (e.g., Easton, 1992; Davidson, 1997; Easton et al., 1999). Instead, there is increasing evidence for significant post-Penokean but pre-Grenville deformation, broadly correlated with the Yavapai and Mazatzal (or Labradorian?) orogenic belts/episodes defined in the southwest USA, i.e. broadly in the interval 1.80-1.60 Ga (e.g., Bailey et al., 2004).

However, at a somewhat larger scale, Grenvillian deformation, specifically loading of the crust by the Grenvillian thrust stack, is likely to have contributed regional tilting of the crust to the south, causing further uplift of the Sudbury area and the Archean craton to the north.

The precisely dated Sudbury impact event (Krogh et al., 1982, 1984; Corfu and Lightfoot, 1996; Davis, 2008; Bleeker et al., 2015; Bleeker and Kamo, in prep.), based on multiple high-precision U-Pb ages on units of the SIC, at 1850 Ma, occurred as a sharply (seconds and minutes!) timed event during the *circa* 1860–1840 Ma Penokean orogeny. From a broader regional perspective, the onset of the major Penokean orogenic event at *circa* 1870–1860 Ma predates the impact event (e.g., Holm et al., 2007). In the Sudbury area, this can be demonstrated, perhaps, by Huronian rocks being folded to some degree prior to emplacement of the melt sheet and associated dykes. However, this requires de-convolving the potentially very complex deformation associated with the impact, the immediately post-impact collapse along ring faults, and the large-scale in-flow of material into the transient crater (e.g., see modelling studies on large impact studies and the spectacular deformation, and folding, it induces in the central uplift and the surrounding annulus; Ivanov, 2005). These latter processes would undoubtedly involve tight folding of Huronian strata in places and satisfy the apparent timing relationships.

The most pertinent structures in this respect occur to the north of the preserved SIC, where melt sheet injection dykes (known locally as “offset dykes”, e.g. the Hess offset dyke) appear to cut both limbs of fold structures in synclinal outliers of Huronian strata. Is this apparent folding pre-impact (e.g., as assumed by Mungall and Hanley, 2004, using an outdated model of the Blezardian orogeny), or syn-impact and due to large-scale in-flow of material into the annulus around the rebounding central uplift?

In any case, these structures were tightened with present dips of Huronian units locally being steep. They were thus likely tightened to some degree by Penokean deformation extending into the foreland. This is the main reason the “Penokean front” in Figure 2 is placed to the north of the preserved SIC, and north of these deformed and folded Huronian outliers.

Zooming back out, most of the more intense and final Penokean deformation post-dates the impact structure and also the emplacement of the foreland depositional wedge of the Whitewater Group turbiditic sediments that must have covered the area and which are now preserved only in the keel of the doubly plunging Sudbury Basin syncline (see map of Figure 1). Figure 7 presents a NNW-SSE cross-section through the western half of the Sudbury structure, as constrained by map and outcrop data, as well as a down-plunge projection of the western fold closure of the Sudbury basin (Bleeker et al., 2014).

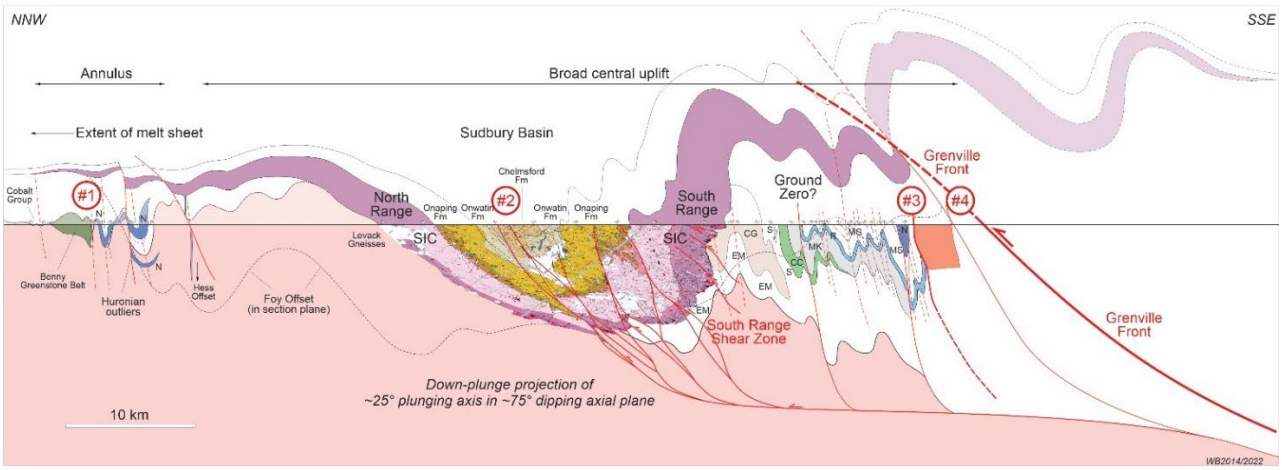


Figure 7: NNW-SSE cross-section through the western half of the Sudbury structure, illustrating the first-order regional scale synclinal fold structure, with more intense deformation and tilting of the southern flank (the "South Range"). This structure is largely Penokean in origin, but, particularly on its southern flank, was further amplified and shortened by younger post-Penokean deformation. Overall shortening on the South Range may reach ~50%, with the basal contact of the SIC being steep to subvertical, and locally overturned. Discrete later shear zones, with south-over-north displacement, further imbricated this steepened southern flank of the structure. However, at the scale of the section, there is no large-scale offset on major discrete shear zones. The folded melt sheet must have extended well to the north, to allow injection of the most distal offset dykes, and also to the south. However, all of the southern half of the folded melt sheet has been removed by erosion. The major structural front in the south is the *circa* 1.0 Grenville Front, with large scale thrust movement. Abbreviations: CG, Creighton Granite; CC, Copper Cliff Rhyolite; EM, Elsie Mt. Formation; MK, McKim greywacke turbidites; MS, Mississagi Formation quartzites; N, Nipissing Diabase sills; R, Ramsay Lake Formation; S, Stobie Formation; SIC, Sudbury Igneous Complex.

This cross-section clearly demonstrates the overall north-verging synclinal structure with the younger than 1850 Ma Chelmsford Formation (*circa* 1840 Ma?) preserved in the core of the syncline. Much of this deformation is likely Penokean, but significantly overprinted by post-Penokean tightening and further shortening on the southern limb of the asymmetric syncline, during deformation associated with the 1.75-1.65 Ga South Range Deformation Zone. The swarm of mafic "Trap dykes", which was emplaced at *circa* 1750 Ma along the South Range (Bleeker et al., 2015; see also the "quartz diabase dykes" of Cochrane, 1984) cuts much of the (Penokean) deformation, but is itself weakly deformed and metamorphosed to upper greenschist facies. It provides an important temporal and structural marker for this post-Penokean interval.

The section demonstrates the intensification of structures and overall shortening on the locally steeply dipping southern flank of the large-scale syncline. Here, bulk shortening may locally reach ~40-50%. Much of this deformation was likely Penokean, but was amplified by younger *circa* 1.80–1.60 Ga deformation that is manifested by a system of southeast-dipping, south-side up reverse shear zones known as the South Range Shear Zone (Shanks and Schwerdtner, 1991; Bailey et al., 2004). Metamorphic titanites in these sheared rocks, which reached epidote amphibolite facies metamorphic grade, date this overprinting deformation as post-Penokean, in the time range of 1.80-1.60 Ga (e.g., Bailey et al., 2004; Papapavlou et al., 2017). Hence, these events have been broadly correlated with Yavapai and Mazatzal orogenic events, as they have in the broader Lake Superior area. Indeed, some of the orebodies on the steep South Range of the SIC are cross-cut and displaced by discrete and well-defined southeast-dipping shear zones. This has been clearly demonstrated in the Thayer Lindsley Mine, and also in the Creighton Mine. This will be discussed during the fieldtrip.

Important as these younger structures are, on the scale of the folded Sudbury structure (i.e. Figure 7), this entire shear zone system only causes minor offsets of the SIC, and the first-order continuity of the SIC at both the western and eastern closures of the folded structure rules out very significant, discrete, fault offsets as has been suggested in some of the published literature (e.g., Milkereit and Green, 1992; Wu et al., 1995; Cowan et al., 1999). The section of Figure 7 is thus best interpreted as a largely Penokean fold structure, moderately overprinted and further shortened by post-Penokean bulk shortening deformation (at the scale of the section), with the final shortening on the South Range reaching perhaps ~50%.

7. Where Was "Ground Zero" and How Big Was the Final Impact Structure?

Given the final deformational state of the very large Sudbury impact structure (Figure 7), a first-order question is: where was the geometrical centre of the impact; or in other words, where was "Ground Zero", relative to the preserved erosional remnants of the structure and those of the melt sheet? An accurate answer to this question has major implications on the interpretation of the size of the impact structure, the volume of impact melt generated, and thus also affects overall mass balance calculations of elements in the melt sheet and the orebodies.

Several important datasets may provide an answer to this all-important question:

1) The analysis of preserved ring structures in the relatively undeformed foreland to the north of the deformed Sudbury structure (Butler, 1994; Spray et al., 2004; see also Grieve et al., 1991).
2) An analysis of the statistical focal point of all well-preserved and least deformed (and possibly somewhat re-oriented) shatter cones (Bleeker, in preparation).

As presented by Butler (1994), a careful analysis of lineaments and possible impact ring structures in the foreland of the Superior craton places the geometrical centre well to the south of Sudbury by as much as 10-15 km. Butler's analysis is supported by a statistical analysis of all intersections of shatter cone axes (*see also* Guy-Bray et al., 1966), which also focus along the South Range, approximately in the Copper Cliff area, or slightly to the northeast in the Frood and Stobie mines area. These findings are summarized in Figure 8.

Superimposed on the diagram of Figure 8 is a model of progressive restoration of the preserved outline of the Sudbury structure (i.e. the base of the SIC), in 4-5 steps, including minor deformation in the foreland, 15-20% shortening in the northern half of the SIC syncline, and up to 50% shortening of the southern half of the structure. This restoration of realistic amounts of shortening places the trace of the South Range, and the original focal point of shatter cone axes, just in range of the Butler's geometrical centre defined by his most robust, lineament-constrained "Ring 3" which is shown in red on Figure 8.

A significant conclusion from this work is that all of the preserved SIC represents only a portion of the northern half of the preserved melt sheet. Hence, there is no *a priori* symmetry between the preserved North Range and the South Range, but rather all kinds of asymmetry: the North Range preserves a thinning northern lobe (but not the edge!) of the original melt sheet, whereas the South Range preserves a more central portion of the melt sheet that was onlapping onto, and partially overlapping the collapsed central uplift. All of the southern part of the melt has been removed by uplift and erosion.

These constraints and conclusions are further summarized in Figure 9. These findings can be modified to some extent by increasing the shortening deformation somewhat, moving the restored South Range of the SIC a bit farther south, but without going to unreasonable shortening estimates it will not change the first-order conclusion that the preserved SIC is all from the northern half of the folded melt sheet and thus preserves a relatively small, fundamentally asymmetric sample of the original melt sheet.

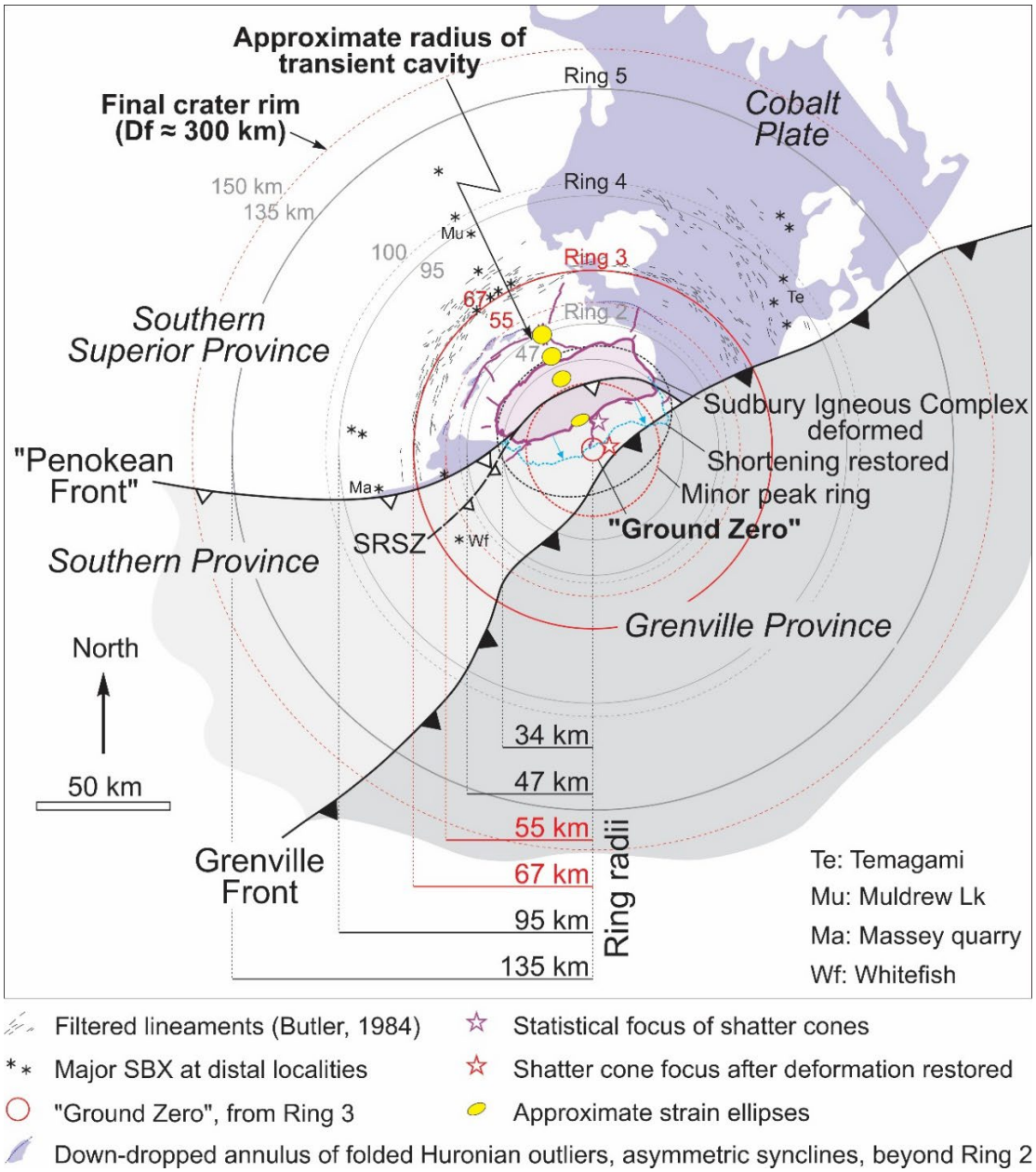


Figure 8: Summary map of the size and location of the preserved Sudbury structure relative to the main structural deformation fronts, and the impact ring structures defined by Butler (1994) based on a detailed lineament analysis (here only lineaments related to Ring 3 are shown for illustration, and some to the northeast of the structure where there is some discordance between lineaments and Ring 4). Of these rings, the one shown in red is the most robust (see Butler, 1994) and it, together with the other rings, defines the centre of the ring structures. This is "Ground Zero" and is shown with the small red circle south of Sudbury. This impact centre is pinned to the foreland as it is largely based on Ring 3. The purple star represents the statistical focal point of best-preserved shatter cone axes and is also located south of the preserved structure. Restoring the shortening deformation in 4-5 progressive steps, including 50% shortening of the southern half, less in the northern half, and ~5% in the foreland up to a radius of 50 km, places the original South Range just in contact with Butler's "Ground Zero" (see the blue trace). Hence, all of the preserved melt sheet represents only part of the northern half of the original melt sheet. Only ~10% of the melt sheet is presently preserved in the erosional remnant of the SIC.

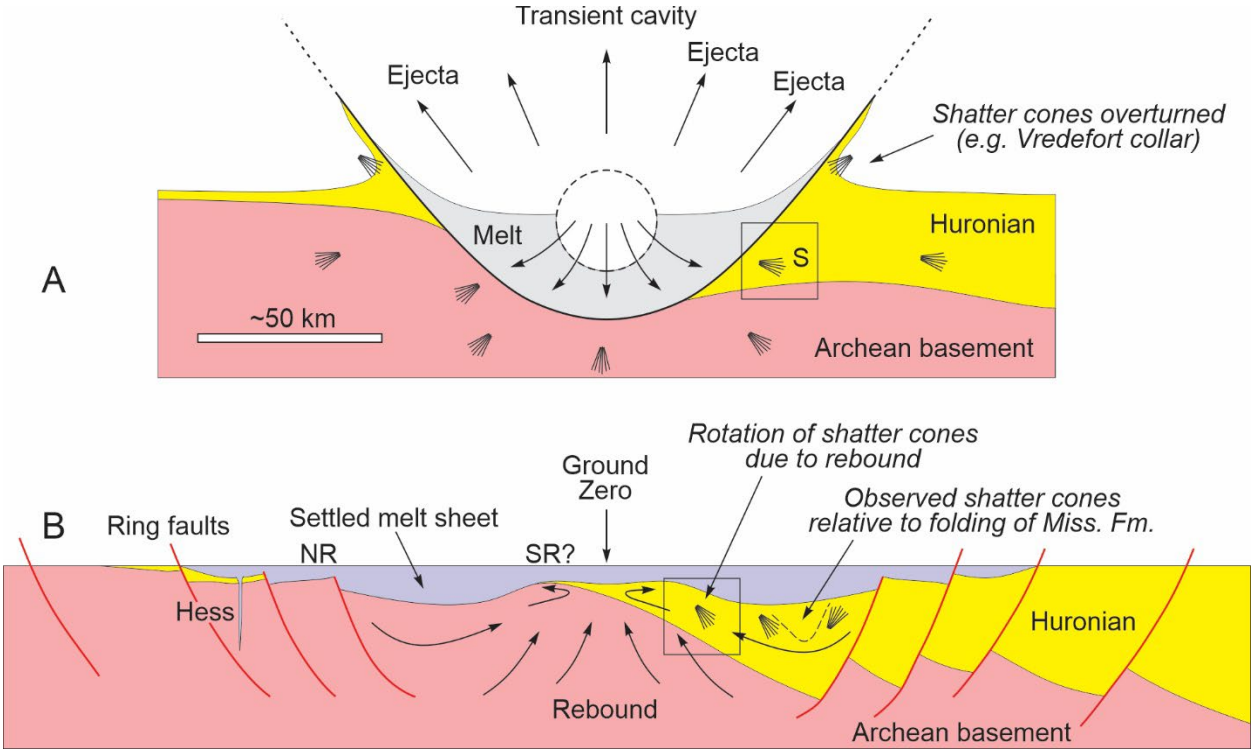


Figure 9: Two critical stages in the formation of the Sudbury impact structure *(*adapted and simplified *after* Grieve, 1994*)*: **A)** early during the impact phase and formation of the transient cavity, as shockwaves are transmitted into the target; and **B)** after collapse and rebound, and final settling of the melt sheet. Note the estimated position of the North Range (NR) and South Range (SR) in figure B. The diagrams also illustrate the general pattern of shatter cones in the target rocks around an impact crater during its formation and excavation (A) and after rebound (B). A shatter cone (S, outlined by square) forms in response to the shockwaves radiating out from the focus of the impact. When the floor of the transient crater rebounds, the cone and the cone and its axis are rotated up and the rocks may also move inwards due to collapse and large-scale material in-flow into the transient crater. In figure B, the concentric Hess Offset dyke is show, at ~50 km distance from "Ground Zero". The Foy Offset dyke, more or less parallel to the section, is not shown but extends out to ~65 km (see Figure 8). The Hess Offset dyke cuts small synclinally folded outliers of Huronian rocks that help to define the down-folded and down-faulted annulus surrounding a broad area of central uplift.

The overall extent of the final melt sheet, after settling, likely reached the well-defined Ring 3 with a radius of ~67 km. This conclusion is supported by at least one of the injection dykes, the Foy Offset, reaching close to this Ring 3, although the originally overlying melt sheet from which it was injected down has been removed by post-Penokean uplift and erosion. Thus, with an estimated radius of 67 km, and an average thickness of the melt sheet of 2.5 km (*see also* Grieve et al., 1991 and Naldrett and Hewins, 1984: 2–3 km on the preserved North Range, 3–5 km on the South Range), this suggests a final melt sheet volume ($\pi r^2 \, x \, d$) of ~35 x10^3 km^3. This estimate does not include the considerable melt component preserved in the ~2 km-thick Onaping Formation or within the ejecta that were blown far beyond the crater (tektites?) and into space. Thus, it is likely an underestimate. Nevertheless, reasonable uncertainty estimates put this total impact melt volume in the range of 25–50 x10^3 km^3, which could be used as an input parameter into various scaling models and models of elemental mass balances.

Of this original volume of settled impact melt sheet, the thin edge of which likely extended out to ~67–70 km radius, only ~2.5–5.0 x10^3 km^3 is preserved in the current extent of the SIC. This estimate can be derived by measuring the cross-sectional area of a central cross-section through the SIC (similar to Figure 7 but a bit farther east), multiplied by 0.5x the preserved width (long axis) of ~60 km.

In other words, only on the order of 10% of the original melt sheet is represented by the currently preserved erosional remnant of the SIC, and all of this represents the proximal part of the northwestern half or lobe of the original melt sheet (see Figure 9), perhaps including a thicker "moat" north of the collapsed central uplift.

As deformational structures, both large and small, typically nucleate on and amplify original "seed structures" or weaknesses, the curved main synclinal axis of the preserved SIC may be inherited to some degree from such a moat surrounding the core of the central uplift. This is where the melt sheet may be thickest and even larger orebodies could have collected. All of these key points should be considered in any interpretation of the Sudbury structure and its endowment of magmatic sulphide orebodies.

Perhaps one of the more robust features of the final impact structure, as presently preserved in the foreland, and in relation to overall size estimates, is the down-folded and faulted annulus of Huronian outliers to the north of the SIC, with a pattern of synclinal fold traces that curves around the entire northern half of the structure, including around the western and eastern first-order fold closures of the structure (Figure 8). This down-folded annulus fits a ring structure with a radius of ~50–55 km. The annulus thus has an apparent diameter of 100-110 km, which is larger than the well-defined annulus of the *circa* 2023 Ma Vredefort impact structure in South Africa, where it is ~90 km in diameter. This can be scaled to a final crater diameter.

The broad, collapsed central uplift must fit within this annulus and therefore has a total radius and diameter of ~45-47 km, and ~90–100 km, respectively, and includes all of the South Range area and the deeply exhumed Huronian to the south, and also the Levack Gneisses in the north. In large complex terrestrial impact craters the diameter of the broad central uplift area is roughly 1/3 of the final crater diameter (Therriault et al., 1997), the scaling relationship being:

$$D_{cu} = 0.31\ D_f^{1.02}$$

where D_{cu} is the diameter of the central uplift, and D_f is the diameter of the final crater rim (Therriault et al., 1997). This would suggest that the final Sudbury structure may have had a diameter closer to 300 km and somewhat bigger than most estimates. Based on the estimate of the annulus diameter alone, it is clear that Sudbury is the largest among known terrestrial impact structures, just slightly larger than the more deeply eroded (no melt sheet preserved) Vredefort structure. Major Sudbury Breccia pseudotachylite occurrences reaching out to a radius of ~120 km supports such a larger estimate based on the scaling relationship of Stöffler et al. (1988):

$$D_{pst}\ .\ 0.8\ D_f$$

Where D_{pst} is the diameter of significant pseudotachylite breccia occurrences, i.e. ~240 km. This would result in a similar estimate of ~300 km for the final crater diameter (D_f).

8. Conclusions

The Sudbury structure represents the deformed erosional remnants of a very large 1850.0±0.9 Ma meteorite (or comet?) impact crater. The current consensus is that the final collapsed crater reached ~200–260 km (final crater diameter, D_f), but it may have been larger based on a robust size estimate of the down-folded and faulted circular annulus, and the most distal observations of significant pseudotachylite occurrences (Sudbury Breccia), some of which occur well beyond the 100 km-radius ring structure (Ring 4 of Butler (1994; *see also* Thompson and Spray, 1994)). We thus estimate the final crater diameter at ~300 km, which makes it the largest known terrestrial impact crater.

The volume of impact melt generated was ~35 x10^3 km^3. Once the crater had collapsed and rebounded, and the melt sheet had settled across a complex peak-ring crater, on a time-scale of mere hours (!) (e.g., *see* Grieve, 1994), the sheet of impact melt reached out to ~67–70 km, and had an average thickness of ~2.5 km. It may locally have reached 5 km or more (Figure 10).

A large proportion of this melt sheet was initially superheated and underwent rapid homogenization and differentiation. Zieg and Marsh (2005; *see also* Golightly, 1994) point out some of the very complex processes involved as the final melting front burned into the footwall and different blobs of melt or partial melt were generated and may not have fully mixed and homogenized by turbulent convection. There likely was an early separation in i) less dense, more felsic melts, and ii) denser mafic melts, the first floating to the top of the melt sheet pool to form or contribute to the Granophyre of the Main Mass of the SIC, whereas the latter collected towards the base to form the Norite, and Sublayer. The Sublayer represents a complex boundary layer with remnant mafic and ultramafic fragments that collected along the base of the SIC, together with a rain-out of sulphides globules. The Transition Zone Gabbro represent the differentiated top of the lower mafic section, enriched in incompatible elements.

The density contrast between the Granophyre and Norite (including the gabbro) is significant, and large enough such that the basal granophyre contact would have equilibrated in a near horizontal position, providing an important reference plane for evaluating the effects of later structural deformation. One of the important implications of this is that it allows a semi-quantitative reconstruction of the footwall topography of the SIC (Figure 10) and aids in general structural reconstructions (e.g., Figure 7).

Some of the enormous volume of ejecta fell back into the crater, or was washed back in by various processes and formed the Onaping Formation. In a general sense, the upwards fining stratigraphy of the Onaping Formation and increasing carbon content reflect the waning energy levels and somewhat longer time scales of deposition, with the uppermost and finer-grained "Black Member" material transitioning into post-impact sedimentary and volcano-exhalative processes of the Vermillion member.

The impact struck at a target site with complex geology, on the boundary of the rifted Superior craton and the developing Penokean fold belt of the Southern Province. The rebounded and settled crater, and the crater-fill deposits, were then covered by the expanding wedge of foreland basin sediments, first deeper water mud- and siltstones (foredeep?) and finally the greywacke turbidites of Chelmsford Formation at *circa* 1850–1840 Ma. Soon after, the crater, the melt sheet, and overlying deposits were deformed and shortened by the climax of Penokean deformation, which transformed the area into a doubly plunging regional syncline. The ~3-5 km-thick SIC melt sheet formed a very thick competent layer during this deformation, which resulted in a single large wavelength open fold structure. This syncline, particularly its southern limb, was further modified and shortened by post-Penokean deformation associated with the *circa* 1.80-1.60 Ga South Range Shear Zone.

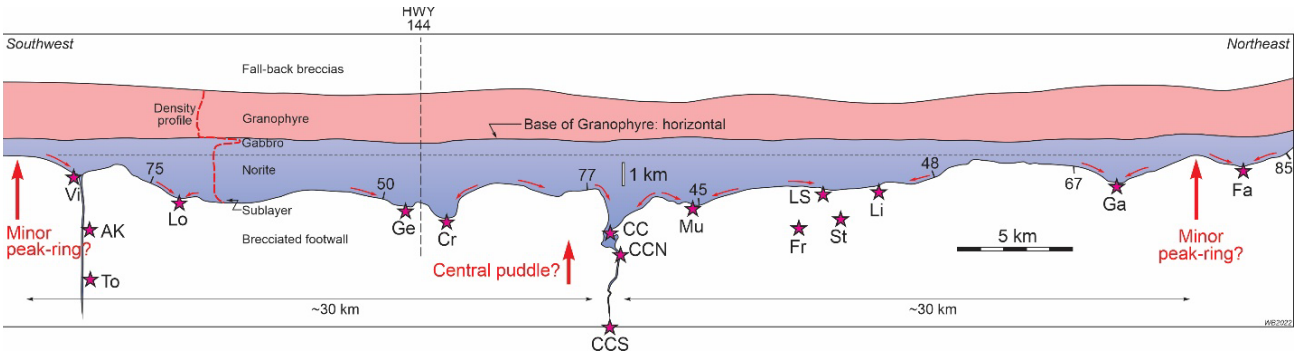


Figure 10: A simplified, first-order reconstruction of the South Range melt sheet with its inherent thickness variations and footwall topography. It highlights the embayment structures where most of the significant orebodies collected (*see* Pye et al., 1984, and contributions therein), as well as the lateral thickness variations which suggest a second-order "peak-ring" on the overall central uplift, where the mafic section of the melt sheet (norites plus gabbro) is much thinner than in the "central puddle" that approximately overlapped "Ground Zero". The dashed horizontal line indicates the level of the topographic highs of the secondary peak-ring. The section was constructed by first restoring minor second-order folding along strike of the South Range, then correcting from apparent thicknesses to true thicknesses, and finally hanging the section segments from the paleo-horizontal reference plane, i.e. the base of the significantly less dense granophyre upper section of the melt sheet. From west to east, star symbols identify the major magmatic sulphide deposits and producers: Vi, Victoria mine; AK, Aer-Kidd; To, Totten; Lo, Lockerby; Ge, Gertrude; Cr, Creighton; CC, Copper Cliff; CCN and CCS, Copper Cliff North and South mines; Mu, Murray; LS, Little Stobie; Fr, Frood; St, Stobie; Li, Thayer Lindsley; Ga, Garson; and Fa, Falconbridge. Of these Creighton and the Copper Cliff system are among the largest deposits known, and associated with the deepest embayments or funnels, in the floor of what appears to be a deep central puddle in the melt sheet, approximately overlying the centre of the collapsed central uplift.

Following various stages of regional uplift and erosion, the end result is that only ~10% to 20% of the original melt sheet is preserved in the current erosional remnant of the SIC. Both the analysis of lineaments and ring structures, and the analysis of shatter cones, suggest that the preserved, folded melt sheet (the SIC) represent parts of the northern half of the original melt sheet and associated crater, with "Ground Zero" being situated south of the main footwall contact of the South Range.

This is significant as most studies have implicitly assumed that "Ground Zero" was underneath the preserved SIC and that the preserved SIC (e.g., Golightly, 1994) more or less preserved a symmetrical patch of the original melt sheet and impact structure centered on "Ground Zero". If the ring structure and shatter cone analysis is correct, this is clearly not the case and there is no inherent basis for symmetry between the North and South Ranges. Rather, the North and South Ranges represent fundamentally different parts of the melt sheet and the collapsed crater; the former represents a more distal northern part of the structure, and the latter a thick, more central part of the melt sheet lapping onto and overlying the rebounded and collapsed central uplift of the final peak-ring crater (see Figure 9).

Petrologists had long recognized that something was odd about the Sudbury structure and its igneous rocks, the "Sudbury Irruptive". In contrast with other large mafic igneous complexes, there was no layered lower section of mafic and ultramafic cumulates near the base, and way too much granitic granophyre near the top (~40% of the volume, rather than ~10% in a fully differentiated mafic intrusive complex). And the

mafic lower part had an unusual SiO_2 content (~55–58 wt%), dominantly characterized by norites rather than gabbros[4]. The regional deformation was also puzzling and under-appreciated, feeding into interpretations of the whole complex as a concave-upwards igneous lopolith.

It took a marine geologist from California, who somehow had solved the mystery of shatter cones in a small number of suspected impact craters in the 1940–1950s (e.g., Dietz, 1947, 1959) [5], and who also was reflecting on the lessons learned from observing the lunar surface (Dietz, 1946), to come and check out Sudbury, and within days confirm his suspicion that it was, fundamentally, a large impact structure with major shock damage in the footwall and no damage in the melt rocks (the igneous rocks of the SIC). Being less familiar with the ores, he still chose to hedge his bets on some of the details, such as the origin of the ores, or the exact nature of the igneous rocks, settling on a hybrid model of a large impact crater that was then intruded by the igneous rocks of the irruptive.

As late as 1970, the debate on Sudbury was summarized as follows in a major paper on the structure (*see* Naldrett et al., 1970):

> "There are two main theories of origin for the Sudbury structure. These are very different from one another and hinge on the interpretation of the Onaping formation and certain unusual pre-irruptive breccia dikes.
>
> Speers (1956, 1957), pointed out that the Sudbury Irruptive lies at the apex of a broad dome some sixty miles in diameter involving Huronian and older rocks. He postulated that uplift of the dome occurred in response to pressure exerted by igneous magma. Successive episodes of uplift, followed by tensional release, gave rise to the breccia dikes and finally resulted in caldera collapse at the apex of the dome. Magma escaping around the rim of the caldera and flowing into the center of collapse produced the Onaping formation. The Irruptive was intruded subsequently, spreading out along the base of this formation. According to this hypothesis, the Nickel Irruptive is a later plutonic manifestation of the igneous activity which previously had given rise to the extrusive Onaping formation.
>
> In opposition to this hypothesis, Dietz (1964), suggested that the circularity and brecciation characteristic of the Sudbury structure could best be explained by an explosive meteorite impact, an interpretation for which his own discovery of shatter cones gave support. This theory led French (1967) to find, in inclusions in the Onaping formation, microscopic features characteristic of shock metamorphism.
>
> Subsequent work has shown that shock metamorphism, a typical feature of impact sites but unknown in volcanic rocks (French and Short, 1968), is widespread and common in the Onaping formation; it also is found in footwall rocks adjacent to the Irruptive and in fragments in Sudbury breccia. According to the meteorite impact theory the sequence of events was as follows: shock waves radiating from the point of impact produced brecciation, melting, microscopic shock features and shatter cones, and excavated a circular crater; part of the material blasted from this crater fell back as a poorly, sorted

[4] See, for instance, the papers by Wilson (1956) and Hamilton (1960), written and published just prior to Dietz' shatter cone revolution, to appreciate the conceptual struggles that petrologist and geologist were dealing with to explain major aspects of the Sudbury structure. In his paper, Hamilton is inching towards an essentially extrusive interpretation for the Sudbury lopolith.

[5] See the amazing review paper by Bourgeois and Koppes (1998) to better understand the historical development of these ideas, and all the players involved, including of course the life and career of Dietz himself. Originally from New Jersey, he was broadly educated with degrees from the University of Illinois and having spent time at the Scripps Institution for Oceanography. He was also well-travelled. A professional posting in Europe during the 1950s had also allowed him to visit the Steinheim and Ries basins in Germany.

breccia—the Onaping formation; fracturing and heating of the rocks, and reduction of pressure in the upper mantle below the crater, triggered the evolution of the Nickel Irruptive; the magma was emplaced in the breccia zone beneath the center of the structure." [End of quote.]

It would take another 15–20 years for clarity to emerge. Once the isotopic evidence became available showing that essentially all the igneous rocks, including the more mafic ones, have bulk Nd isotopic signatures that reflect melting of the crust, rather than melting of the mantle, the pendulum finally swung to an impact-only interpretation (Stöffler et al., 1989; Grieve et al., 1991). Bulk sample isotopic values and mixing equations may still hide a very small mantle component into some of the melts but, to date, no conclusive evidence for this has emerged.

What lessons can be learned from all of this? One is that it is critical to think "big", always, and broaden one's horizon, and to to reflect on new ideas from related or not-so-related sciences. The other is that, as our understanding of the impact record grows, particularly on nearby planetary surfaces, there got to be other Sudbury's out there, with perhaps less than ~10% preservation: just the odd bit of breccia; or a poorly preserved shatter cone here or there; or some odd dyke of quartz diorite with some sulphides in it, and which was just a little too hot for your average diabase dyke!

Remember, every breccia is an interesting breccia!

Acknowledgements

The first author (WB) was first introduced to some aspects of Sudbury geology in 1987, during a fall fieldtrip of the Canadian Tectonics Group. He has worked, on and off, on Sudbury geology since the early 1990s, first as a researcher for Falconbridge Ltd. at Onaping Mine, and later for the Geological Survey of Canada. The second author (SK) has been involved in precise U-Pb dating of many of the rocks in and around Sudbury. The geology of this unique area never stops to fascinate. WB would like to thank co-leaders for helping to put this fieldtrip together, and also those who gave the guidebook a proof-read to catch some of the imperfections.

Excursion Stops, Day 1:

Stop 1: Well-developed shatter cones in Mississagi Formation quartzites, south of Sudbury

46.432448° N, 81.072267° W
494448E 5142100N

This is one of the classic shatter cone localities south of the Sudbury structure, and probably one seen by Dietz early on during his 1962 fieldtrip to Sudbury (Figure 11). The conical, striated fracture surfaces can be seen both in the blasted roadside outcrop on the south side of the gravel road (Gibson Road), but also on top on natural outcrop surfaces. Typically, shatter cones are more obvious in blasted outcrops where the rock surface has been opened up, and are more easily missed on natural surfaces. We will see examples of that on this fieldtrip, but this is not a limitation here. Spectacular shatter cones are visible south of the road, on top of the outcrop.

Bedding at this locality dips and youngs to the southeast. The shatter cones are well developed and enough of the cone surfaces are visible, such that their axes can be measured accurately. These axes point up towards the north-northwest (~340º/+45º, i.e. up), and they plunge up more or less parallel to the southeast dipping bedding surface, perhaps just a bit shallower than the bedding dip angle (~45–55º). Although there is plenty of impact brecciation in these footwall rocks, locally chaotic, these outcrops are not chaotic and align with the regional pattern. The shatter cone axes therefore point approximately to "Ground Zero", the centre of the impact structure.

In meteorite impacts, the shockwaves travel out from the centre of the transient crater, outwards, and interaction of the high-velocity shock waves with imperfections in the rocks nucleate the conical fracture surfaces. The apparent point of origin of the shockwaves is below ground, as the crater is being excavated and the ground is depressed. Therefore, away from "Ground Zero", shatter

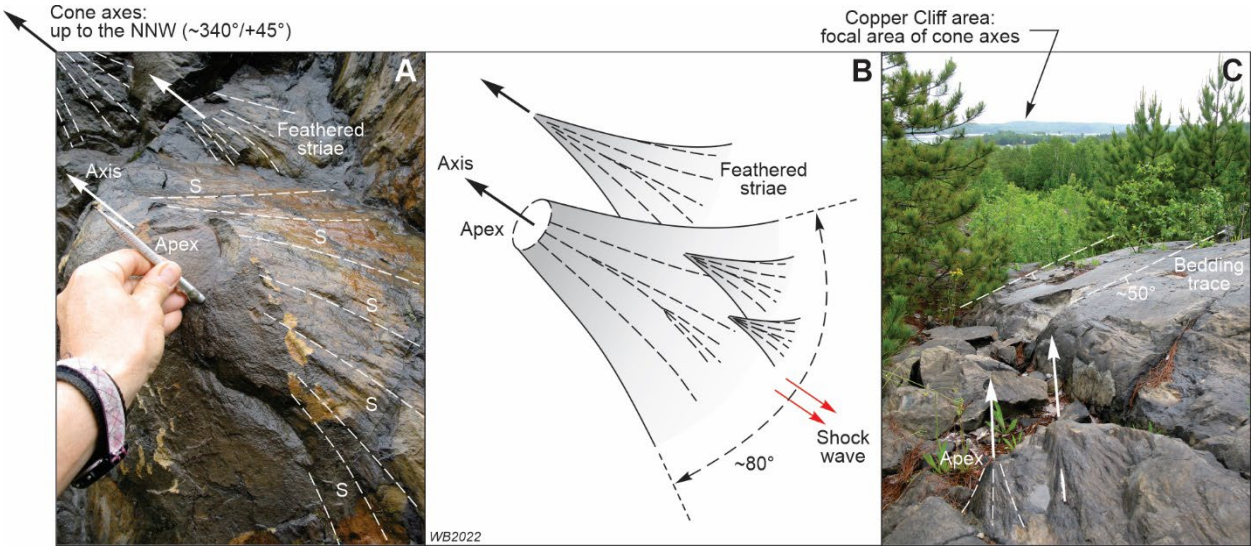


Figure 11: Classical "shatter cones" in the shocked footwall and target rocks of the 1850 Ma Sudbury impact crater, well developed in quartzites of the circa 2.4 Ga Mississagi Formation. With the newly recognized knowledge, in the mid- to late-1950s, that these conical, radiating fracture surfaces represent unique "trace fossils" for high-velocity, very high pressure shock waves associated with meteorite impacts (Dietz, 1959), Sudbury was quickly recognized as an astrobleme—the scar of an ancient impact crater (1962; see Dietz, 1964; and Dietz and Butler, 1964).

cones will fan outward and their axes will be shallow. On rebound of the crater floor, the central uplift moves up and rotates the cones, with their axes now pointing up (see Figure 9). If marginal rocks are turned up or even flipped over during crater formation, the cones may point outwards in rocks that are flipped over (as in the collar around the Vredefort central uplift). Here this is not the case, the rocks are right-way up, dipping moderately to the southeast, and the cones plunge up to the north-northwest.

Hence, just from this outcrop alone, it seems "Ground Zero" was to the north, and their dip was subhorizontal when the shockwaves hit. The cones can be examined on top and in section, to determine their axes, as will be demonstrated during the visit.

Stop 2: Well-developed shatter cones in blasted Mississagi Formation quartzites, south of Sudbury

46.422030° N, 81.086392° W
493367E, 5140943N

This roadside outcrop is again in Mississagi quartzites, somewhat to the southwest of the previous locality. Here beds dip and young towards the north, so we have travelled across one of the many folds in the Mississagi Formation. Several reasonably developed shatter cones are visible in the rocks on the east side of the road (Figure 12). The cone axes here plunge down and to the north, in approximately identical relative orientation to the bedding surface, but now plunging down!

Just from these two outcrops alone, it is clear that there is a correlation between final bedding attitude and final attitude of the cones: both have been affected by moderately tight folding that produced the local synclines and anticlines in the Mississagi Formation. Unfolding of these folds and restoring bedding to approximately horizontal will also align the cones between Stop 1 and Stop 2. This general story is repeated all through the area and the conclusion must therefore be:

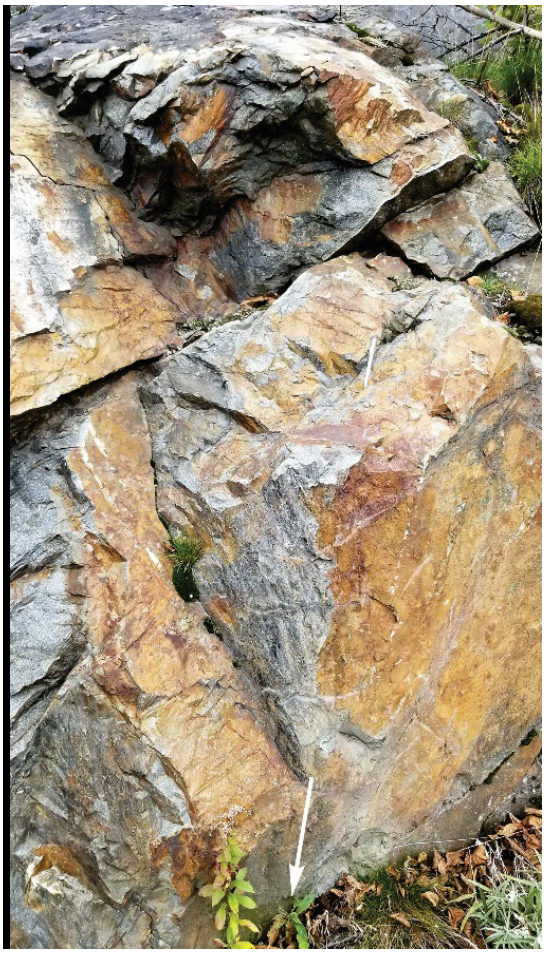


Figure 12: Shatter cones plunging down to the north-northwest, on the southern limb of the local syncline. Both bedding and the shatter cone axes have been re-oriented by the Penokean folding.

1) The local Huronian strata were (sub)horizontal at the time of impact;
3) The shockwave traveled from north to south and formed more or less flat lying shatter cones with their apices plunging gently to the north;
4) And both bedding and cones were affected by the folding that affected the south range of the Sudbury structure.

Duplicating this exercise at as many as possible localities, and then intersecting all cone axes statistically, looking for a maximum of intersections, can define the origin of the shockwaves, i.e. "Ground Zero", or perhaps a point below "Ground Zero".

This outcrop also demonstrates that shatter cones are rare on naturally weathered outcrop surfaces but are more clearly developed or visible on certain broken rock surfaces.

Stop 3: Southern extent of the Copper Cliff "Offset dyke" cutting across Pecors and Mississagi Formation strata

46.434569° N, 81.068932° W
494704E, 5142335N

The sedimentary rocks on the south side of this Gibson Road locality represent the transition from the finer-grained Pecors Formation to the quartzites (quartz arenites) of the Mississagi Formation. Dips and younging direction are to the south, as at Stop 1, which is just along strike. The Huronian strata are cut by a ~25 m-wide mafic dyke, which at first sight looks not unlike a diabase dyke. This dyke is subvertical and trends south, and its contact relationships look rather typical for a diabase dyke, wandering a little bit and stepping sideways a little bit here and there. There is, however, no known swarm of this trend and/or the right age to explain this dyke.

Petrography shows it to be a medium-grained quartz diorite, with some chilling near the margin. A quartz diorite is rather atypical for a regional diabase swarm, which are essentially all basaltic. It is perfectly along strike of the Copper Cliff Offset dyke to the north, on the north side of Kelly Lake, which can be mapped into the base of the Sudbury Igneous Complex (SIC).

It is tempting to think of these "radial offset dykes" as having been emplaced laterally from the north, but that is not the right interpretation. Almost certainly, these dykes of basal melt rock from the SIC were injected downwards from the overlying melt sheet into active fracture planes in the deforming footwall, during the overall crater modification processes and settling and early differentiation of the melt sheet. Overall homogeneity of these quartz diorite dykes indicates that the melt sheet had already mixed and homogenized to some extent. The overall time scale of this dyke emplacement, relative to "time zero" could be years to thousand of years?

Here the dyke does not show evidence for superheating and melting of adjacent country rocks, but elsewhere melting can be observed along the contact.

After emplacement and cooling, the dyke was moderately deformed together with the country rocks during the largely Penokean shortening and folding deformation.

Stop 4: Unsorted matrix-supported conglomerate (reworked diamictite) of the Ramsay Lake Formation

46.433508° N, 81.077564° W
4947041E, 5142218N

This glacially polished outcrop is on the north side of the Highway 17 Bypass just south of Kelly Lake. Note that this is a very busy highway with heavy, high-speed traffic. The shoulder of the highway allows cars to stop safely here but make sure to pull well on to the shoulder.

This is an outcrop of poorly sorted, matrix-supported, sandy conglomerate typical for the lowermost of the three glacial formations in the Huronian succession. The Ramsay Lake Formation is the basal unit of the Hough Lake Group (see Figure 4), which overlies the rift and rift-fill succession of the Elliot Lake Group, locally with a sharp contact.

This outcrop shows the unsorted nature of the conglomerate and on an outcrop like this one could debate the evidence for a glacial origin. Some bedding surfaces and cross-bedding are visible. Long (2009) describes these rocks as sub-glacial melt-out till. Elsewhere the Ramsay Lake Formation, which typically is thick to very thickly bedded, or even massive, has a finer-grained, darker matrix and looks more like a typical diamictite (Figure 13). Here we are at the top of the Formation, where the till material was reworked and sorted to some degree by sub-glacial processes.



Figure 13: Unsorted more typical diamictite of the Ramsay Lake Formation, with a dark fine-grained matrix (not this outcrop). Pebbles and cobbles are mostly granitoid rocks. Dark patches with striae are a rare example of how shatter cones are exposed on naturally weathered (and polished by Pleistocene ice movement) surfaces.

Stop 5: Mineralized Copper Cliff Offset dyke, up from the walking trail at Copper Cliff

46.470977° N, 81.075562° W
494199E, 5146381N

This series of outcrops on the side of a hill overlooking the urban area of Copper Cliff exposes parts of the major Copper Cliff Offset dyke which can be mapped to the north into a major "funnel" or narrow embayment structure at the base of the SIC. Major Cu-Ni sulphide mineralization occurs in this funnel structure and into the dyke (Figures 14 and 15).

Walking up toward the dyke, one walks across heavily brecciated wackes and arenites at the base of the McKim Formation, and into the top of the Copper Cliff Rhyolite Formation. Bedding features at the base of the McKim Formation, such as graded bedding, scours, and truncated bedding indicate younging is towards the southeast in steeply southeast dipping strata, i.e. stratified sediments that overlie the rhyolites to the north. The latter show quartz phenocrysts and beautiful flow lamination in places.

The brecciation, locally approaching large-scale chaos, is typical for many areas in the footwall. Dykes and veins, and more irregular domains of fine-grained, dark pseudotachylitic rocks, are what is referred to as "Sudbury Breccia", indicating intense brecciation and pulverization of the target rocks during impact and intense deformation associated with rebound and collapse of the central uplift.

All these rocks are cross-cut by an irregular dyke-like body of quartz diorite emanating from the Copper Cliff funnel. The outer contact of this quartz diorite dyke induced melting and mixed with the adjacent rhyolite (Figure 16). The quartz diorite itself shows typical quench textures: more or less spheroidal structures of radiating, fine acicular pyroxene/amphibole crystals. Together, these observations show that this quartz diorite was injected in a superheated state and then quickly cooled due to interaction with wall rocks. Locally the quartz diorite dyke rock interacted with dark Sudbury Breccia pseudotachylitic material, overall showing that the latter was marginally older but not yet acting fully lithified or brittle.

The outer phase of the dyke is known as typical quartz diorite, or "QD", and is generally similar to quartz diorite of many of the offset dykes (e.g., see Stop 3). It generally does not contain sulphides, and does not weather rusty. This outer QD was intruded by one or more phases of dyke injections that are characterized by carrying along inclusions of varying size (inclusion-bearing quartz diorite, or "IQD"), many of which are more mafic, together with variable amounts of sulphides, either disseminated or as conspicuous cm-size globules. A central phase of the dyke carries semi-massive sulphides, which are enriched in Cu. Hence, in these outcrops here, three successive phases of dyke injection can be demonstrated, which mark different stages in the evolution of the melt sheet:

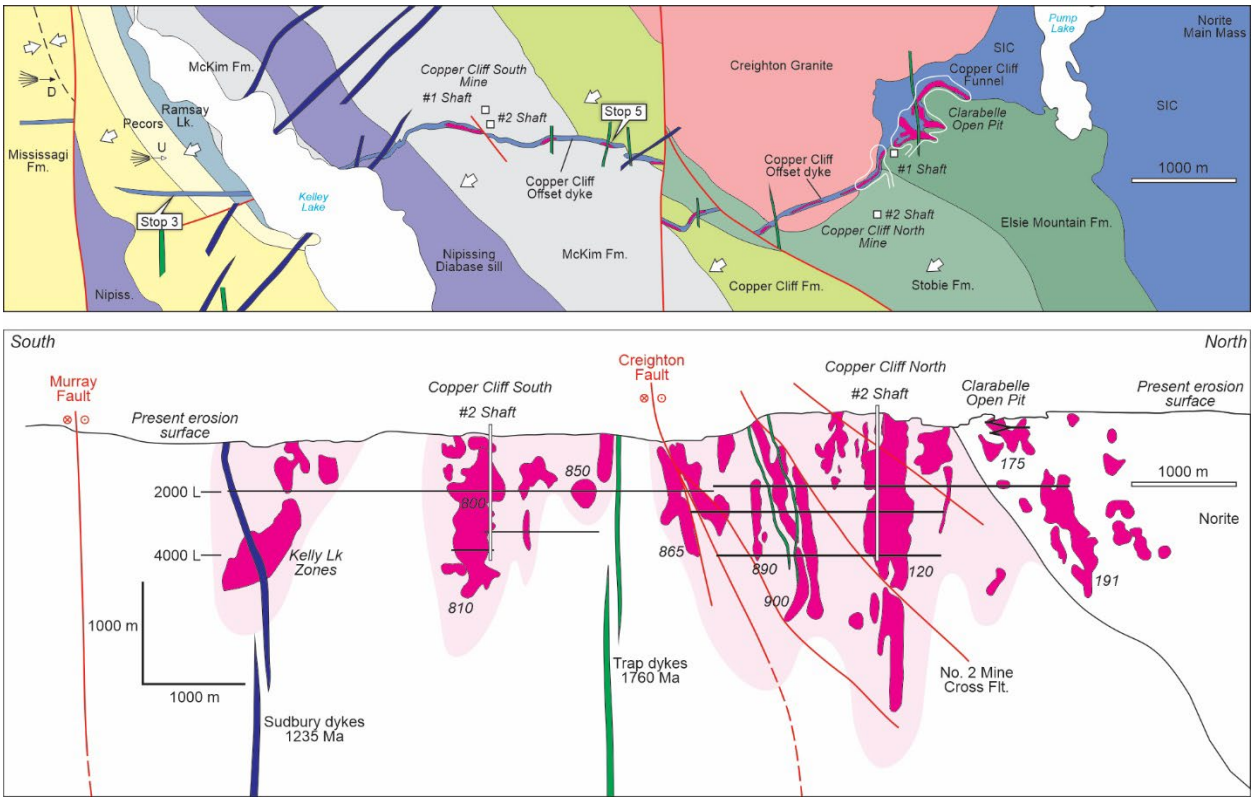


Figure 14: Map and corresponding longitudinal section, looking west, of the Copper Cliff Offset dyke and its associated orebodies, supporting several active mines. North is to the right in the map figure. Modified after Cochrane (1984) and Farrow and Lightfoot (2002), based on mine and exploration sections.

1) Initial homogenization and injection of unmineralized quartz diorite, probably just prior to sulphide saturation having taken place.
2) Renewed injection from near the base of the melt sheet, with abundant inclusions and sulphide globules being entrained.
3) A final phase of injection, after sulphides had collected and had become enriched in Cu due to Fe-rich and Cu-poor monosulphide solid solution (MSS) having separated out, enriching residual sulphide liquid in Cu and other MSS-incompatible elements (e.g., Craig and Kullerud, 1969).

Elsewhere, a fourth and final phase of injection can be recognized, consisting of plagioclase-phyric, sulphide-free quartz diorite, representing differentiated norite from the main melt sheet, making it into the footwall.

On Figure 14, the mineralized bodies are shown in magenta along the extent of the offset dyke (after Cochrane, 1984). In the corresponding longitudinal section, looking west, the overall extent of the sulphide bodies in the dyke are shown, forming km-scale steeply plunging "fingers", generally along the centre of the dyke, but sometimes along the margin. These steeply plunging fingers of mineralized IQD clearly indicate emplacement was downwards from the base of the overlying SIC (now eroded away) where sulphides had collected.

The vertical plunge of these IQD "dyke in dyke" injections was further amplified by N-S shortening during folding, and vertical extension, but deformation intensity is insufficient to explain the observed aspect ratios. So, the long axis of the orebodies, well defined by drilling and mining, indicates the injection direction: i.e., down from an overlying but now removed melt sheet undergoing critical stages in magmatic evolution:

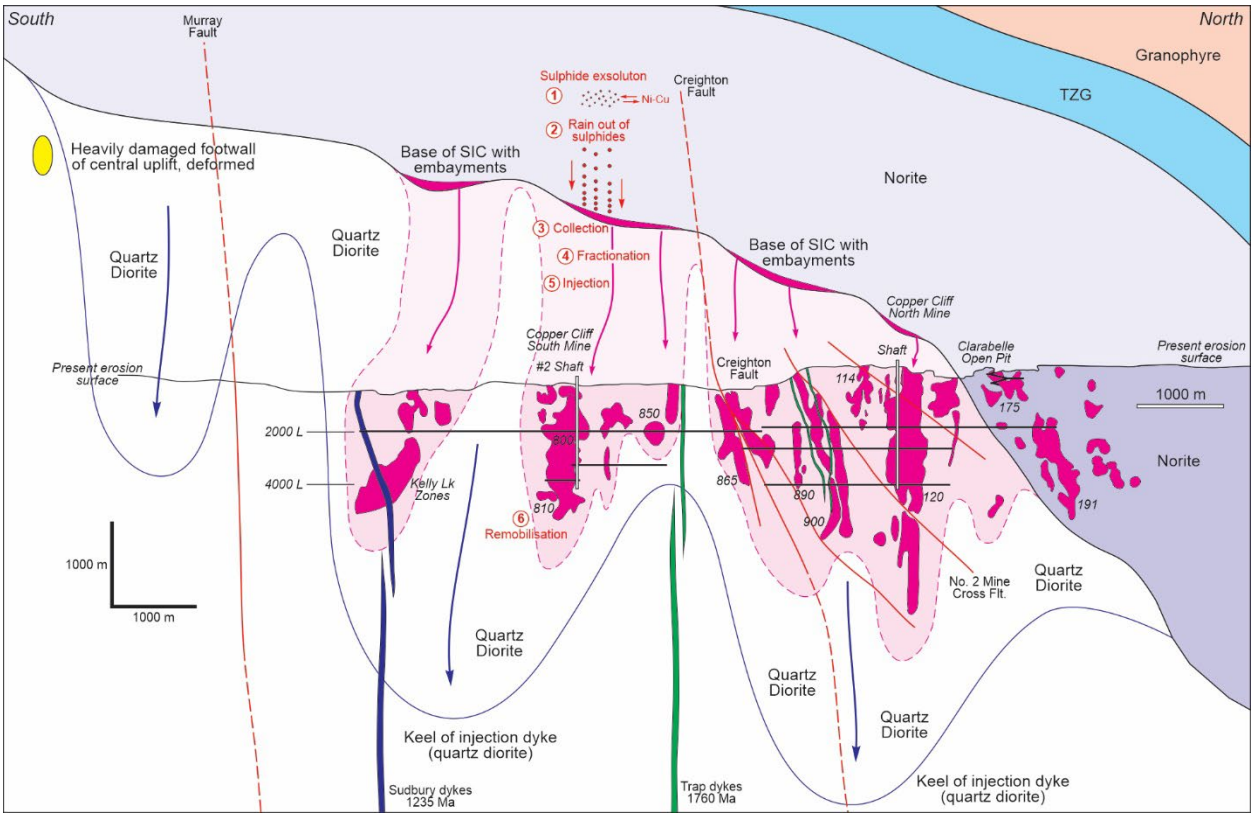


Figure 15: Longitudinal section similar to but expanded from Figure 14, with full interpretation. Note the full extend of the Copper Cliff Offset dyke with an initial "leading sheet" of relatively homogeneous quartz diorite, followed by later injections of mineralized inclusion-bearing quartz diorite and entrained sulphides, all down from the overlying but now eroded Sudbury melt sheet, which underwent final differentiation into a thick basal norite section (~3 km), a thin transitional quartz gabbro section (~500 m, TZG in blue), and an upper "granophyre" section of broadly granitic composition (~1–2 km, pink). As shown on the figure, the overall ore formation process can be divided into six stages: 1) sulphide saturation, 2) growth and sinking of sulphide globules, 3) collection of sulphides along the basal contact, particularly in topographic low or "embayments", 4) the onset of liquid fractionation of the sulphide melt due to MSS fractional crystallization, and progressive enrichment of the residual sulphide melt in Cu, 5) episodic injection of dense melts into footwall fractures, and 6) further sub-solidus remobilization of ore components during later deformation. The dip of the SIC is due to subsequent deformation.

1) Initial homogenization due to rapid convection in a superheated impact melt;
2) Rapid sulphide saturation;
3) Collection of basal sulphides and entrained inclusions in a still very hot magma;
4) And repeated injection of basal phases of the evolving melt sheet into footwall fractures forced open by i) on-going movements in the adjusting footwall, and ii) fluid pressure of the magma.

The density of the overall "dioritic" (more or less, average crust, well mixed) is about 2.8 g/cm3 and is denser than the average density of the footwall rocks.

Figure 15 above shows the completed section and interpretation. In many ways this one section tells much of the story of the Sudbury structure, its evolving melt sheet and its orebodies.

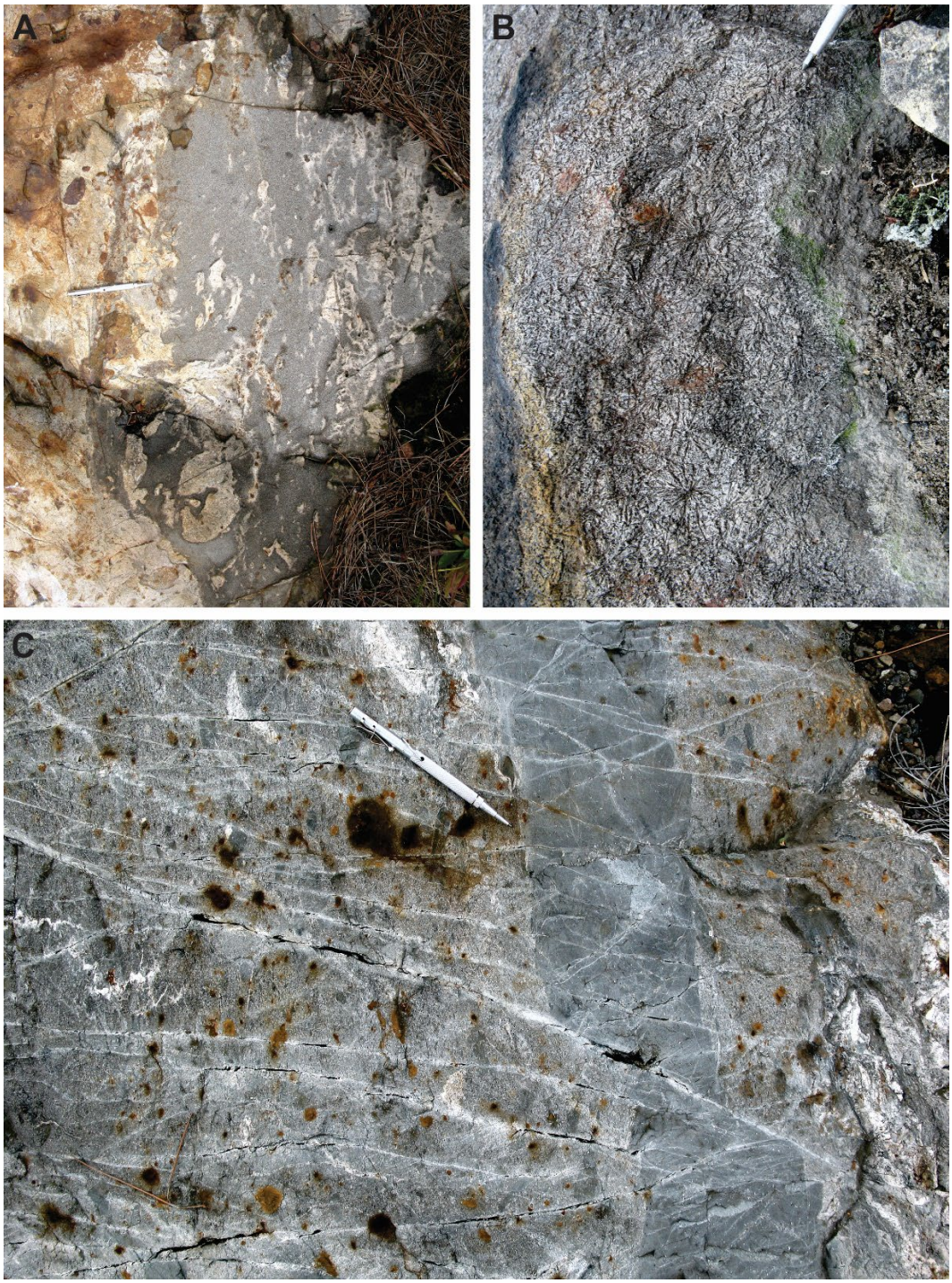


Figure 16: Some key features associated with the quartz diorite offset dykes. **A)** Margin of the Copper Cliff quartz diorite dyke, against melted rocks of the Copper Cliff Rhyolite country rocks. This degree of melting is highly unusual for normal mafic dykes of this size and clearly indicates the superheated state of the first injections of the Copper Cliff Offset dyke. **B**) Quench textures in outer (unmineralized) quartz diorite indicating rapid cooling from a superheated states (no crystallization nuclei), followed by rapid crystallization. **C)** Inner inclusion-bearing quartz diorite with rusty sulphide blebs, cut by a last phase of plagioclase-phyric quartz diorite dyking along the core of the Worthington Offset dyke, Aer-Kidd Mine area.

Stop 6: Extensively developed Sudbury Breccia developed near the top of the Copper Cliff Rhyolite Formation, Lively Arena

46.427506° N, 81.144626° W
488888E, 5141558N

This glacially polished outcrop shows spectacular development, essentially chaotic at the ~5–10 m scale, of Sudbury Breccia (often abbreviated to SBX). Metre-size clasts, some rounded, float around in a dark matrix of pulverized rock flower that in some cases may have melted. Many of the fragments can be linked to the surrounding Copper Cliff Rhyolite Formation, the rhyolite that formed the uppermost felsic component of the basal rift succession of the Huronian. However, other fragments are more "exotic", relative to local outcrops, and illustrate significant movement of breccia material, and injection for some distance into dilating fractures.

This outcrop illustrates an important truism: any kind of breccia is an interesting breccia!

Seeing this amount of breccia, and the dynamic processes that must have been involved, requires a generative process with sufficient cause and energy, and volcanism clearly is not it. Although this particular outcrop is very spectacular, similar breccia bodies, dykes, and veins occur all over the Sudbury area (in the footwall), and even early workers were familiar with the fact that SBX occurred up ~60–70 km away from the Sudbury Igneous Complex.

In hindsight, and with our present understanding of Solar System geology, it is easy to see that only an ancient meteorite impact had sufficient energy to do this much damage, and at this scale. As stated in the introduction, it is perplexing how early workers clung to the cryptovolcanic explosion model for that long, perhaps largely due to non-familiarity with the emerging knowledge of the geology of the Moon and other planetary bodies. Robert Dietz, being exposed to this relatively new science of lunar geology in the 1950s, and also having carefully read the early papers/reports by Boon and Albritton (1936, 1937, 1938), quickly broke through this stasis in geological thinking.[6]

There is a general lesson here: always broaden your horizon, think out of the box, and listen to what may be totally new views out of left field.

One final anecdote: One of us remembers, as a 21-year-old undergraduate student back in Amsterdam, in 1980, sitting through a (very) long lecture by an eminent metamorphic petrologist talking about the Vredefort Dome in South Africa, and how it could only be an endogenic crypto-explosion domal structure based on this or that metamorphic reaction ... . This was 20 years after Hargraves (1961) and Dietz (1961) showed a systematic pattern of shatter cones around the dome, which even then was long known to be riddled with large pseudotachylite bodies and dykes (Shand, 1916).

Stop 7: Pillow lavas of the Elsie Mountain Formation, lower Huronian volcanic rocks

46.442417° N, 81.147691° W
488655E, 5143216N

In these roadside outcrops along the main road north out of Lively, the mafic volcanic rocks of the lowermost Huronian rift succession are well exposed. Plagioclase-phyric basaltic pillow lavas of the lowermost Elsie Mountain Formation are steeply dipping and facing south (Figure 17). The pillows show well developed rims and, although moderately flattened, show enough asymmetry to determine top directions, to the south.

These lavas, together with the Copper Cliff Rhyolite, comprise a bimodal succession typical for continental rifts. Their overall age is 2460–2480 Ma. The mafic lavas, with their prominent plagioclase crystals, were fed by the similarly plagioclase-phyric diabase dykes of the *circa* 2460

[6] The short and clearly written papers by Boon and Albritton in the 1930s could be characterized as "Sleeping Beauties" (van Raan, 2004; see also Ke et al., 2015; Miura et al., 2021), i.e. papers that were ahead of their time, and went largely unnoticed until they were "discovered" decades later. Boon and Albritton (1937, 1938) add Vredefort to their growing list of suspected impact craters.

Ma Matachewan swarm (e.g., Heaman, 1997), which are of the same age and which riddle the Archean basement to the north. None of these dykes cut across the uppermost volcanic unit, the Copper Cliff Rhyolite, where they would be easily noted. Hence, the Copper Cliff Rhyolite marks the final, felsic phase of this rift magmatism. Together with the subvolcanic A-type granite bodies of the Creighton Granite, and similar plutons along strike, this final phase has been dated at *circa* 2460 Ma (2459±7: Bleeker et al., 2015).



Figure 17: Pillow lavas near the top of the Elsie Mountain Formation, steeply dipping and younging to the south.

Stop 8: Creighton Mine, among the largest and deepest mines of the structure

46.461018° N, 81.176757° W
486427E, 51415287N

We will make a brief stop here on the access road to the Creighton Mine, one of the largest and deepest mines of the Sudbury structure. Various magmatic sulphide deposits occur at or near the basal contact of the SIC in what is one of the more prominent "embayments" along the basal contact. Creighton Mine geology and structure will be introduced and discussed by means of the composite cross-section shown below (Figure 18).

Overall, the structure of Creighton Mine is typical for the more strongly deformed South Range of the Sudbury structure, with the basal contact of the SIC dipping ~45–50º at surface and steepening at depth, with second-order structures superimposed. At depth, the steep basal contact is cut and offset by a discrete south-dipping shear zone, with south-side up displacement (e.g., Papapavlou et al., 2018), which is part of the post-Penokean "South Range Shear Zone" deformation that has further shortened the South Range.

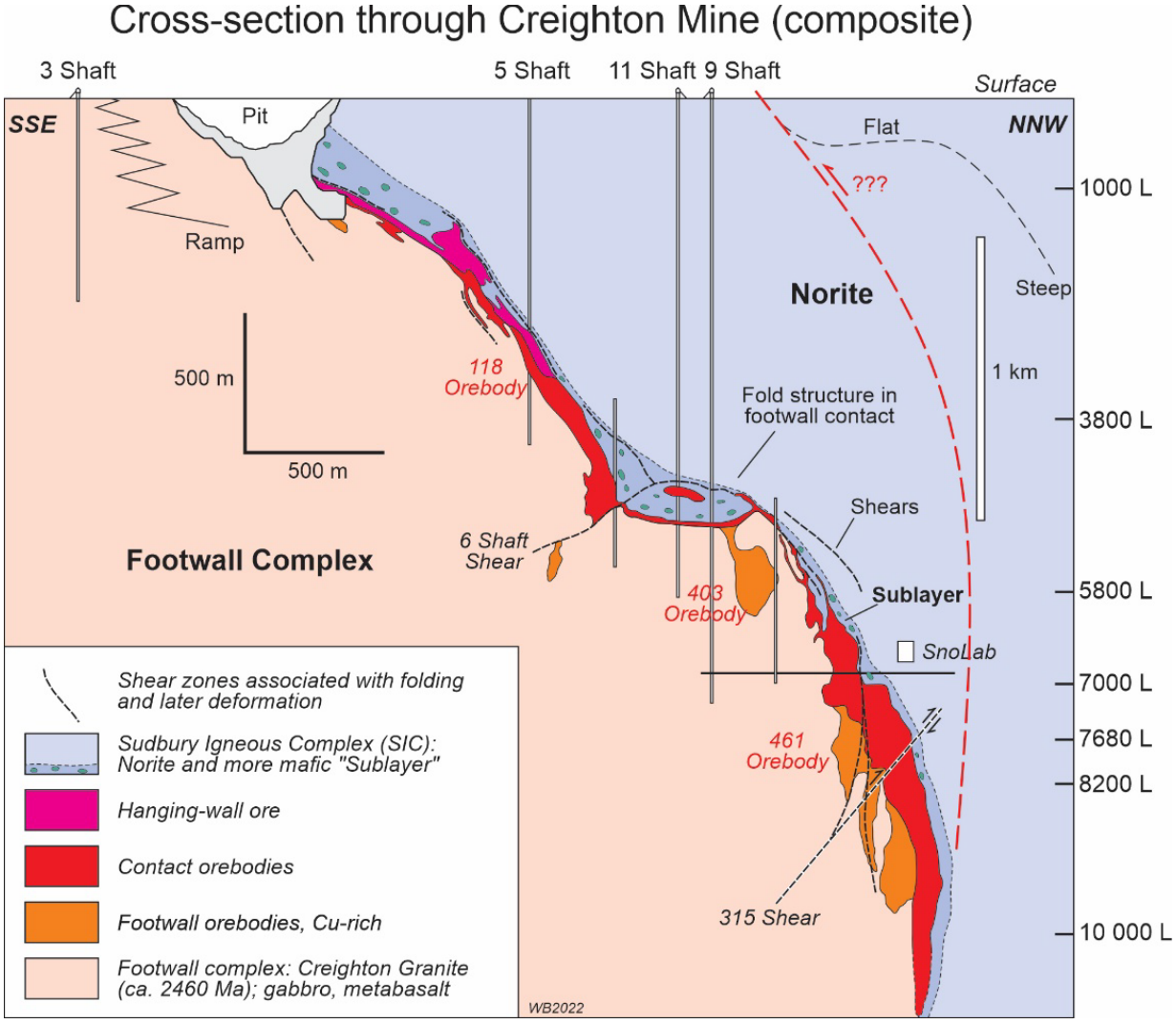


Figure 18: A) Composite cross-section of the Creighton Mine with its various ore lenses (projected onto a common section). Creighton Mine is one of the largest producers and it is getting very deep. As shown on the section, there is the typical variety of ore types, including more fractionated Cu-rich ore in the footwall, often controlled to some degree by local structures. Section modified after various published sources and original Inco mine sections. Because of the large depth and deep mine infrastructure, the mine also hosts one of the major neutrino labs in the world, the SnoLab, at about 2 km depth. Research at this lab was among the work that has demonstrated some of the fundamental characteristics of these elementary particles.



Figure 18: B) Photo of an active mining face in a Cu-rich orebody in the footwall complex below the SIC at Creighton Mine. The vein-like body is several metres wide and shows sharp, in part structurally-controlled, contacts. Inclusions of wall rock are suspended in the sulphide matrix and in various states of dismemberment by both physical and magmatic processes.

Stop 9: Creighton Granite, gabbro enclaves, and Sudbury Breccia

46.453751° N, 81.186197° W
485700E, 5144481N

To the south of Creighton Mine, a ~300 m-long section of the Lively regional road exposes superb outcrops of the Creighton Granite pluton, variably affected by dykes and veins of dark Sudbury Breccia (SBX). The granite hosts major enclaves of early Huronian gabbro/diabase sills or dykes, some with very prominent zone calcic plagioclase megacrysts, which clearly link them to the Matachewan magmatism and large igneous province.

There are also remnants of lower Huronian volcanics with interlayered sandstone layers, which here show graded bedding suggesting tops are to the north, not to the south.

The somewhat coarser grained gabbro enclave at the east end of the road section has been dated at *circa* 2479 Ma (Bleeker et al., 2015), thus being part of the early Matachewan (Matachewan I) event, which also emplaced larger layered intrusions at the base of the Huronian section, at or near the unconformity with Archean basement.

Overall, it is rather challenging to get highly precise U-Pb ages on many of units in this area, due mainly to two related reasons: 1) all the pre-1850 Ma zircons are shocked and disturbed, and 2) the zircons are generally altered. Results on the Creighton Granite and the Copper Cliff Rhyolite are shown in Figure 19, showing the scatter and general “pull down” due to shock-induced Pb loss at 1850 Ma, with superimposed younger Pb loss. Many individual analyses, whether single grain or multigrain, are almost meaningless due to these

combined Pb loss effects. This resulted in the *circa* 2350 Ma early age estimates for the Creighton Granite (Frarey et al., 1982), a result based on several un-abraded multigrain zircon fractions that would have averaged out all these effects and plot in the middle of the Pb loss array or triangle (see Figure 19). The current apex of this Pb loss triangle is constrained by analyses of single, small, best-preserved zircon grain fragments pre-treated by chemical abrasion. Collectively the data indicate a minimum age of 2455 Ma for the Creighton Granite and the co-magmatic Copper Cliff Rhyolite, with a most likely upper intercept age of 2459 $^{+7}/_{-4}$ Ma.

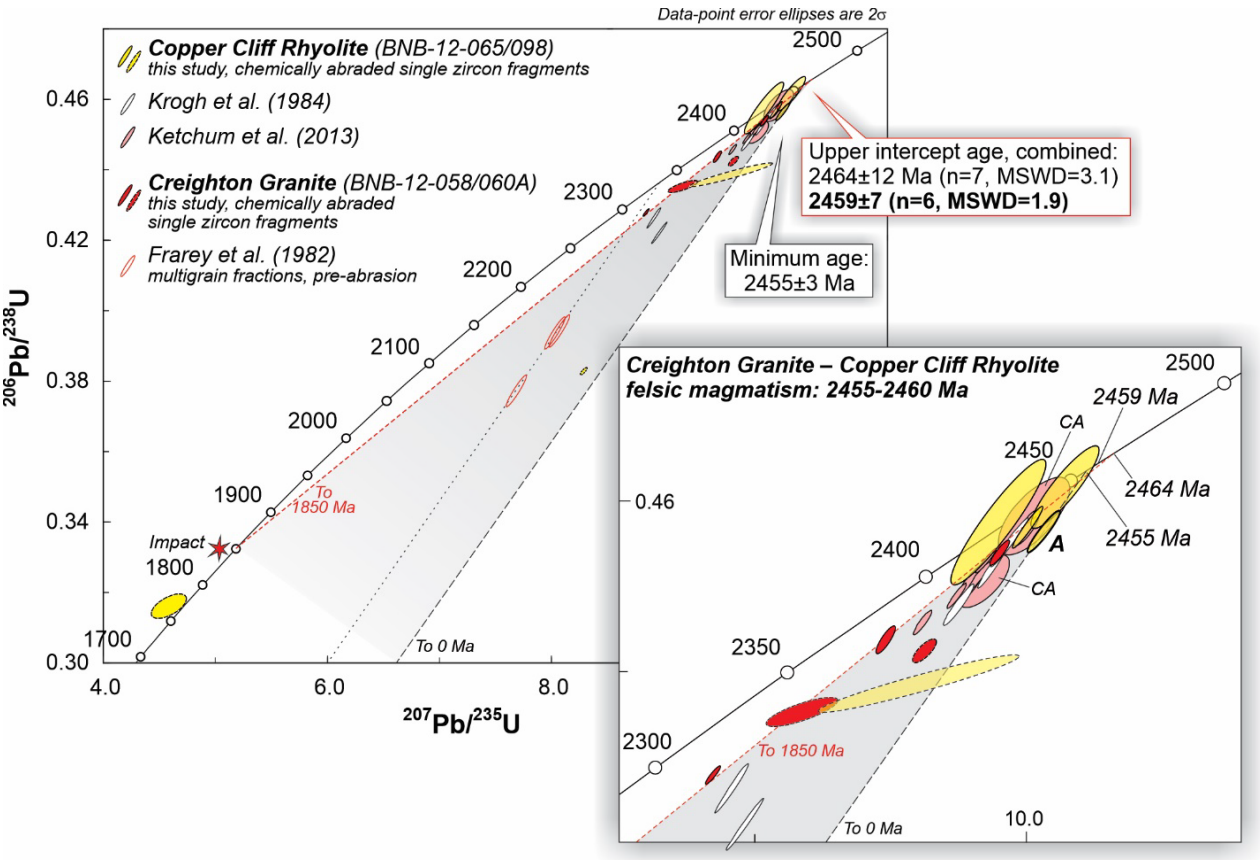


Figure 19: U-Pb concordia diagram of the combined results on Creighton Granite and Copper Cliff Rhyolite samples, showing the complex Pb loss array or "triangle" formed by various Pb loss processes. Shock effects of the Sudbury impact have damaged most if not all of the zircon crystals to varying degrees (e.g., Krogh et al., 1996), with results being pulled down to an 1850 Ma lower intercept. The variably altered zircons were then affected by varying stages and degrees of younger Pb loss, including recent Pb loss. Large multigrain fractions, non-abraded, from the early Frarey et al. (1982) study, plot in the middle of the triangle, having averaged out all the various Pb loss processes. Only tiny, best-preserved, single zircon fragments pre-treated by chemical abrasion (CA) from recent studies (Bleeker et al., 2015) plot near the apex of the triangle, constraining both a minimum (2455 Ma) and most likely upper intercept age of 2459+7/-4 Ma for the felsic magmatism. In contrast, comparatively large laser ablation spots on these complicated zircons, without CA pre-treatment, will simply sample all this complexity in Pb loss and result in meaningless upper intercept ages.

Stop 10: Magma mingling structures between Creighton Granite and mafic magmas

46.450530° N, 81.195107° W
480515E, 5144125N

Just south of the intersection of the Lively regional road with the Highway 144 Bypass, on the east side of the highway, are roadcuts through the Creighton Granite with classic magma mingling structures: rounded blobs of mafic magma suspended in surrounding Creighton Granite. This is significant in the sense that it clearly demonstrates contemporaneous mafic and felsic magmatism in an overall bimodal magmatic system.

This is relevant to the interpretation of the Creighton Granite and the confusion about the "Blezardian orogeny". The Creighton Granite is not a terminal collisional granite, i.e. the interpretation that fed the idea of a Blezardian orogeny, but rather an early A-type granite associated with the final rift-related magmatism at the base of the Huronian Supergroup.



Figure 20: Typical magma co-mingling structure of rounded blobs of mafic magma interacting with K-feldspar porphyritic granitoid magma of the Creighton Granite.

Stop 11: Basal contact of the Sudbury Igneous Complex

46.459667° N, 81.203009° W
484411E, 5145142N

Roadcuts along the Highway 144 Bypass expose the basal contact of the SIC on underlying Elsie Mountain dark, mafic volcanics and gabbro sills. As was shown at Stop 9, younging in the volcanic section is most likely to the north here, into the base of the SIC.

These outcrops show how difficult it is to actually put one's finger on the lower contact of the SIC, due to the nature of immediate footwall rocks.

The section is locally cut by felsic dykes, which are "rheomorphic" dykes of melted footwall that were back-intruded into the base of the SIC. These dykes are of interest in terms of constraining final melt formation and migration in the footwall rocks, but they are very hard to date because they contain essentially all xenocrystic zircons from the underlying rocks (e.g., Creighton Granite), with all their shock damage, and very few newly grown zircon crystals. A couple of hundred metres farther to the north, there are the first outcrops of typical, massive, homogeneous norite from near the base of the Main Mass of the SIC.

Stop 12: Top of the norite section, across the transition zone gabbro, and into the base of the granophyre section

46.487202° N, 81.207053° W
484109E, 5148202N

About 3 km north of Stop 11, just north of the powerline and in a lazy curve of the Highway 144 Bypass, occurs the transition zone from uppermost norite, into a ~500 m thick gabbro section where augite becomes the dominant pyroxene, rather than hypersthene, and into the base of the thick granitic granophyre section.

Figure 21 shows the typical variation in modal mineralogy across the SIC, as compiled from various sources. The "Transition Zone Gabbro" crystallized oxides and apatite, giving it a much higher magnetic susceptibility. Figure 22 shows some petrographic details of the "black norite".

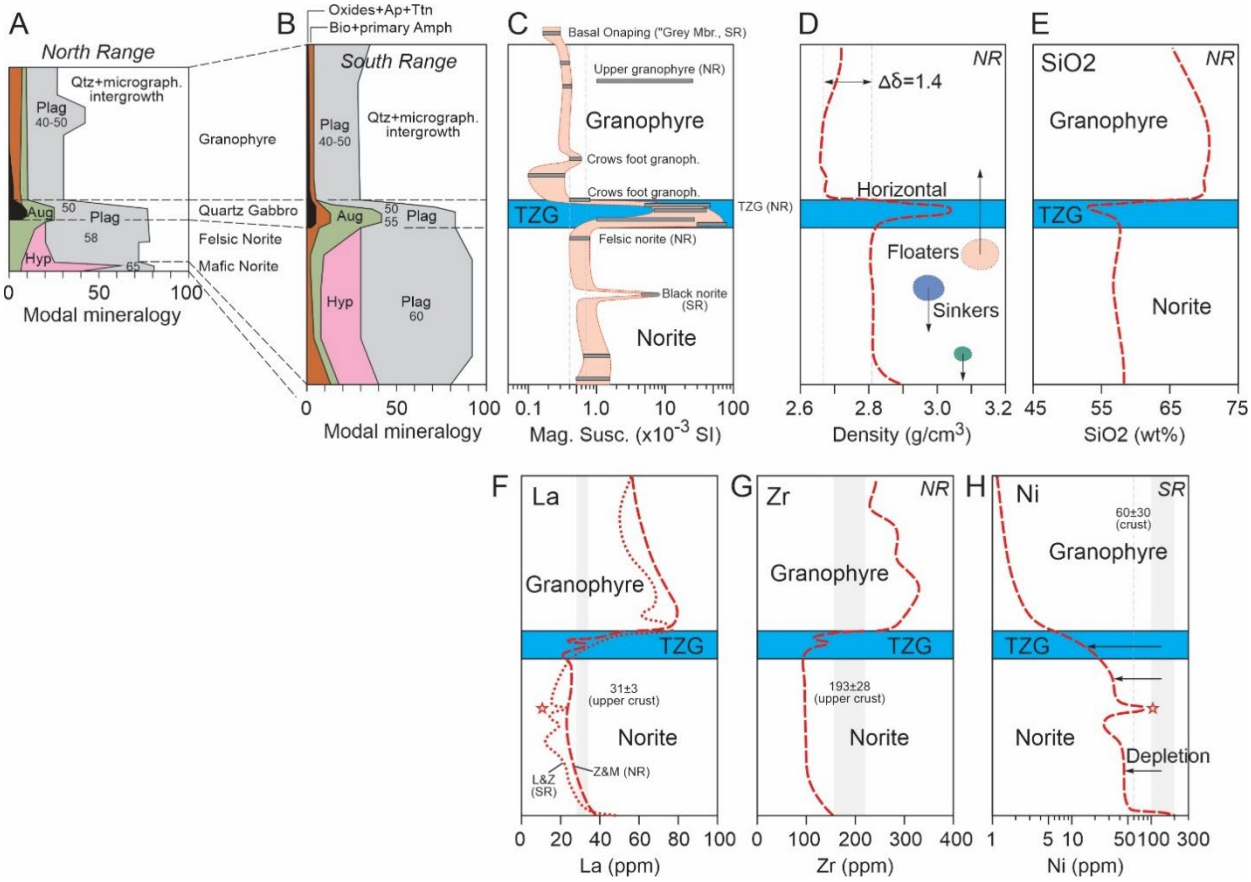


Figure 21: First-order variations in modal mineralogy, density, and whole-rock chemistry across the Main Mass of the SIC. Figure A and B from Naldrett and Hewins (1984); C from observations by the first author; D, E, F and G from Zieg and Marsh (2005), North Range; F and H from Lightfoot and Zotov (2005), South Range along Highway 144. Profiles resized to a common scale to highlight first-order features. Note the interesting spike in Ni values in the South Range "black norite" (see star in H), where samples also show large scatter in La (see G)), approaching values in basal norite. This area in South Range norites also shows shallow mineral lamination, suggesting a major structure may repeat the basal norites (see section of Figure 18). Grey bars are: in H, initial Ni values in early quartz diorite, highlighting the large Ni depletion in much of the SIC; the dashed line indicates Ni values in glassy melt fragments in the Onaping Formation (Ames et al., 2002); in F and G, upper crustal average values from Rudnick and Gao (2005).

Toward the top of the gabbro section occurs a ~50 m-wide zone of very coarse-textured "Crows-foot Granophyre" (Figure 23). This probably reflects the final accumulation of H_2O in an evolved residual gabbro magma, promoting coarse crystal growth in the residual melt. So, although this unit is called "granophyre" it is probably better seen as the top of the gabbro section.

Up from here, one enters the base of the "Granophyre" proper, which is of broadly granitic composition and fairly evenly grained and homogeneous.

Based on normal liquidus and solidus temperatures for these various compositions, one predicts that the base of the granophyre would crystallize last (see Figure 6). Overall heat loss would be highest from the roof of the SIC, and less so into the footwall of the SIC. Crystallization of the norites and gabbro would add latent heat of crystallization into the base of the granophyre section, thus keeping it hot and molten until the upper crystallization front closed in on it from above.

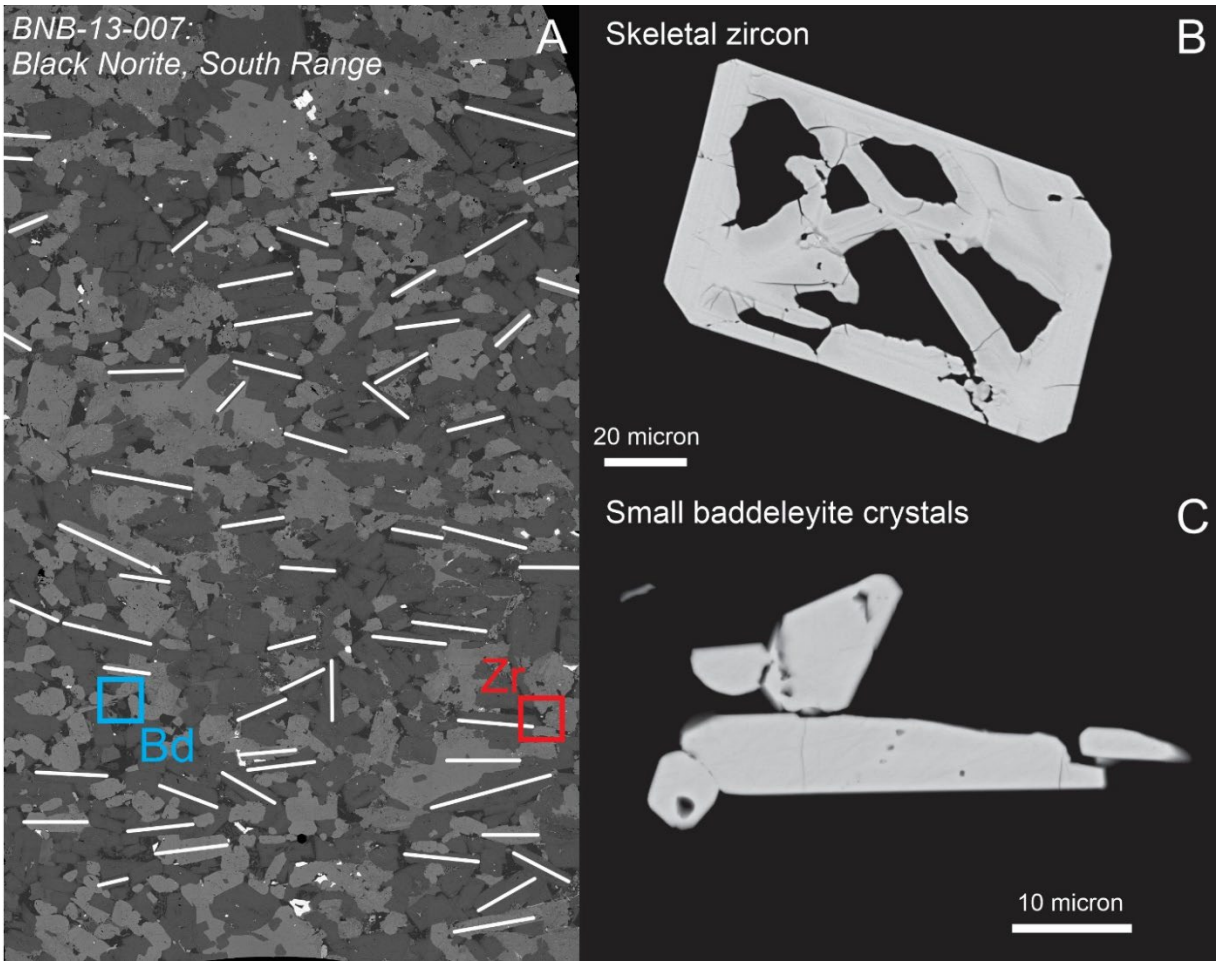


Figure 22: Thin section image (SEM, back-scattered electron map) of "black norite" showing typical mineral lamination of platy feldspars; typical zircon and baddeleyite crystals are highlighted and shown in close-ups on the right. Note the characteristic dendritic or skeletal habit of the zircons, typical for the norites, suggesting crystallization in a rapidly cooling melt, originally superheated with no nuclei. Zircons from this sample were dated using the Pb evaporation technique, resulting in an age of 1849.7±0.2 (Bleeker et al., 2015).

With these expectations, we attempted to precisely date final crystallization of the Main Mass and this resulted in a precise and concordant zircon age of 1850.0±0.9 Ma, which fully overlaps with the best results on ages for the norites. From this we can conclude that crystallization of the entire SIC melt sheet was all within a million years, and likely well within the current resolution of the best available ages. As our ages get better and more precise, this age range of crystallization and cooling may shrink further.

The boundary between the top of the Transition Zone Gabbros and the base of the Granophyre represents a major boundary in physical parameters, among them density (Figure 21). The granophyres are on average 1.4 g/cm^3 less dense than the norite section and much less dense than the oxide-rich gabbros. The base of the thick granophyre section, also last to crystallize, would thus have equilibrated in an essentially horizontal position after settling of the thick melt sheet. It thus represents an important "paleo-horizontal" reference surface when thinking about the overall structure.

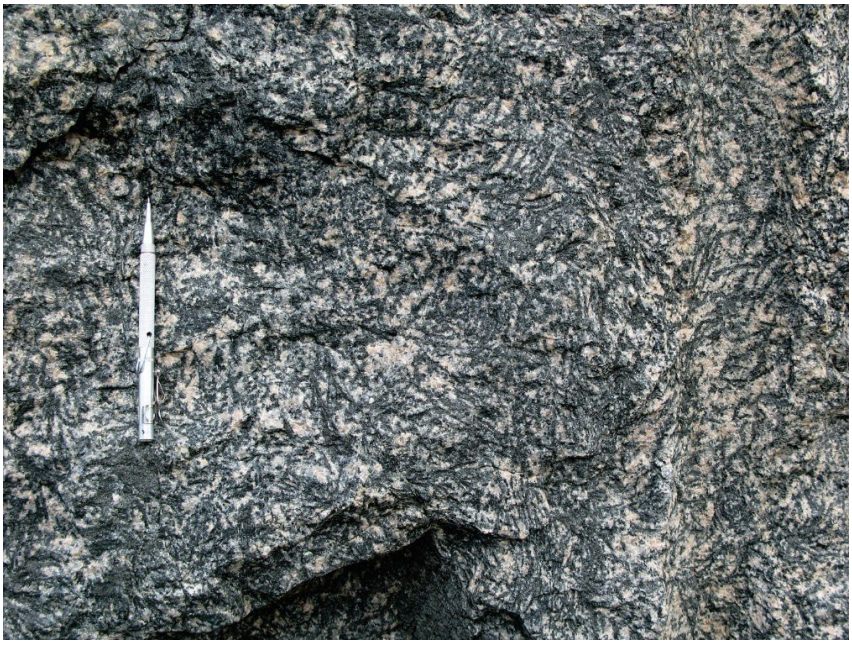


Figure 23: Coarse-grained radiating textures of the "crows foot granophyre", which probably represents the H2O-rich residual liquid at the top of the Transition Zone Gabbro section. It resembles typical pegmatitic zones at the top of other large differentiated gabbro sill complexes (e.g., Nipissing Diabase sills). If so, this is indeed better interpreted as the top of the gabbros, rather than the base of the granophyre section of the SIC.

Stop 13: Typical granophyre, lower half of the Granophyre section

46.498015° N, 81.204066° W
484341E, 5149403N

A large pull-out on the west side of the Highway 144 Bypass provides an easy place to examine typical granophyre about 1 km above the transition with the gabbros. Here the granophyre is medium grained, and evenly textured.

Early workers on the "Sudbury Irruptive", among them petrologists, had noted of course that the SIC was unusual in two main ways: 1) the basal norites being more silica-rich (~56–58 wt%) than other large mafic intrusions, and 2) being characterized by a very thick granitic "granophyre" section. Typical large layered intrusion would differentiate into a mafic-ultramafic layered base, and overall perhaps ~10% of evolved granitic granophyric material underneath the chilled roof section. Clearly something was odd about Sudbury!

Stop 14: Basal section of the Onaping Formation, "Grey Member" rich in angular to rounded Huronian quartzite fragments

46.524529° N, 81.195052° W
485040E, 5152347N

This spectacular outcrop shows a perfect example of the clast-rich base of the Onaping Formation, with a clast population dominated by Huronian quartzite fragments 1–20 cm in size, somewhat rounded to angular. The fragmental material is tightly compacted into a "suevite breccia", the latter name used when impact melt clots can be recognized.

The fragmental rocks are moderately to strongly deformed with a strong, southeast-dipping schistosity/cleavage that is axial planar to the overall syncline of the Sudbury structure, with a stretching lineation that plunges down-dip (on the cleavage plane). Although these structural elements are largely Penokean in origin, they were likely amplified by the younger deformation of the South Range Shear Zone (Shanks and Schwerdtner, 1991).

Stop 15: Upper member of the Onaping Formation ("Black Member")

46.535044° N, 81.185639° W
485765E, 5153514N

With extensive rock cuts on both sides of the Highway Bypass, this is a good outcrop to examine the upper part of the Onaping Formation, which is finer grained and overall upward fining (see Figure 6), darker, and traditionally was called the "Black Member" of the "fall-back breccias". The darker colour of the breccias going up-section reflects, in part, a finer-grained matrix, and also an increasing carbon content.

In a general sense this "Grey Member" and "Black Member" terminology is still useful, but more detailed mapping in the last two decades has shown a more complex system of different depositional units including volcanic deposits due to venting of a still active melt sheet, and debris flows (*see* the work by Ames et al. (2008a,b) and references therein). So only part of the Onaping Formation represents true suevitic fall-back breccias. Other parts were reworked or washed back into the crater by processes other than strict fall back.

Stop 16: Greywacke turbidites of the uppermost Whitewater Group, the Chelmsford Formation

46.575577° N, 81.289467° W
477820E, 5158042N

This roadside outcrop shows the folded and cleaved greywacke turbidites of the Chelmsford Formation, the uppermost formation preserved in the core of the Sudbury Basin syncline. The approximate age of these turbidites is *circa* 1840 Ma and they represent the foreland depositional wedge of the Penokean orogen.

Participants of the fieldtrip who have seen the Rove Formation in the Thunder Bay area will recognize the great similarities and, indeed, these two formations are broadly related. The reason these turbidites are only locally preserved has to do with the fairly high-amplitude down folding in the core of the doubly-plunging Sudbury Basin syncline, which preserved the folded foreland sedimentary wedge from regional uplift and erosion.

Prior to the regional uplift, over the course of the Proterozoic, the Chelmsford and Rove Formation foreland wedges may well have connected.

Bedding in the Chelmsford Formation is varied, with locally thick sandy turbidite beds. Bedding is folded and overprinted by the regional cleavage that is perfectly aligned with the axial plane of the overall fold structure. Large carbonate concretions are deformed and give an indication of the finite strain.

Stop 17: Black Member of the Onaping Formation on the North Range, along Highway 144 at Onaping Falls

46.589851° N, 81.382453° W
470702E, 5159658N

Here along the highway, and the railway cut above, the Black Member of the Onaping Formation is well exposed in large rock cuts. Micro-diamonds have been reported from these rocks (Masaitis et al., 1999; see also French, 2004).

Stop 18: Transition Zone Gabbro, North Range, at Highway 144 – Highway 8 intersection

46.617271° N, 81.413896° W
468309E, 5162717N

Outcrops near this intersection expose the Transition Zone Gabbro. The gabbro is medium grained, and characterized by augite being the main pyroxene, with little or no hypersthene. The gabbros also show a spike in oxide (Fe, Ti) and apatite (P) crystallization, typical of the "peak" in relative Fe-Ti concentrations during progressive crystallization, as seen in AFM diagrams of relatively reduced and anhydrous mafic magmas. This is reflected in a spike in magnetic susceptibility, with values increasing by an order of magnitude. The transition zone gabbro is several hundred metres thick and upward transitions into the base of the granophyre section.

Stop 19: Felsic norite towards top of norite section on North Range

46.615361° N, 81.428390° W
467198E, 5162571N

These outcrops in the broad curve of the highway at the bottom of the hill expose the top of the North Range norites, referred to as "Felsic Norite" (e.g., Naldrett and Hewins, 1984). The norite is massive and homogeneous, as is typical for most of the SIC rocks. SiO_2 contents of these norites are ~58 wt%. Ni values are ~20–30 ppm, which is significantly depleted from values of what are thought to have been primary Ni values in the early undifferentiated impact melts that were perhaps as high as ~100–200 ppm (Zieg and Marsh, 2004; Lightfoot and Zotov, 2005; Lightfoot, 2006).

A subtle mineral lamination, formed mainly by alignment of platy plagioclase crystals, can be seen and dips moderately to the south, parallel to the attitude of the basal SIC contact in this part of the North Range. Magnetic susceptibility values are ~7.5±1.0 $x10^{-3}$ SI units.

Stop 20: Levack Gneisses, cut by Matachewan diabase dykes, all overprinted by Sudbury Breccia and southeast pointing shatter cones

46.624698° N, 81.444722° W
465954E, 5163556N

Roadcuts on both sides of the highway show typical "Levack Gneisses", high grade migmatitic gneisses that characterize the Archean basement to the north of the SIC. These rocks reached pyroxene granulite facies grade, before being retrogressed to amphibolite facies in the latest Archean (e.g., Prevec et al., 2005, and references therein). The gneisses contrast with more homogeneous late Archean granites farther north, the Cartier Granites, dated at *circa* 2640 Ma (Meldrum et al., 1997).

The gneisses are cut by large, SSE-trending, subvertical mafic dykes with conspicuous plagioclase pheno/megacrysts. These dykes are *circa* 2460 Ma in age and part of the giant Matachewan (II) dyke swarm. They represent the feeders to the plagioclase-phyric basalts of the lower Huronian mafic volcanic rocks seen at Stop 7 north of Lively. They are numerous in this part of the Superior Province and provide an important time marker. As pointed out earlier, they do not cut the Copper Cliff Rhyolite, which represents the terminal phase of the Huronian rift magmatism and volcanism.

Both the gneisses and the Matachewan dykes are overprinted by Sudbury Breccia, and shatter cones can be seen at several localities, with the cone axes projecting to the southeast. One well-developed cone in the gneisses has a cone axis of ~120º/+20º (up).

Well-preserved pseudotachylite of the Sudbury Breccia bodies, locally up to ~1 m wide, is very dark in colour here below the North Range, and only weakly recrystallized. This contrasts with pseudotachylite on the South Range (e.g., Stop 9), where it is typically strongly recrystallized and metamorphosed to epidote amphibolite facies and shows cleavage/foliation development due to the more intense deformation of the South Range.

Given this relatively good state of preservation of the black pseudotachylite matrix here at these outcrops, it is unlikely that the Levack Gneisses were at a lower crustal level, and thus hot, at the time of impact. The gneisses were likely exhumed to shallower crustal levels during the latest Archean or earliest Paleoproterozoic, well prior to the impact (e.g., James et al., 1992). This point has been debated in the Sudbury literature and is relevant to the question of where "Ground Zero" is located. To the northeast of the Sudbury structure, Levack Gneisses occur in close proximity to the unconformity with the Huronian succession.

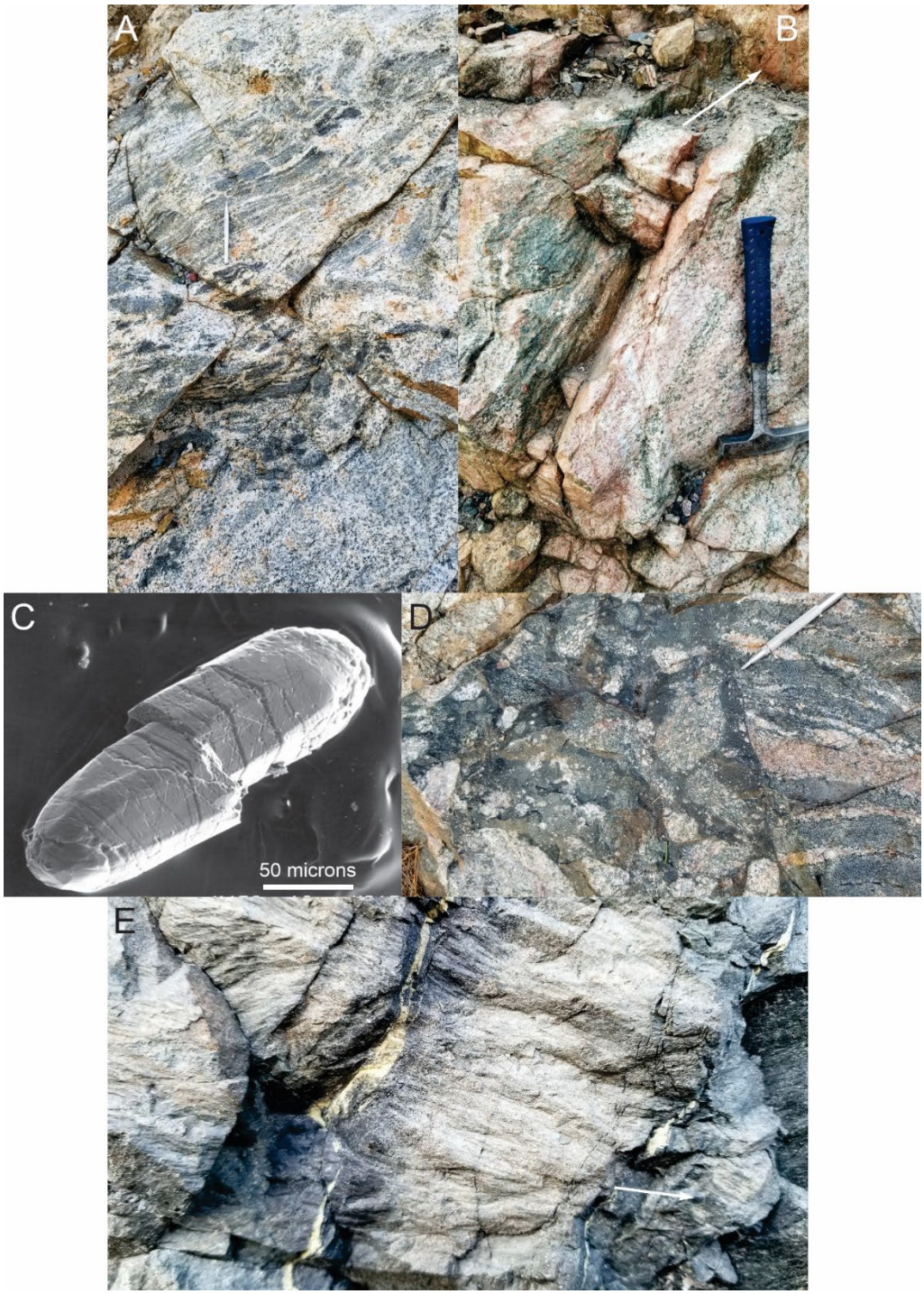


Figure 24: Shatter cones and Sudbury Breccia in the Levack Gneisses ~0.5 km below the basal contact of the SIC on the North Range. **A)** Photo of typical, heterogeneous, coarse-textured Levack Gneiss. **B)** Relatively well-developed partial shatter cone in Levack Gneiss, ~0.5 m in size. **C)** Shocked zircon with multiple sets of planar deformation features and fractures, from the original study of U-Pb dating of these rocks (see Krogh et al., 1984). **D)** Well-developed Sudbury Breccia with displaced and variably rounded gneiss fragments floating around in a black pseudotachylite matrix. **E)** Small shatter cones, ~10-25 cm in size, developed within the Matachewan dyke at this stop, with cone axes focusing toward the southeast (to the right in this picture).

Stop 21: Cartier Granites, with locally conspicuous Sudbury Breccia, and rare shatter cones

46.668937° N, 81.532490° W
459268E, 5168513N

This final stop or stops, ~10 km out (horizontally) from the footwall contact of the SIC, will examine the more homogeneous late Archean granites of the "Cartier Batholith", part of the "Algoman granites" to the south of the main Abitibi granite-greenstone terrane. Although somewhat variable and locally showing relict layering, typical parts of the Cartier Granite are homogeneous, relatively massive, pink, and relatively K-feldspar-rich late granites with little structure. However, despite locally conspicuous K-feldspar, petrographically the granites are largely monzogranitic to granodioritic in composition, typical for late Archean granites. They contrast with and intrude the older Levack Gneisses. These late-stage granites, reflecting final re-melting of earlier tonalite-trondhjemite-granodiorite (TTG)-dominated granite-greenstone crust, have been dated at *circa* 2640 Ma (Krogh et al., 1984; Meldrum et al., 1997)). They represent a final stage in the Archean crustal evolution of the southern Superior craton prior to stabilization and "cratonization".

Occasional shatter cones can be seen in the road cuts, all pointing to the southeast (e.g., GPS waypoint #2686). And there is abundant Sudbury Breccia in places (GPS #2522), forming veins, dykes, and larger breccia bodies with rounded fragments of granite in a black pseudotachylite matrix. Similar Sudbury Breccia occurrences can be mapped radially outwards for another ~75 km, out to ~120 km from "Ground Zero" (*see* Figure 8; *see also* Butler, 1994; and Thompson and Spray, 1994) [7].

Stop 22: Discovery outcrop along railway cuts, just east of Murray Mine, South Range

46.521393° N, 81.052983° W
495936E, 5151982N

At this final stop we visit the approximate "discovery outcrop" just east of the Murray Mine. When in 1883 a railway line was cut through this area, Fe-sulphides and some chalcopyrite were noted in the rock cuts. This led to more prospecting and, eventually, the first mining operations in the area. At the time, the main interest was in Cu, as the application of Ni to steel manufacturing had not yet been invented.

The original rock cut was a bit farther to the west, but both the railway and the roads were moved east to allow for the development of the Murray open pit, which is located on the other side of the highway.

In the present railway cut, semi-massive sulphide veins anastomose around somewhat deformed mafic fragments, some containing minor disseminated sulphides, others with no sulphides. This kind of material is typical for the basal contact of the SIC and is generally referred to as the "Sublayer" or "Contact Sublayer" (e.g., Pattison, 1979). To the west, these minor sulphide stringers broaden out into the orebodies of the Murray Mine (Figure 25).

Figure 25 shows a composite section across the basal contact of the SIC in this area, illustrating the SIC dipping to the north at ~40–45º. Relationships are approximately similar to that shown in the Creighton Mine section. The Murray Mine area could be described as yet another minor embayment, just to the east of the major Copper Cliff funnel structure.

[7] Confusion is possible with pseudotachylite veinlets formed in relation to regional faults, unrelated in time to the Sudbury structure. However, such veins are typically 0.5–2 cm in width and can mapped along or in proximity to observed fault or slip planes. Almost all Sudbury Breccia, such as discussed here, is developed at a different scale, often as dykes or bodies up to 10-100 cm wide.

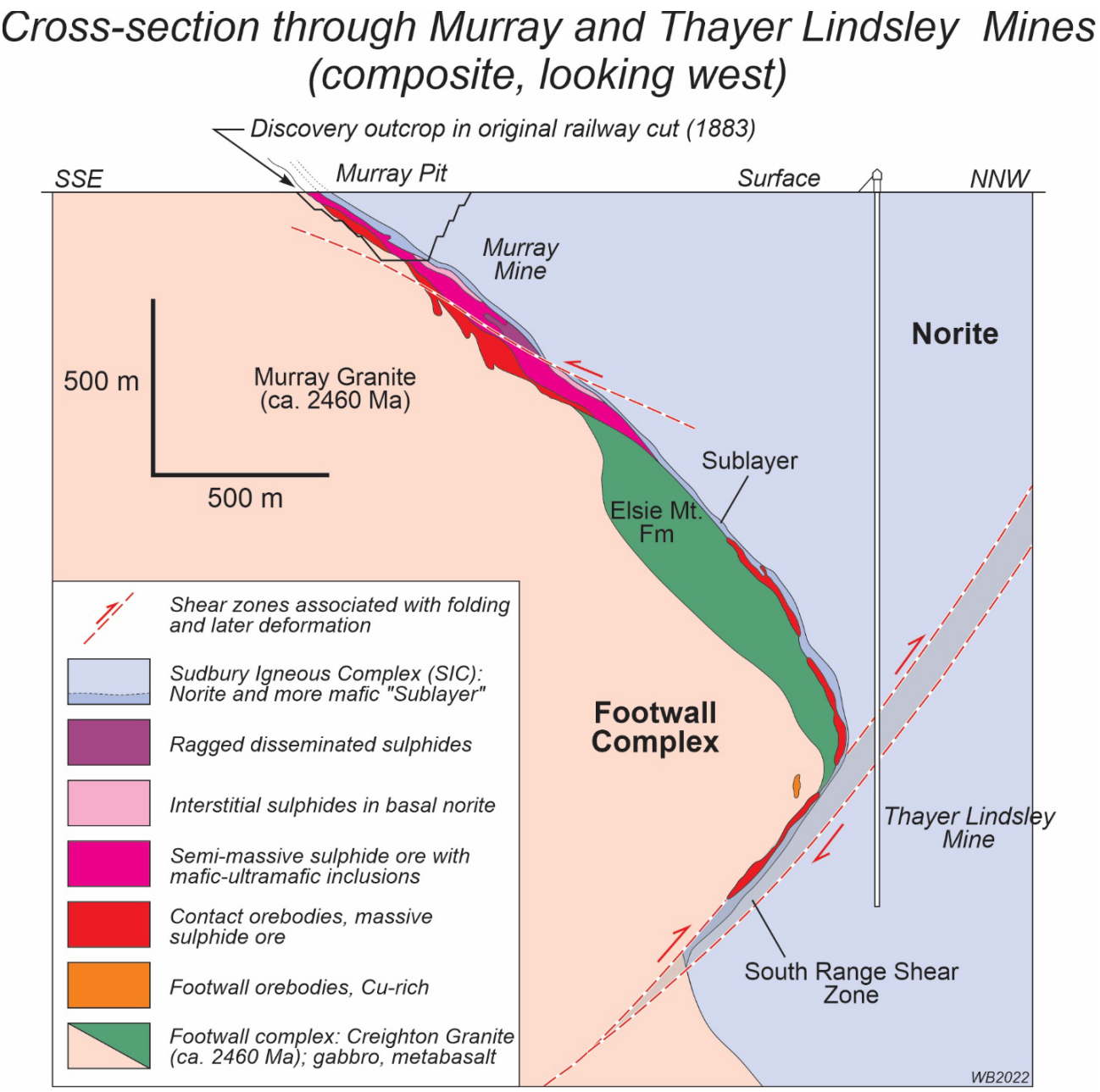


Figure 25: Composite cross-section, looking west, across the footwall contact of the SIC illustrating the first-order relationships at the Murray Mine and open pit, just to the west of the "Discovery Outcrop" locality along the railway bed, and the Thayer Lindsley Mine farther to the east. Sections of both mines are integrated at the same scale. The Murray Mine section is from old Inco data, as published by Naldrett (1984), and differentiates some of the ore types, all in close proximity to or right along the footwall contact. The outline of the open pit is schematic. The Lindsley Mine section is from old Falconbridge data (see Binney et al., 1994), as published by Bailey et al. (2004). It is one of the localities where the significant offsets due to structures associated with the South Range Shear Zone was first recognized. Ore bodies along this shear zone were highly deformed. A Cu-rich ore body occurred within the footwall complex, dominated by Murray Granite.

Incorporated into the section of Figure 25, at the same scale, is a section across the Thayer Lindsley Mine. This latter mine is located ~5 km to east-northeast along strike and represents a next shallow embayment along the footwall contact of the SIC.

At the Lindsley Mine, at depth, the basal contact of the SIC is offset by a well-defined, discrete, shear zone that dips to the southeast and shows significant south-side up displacement (Binney et al., 1994). Metamorphic grade in this shear zone is lower amphibolite facies and titanite crystals from sheared norite show two distinct growth phase, brown titanite overgrown by colourless titanite (Bailey et al., 2004). U-Pb data for two brown titanite fractions suggest they grew at *circa* 1815±15 Ma, whereas the colourless titanites, interpreted to be syntectonic relative to the shear zone fabric, record an imprecise but somewhat younger age, *circa* 1670±70 Ma (Bailey et al., 2004).

The latter age dates this marked shear zone, which is part of the South Range Shear Zone system that further shortened and imbricated the South Range of the SIC.

End of Road Log for Day 1

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Day 2
Sudbury Ore Environments and Offset Dikes – Examples from Whistle and Parkin, NE Sudbury

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1. Organization of Day 2

The second day of pre-meeting Trip 1 will introduce participants to the ore environments at the base of the Sudbury Igneous Complex (SIC) and to the general geology of offset dikes that are exposed on mechanically and hydraulically stripped outcrops in the northeast corner of the Sudbury basin. The day starts adjacent to the former Whistle mine, mined by Inco (now Vale) from 1993 to 1997, where large outcrops highlight the complexity of Superior Province footwall rocks near the contact with the overlying SIC and the transition from mineralized Sublayer through anatectic breccias to traditional offset dike lithologies. The Podolsky North Zone outcrop to the northeast contains footwall-style mineralization on surface, extensions of which were mined underground by FNX/KGHM from 2008 to 2013. In the afternoon, outcrops south (distal Whistle) and north (proximal Parkin) of the Post Creek fault will be compared. The field trip will end with a visit to Rocky's Restaurant on Lake Wanapitei.

In case we are not able to access the Whistle mine outcrops, several interesting and well-preserved outcrops of the Worthington, Trill, and Hess offset dikes to the southwest and northwest of the Sudbury Structure will be visited. Participants will be introduced to typical characteristics of offset dikes, emplacement mechanisms and formation models.

2. Introduction

The Sudbury mining camp is the one of the largest magmatic Ni-Cu-PGE mining camps in the world (Fig. 2.1) and has been mined for over 135 years (see review by Lightfoot, 2016). Mineralization is associated mainly with breccias along and near the lower contact of the Main Mass of the Sudbury Igneous Complex (SIC) and within associated offset dikes (Fig. 3.1).

Breccias in the Sudbury Structure include 1) pre-impact magmatic breccias (e.g., Levack Breccia), 2) syn-impact pseudotachylitic breccias, locally referred to as Sudbury Breccia (SUBX; e.g., Rousell et al., 2003), 3) syn- to post-impact magmatic breccias directly derived from the SIC, such as Inclusion-Bearing Quartz Diorite (IQD; e.g., Grant and Bite, 1984) and inclusion-bearing Sublayer Norite (SLNR; e.g., Lightfoot and Farrow, 2002; Lightfoot et al., 1997a), and 4) contact metamorphosed and/or partially melted (anatectic) breccias, variably referred to as Footwall Breccia (FWBX; e.g., McCormick et al., 2002) and "Metabreccia" (MTBX; e.g., Lafrance et al., 2014).

The ore deposits in the Sudbury Structure occur in two distinct environments (Fig. 3.1): 1) mineralization along or near the basal contact of the Main Mass, including a) disseminated to semi-massive Fe-Ni-Cu sulfides in SLNR and FWBX, and b) veins and disseminations of Fe-Cu-Ni sulfides in underlying SUBX and associated footwall rocks, and 2) disseminated-blebby to semi-massive Fe-Ni-Cu and Fe-Cu-Ni sulfides in offset dikes (e.g., Souch et al. 1969; Naldrett, 2004; Lightfoot, 2016).

The Whistle embayment and Whistle-Parkin offset dike at the northeastern corner of the SIC contain elements of both environments, providing an excellent introduction to Sudbury ore systems.

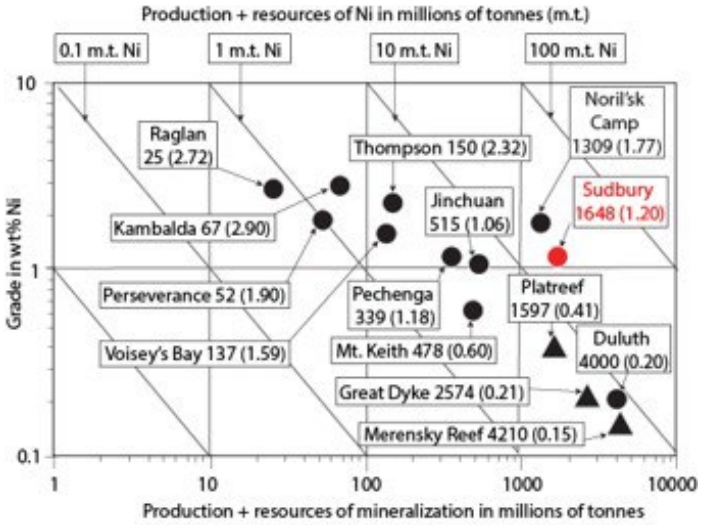


Figure 2.1: Pre-mining Ni resources (past production + current resources) and grades of the world's largest magmatic Cu-Ni (circles) and PGE (triangle) deposits. *Modified from* Naldrett (2004).

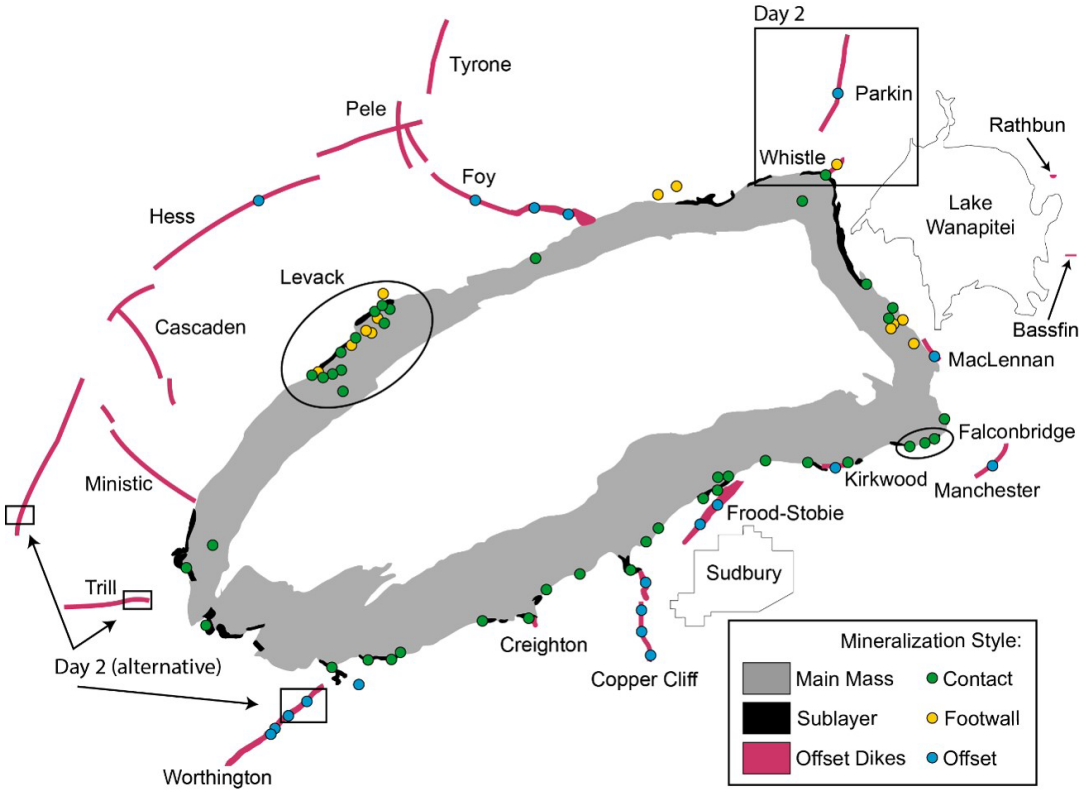


Figure 3.1: Distribution of contact, footwall, and offset dike deposits and occurrences in the Sudbury Impact Structure. Note that locations for deeper ore bodies are projected to surface. Simplified *after* Ames et al. (2008).

3. Sudbury Ore Environments

Ore deposits and occurrences occur all around the Sudbury Structure (Fig. 3.1), but some areas, such as Levack in the North Range and Frood-Stobie, Creighton, and Copper Cliff in the South Range, are much better endowed.

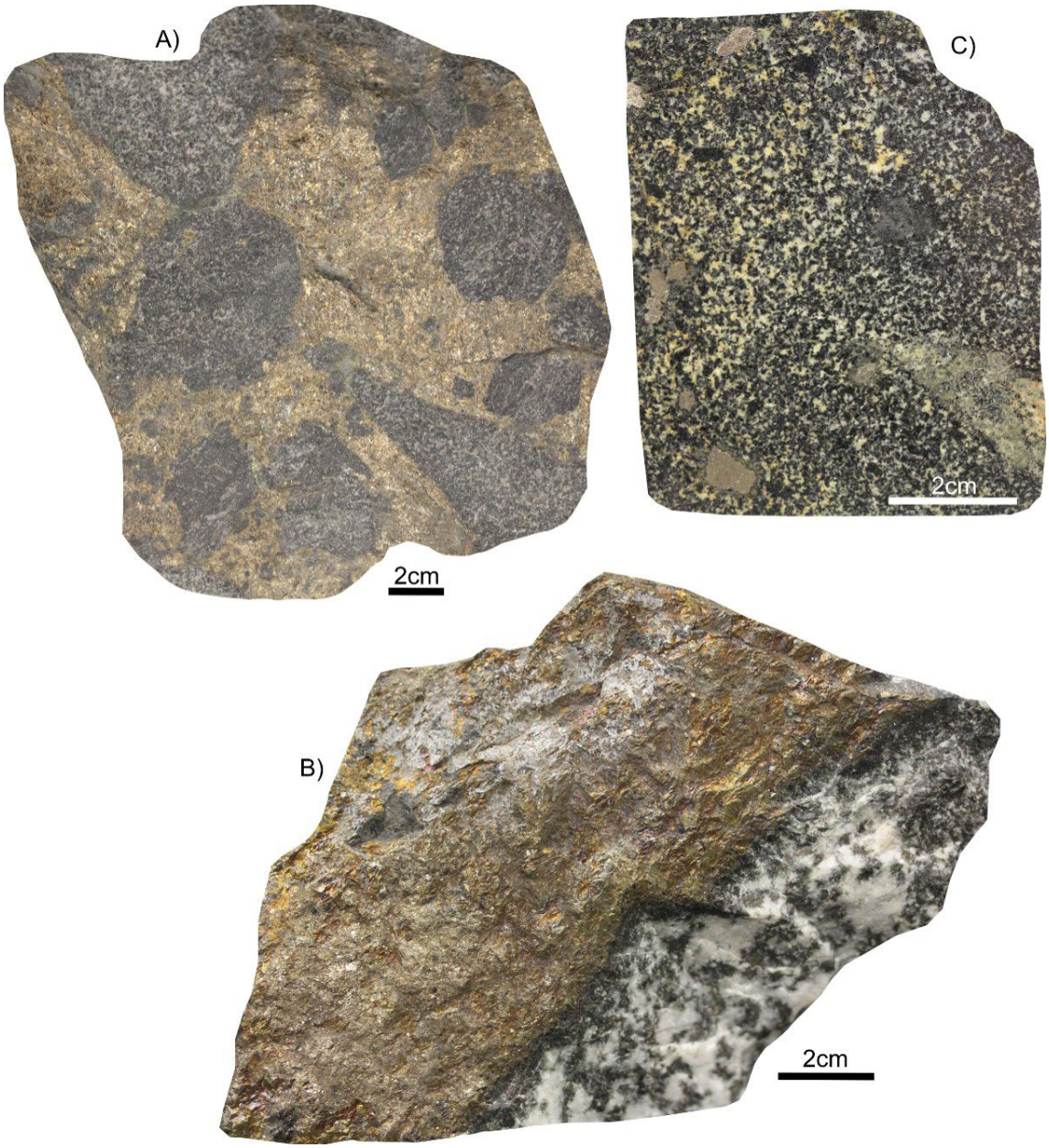


Figure 3.2: A) Inclusion-bearing massive sulfide from the Stobie Mine (contact deposit). **B)** Sharp-walled Cu vein in contact with tonalitic country rock from the Strathcona mine (footwall deposit). **C)** Disseminated sulfide blebs in variable altered inclusion-bearing quartz diorite from the Copper Cliff North Mine (offset deposit). Courtesy of Harquail School of Earth Sciences (Laurentian University).

Contact-Footwall Environment

Contact ores occur discontinuously along the SIC-footwall contact and are typically hosted by SLNR and FWBX (also referred to as Granite Breccia or Late Granite Breccia on the North Range). Mineralization typically occurs in funnels (e.g., Whistle, Foy, Copper Cliff), troughs (e.g., Creighton), and embayments (e.g., Levack) (Fig. 3.3) along the basal contact of the SIC, and are subeconomic or absent outside of those features. Mineralization typically grades downward from sparse disseminated sulfides in overlying Main Mass norite through fine and coarse (blebby) disseminated sulfides in Sublayer to semi-massive sulfides in Footwall Breccia (Fig. 3.2A) (see review by Lightfoot, 2016). Contact ores contain a typical magmatic

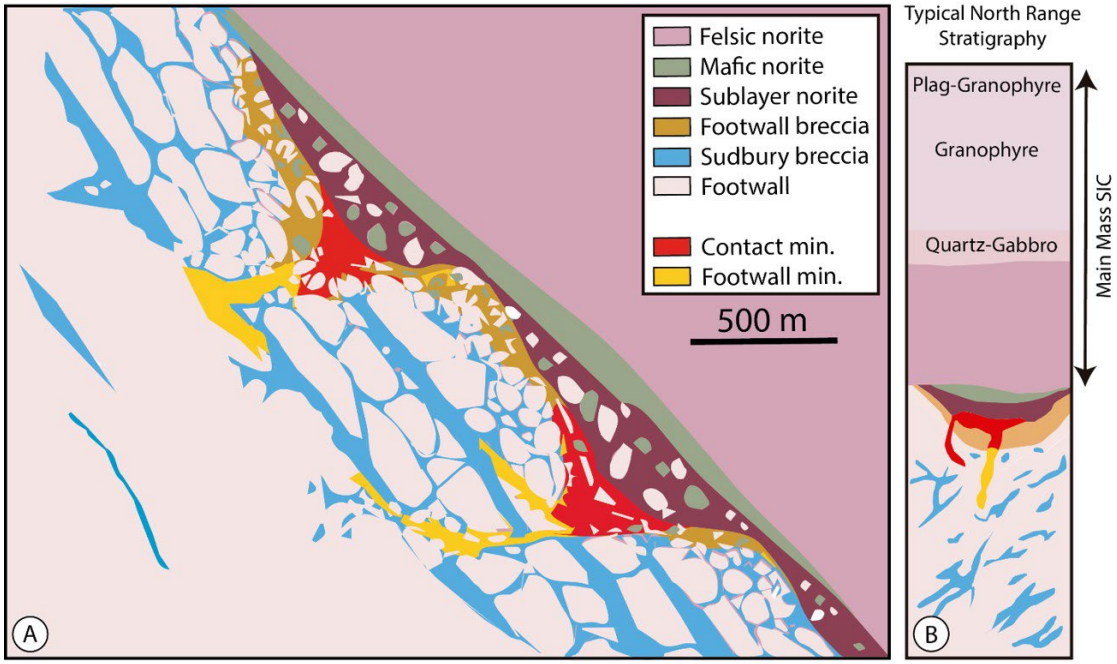


Figure 3.3.: Distribution of contact, footwall, and offset dike deposits and occurrences in the Sudbury Impact Structure. Note that locations for deeper ore bodies are projected to surface. Simplified *after* Ames et al. (2008).

sulfide assemblage of pyrrhotite (containing up to 1% Ni) > pentlandite > chalcopyrite with minor magnetite and platinum-group minerals (PGMs) (see review by Ames et al., 2008). Ore tenors (metals in 100% sulfides) vary widely depending on the composition of the magma, magma:sulfide mass ratio (R factor: Naldrett et al., 1979), and degree of MSS fractionation (e.g., Li and Naldrett, 1994), but typically range from 3.9–6.1% Ni100, 1.3–7.1% Cu100 and 0.7–5.6 ppm (Pd100+Pt100) (Naldrett, 2004). Some contact deposits in the South Range (e.g., Thayer Lindsley: Bailey et al., 2004; Garson: Mukwakwami et al., 2014) are faulted, sheared, and remobilized.

Footwall ores appear to occur only below contact deposits, up to 700m (Golightly, 2009) but more typically up to 200–300m (Farrow and Lightfoot, 2002) from the base of the SIC (Fig. 3.3). They are common in the North and East Ranges, but rare in the South Range (Fig. 3.1). Farrow et al. (2005) discriminate three types of footwall mineralization: 1) sharp-walled veins that can reach up to several meters in thickness with predominant chalcopyrite and lesser pentlandite, millerite, and cubanite grading downward and outward into bornite ± millerite veins (Fig. 3.2B), 2) disseminated sulfides often with high PGE/S ratios, and 3) a hybrid type containing both mineralization styles. Ore tenors typically range from 3.5–8.7% Ni100, 28.8–38.3% Cu100, and 13.4–33.5 ppm (Pt100+Pd100) (Naldrett, 2004). A gradual transition from Fe-Co-(Ni)-IPGE-rich contact mineralization to Cu-(Ni)-PPGE-Au-rich footwall mineralization at several deposits (e.g., Frood: Hawley, 1965; Strathcona: Li and Naldrett, 1994; McCreedy East: Gregory, 2006; Levack-Morrison: Nelles, 2012; Nickel Rim South: Glencore Ltd., unpubl.; Podolsky, KGHM unpubl.) is consistent with fractional crystallization and accumulation of Fe-Co-IPGE-rich monosulfide solid solution (MSS) in contact ores and segregation of Cu-rich intermediate solid solution (ISS) and Cu-PPGE-Au-rich residual sulfide liquid in footwall ores (e.g., Mungall, 2007). The formation of bornite-millerite ores across a thermal divide in the Fe-Cu-S system appears to require reactions with wall rocks (Nelles, 2012; see also Lesher, 2017). The formation of distal disseminated PGE-Au rich sulfides appears to require deposition from hydrothermal fluids (e.g., Hanley et al., 2004; Stout, 2009).

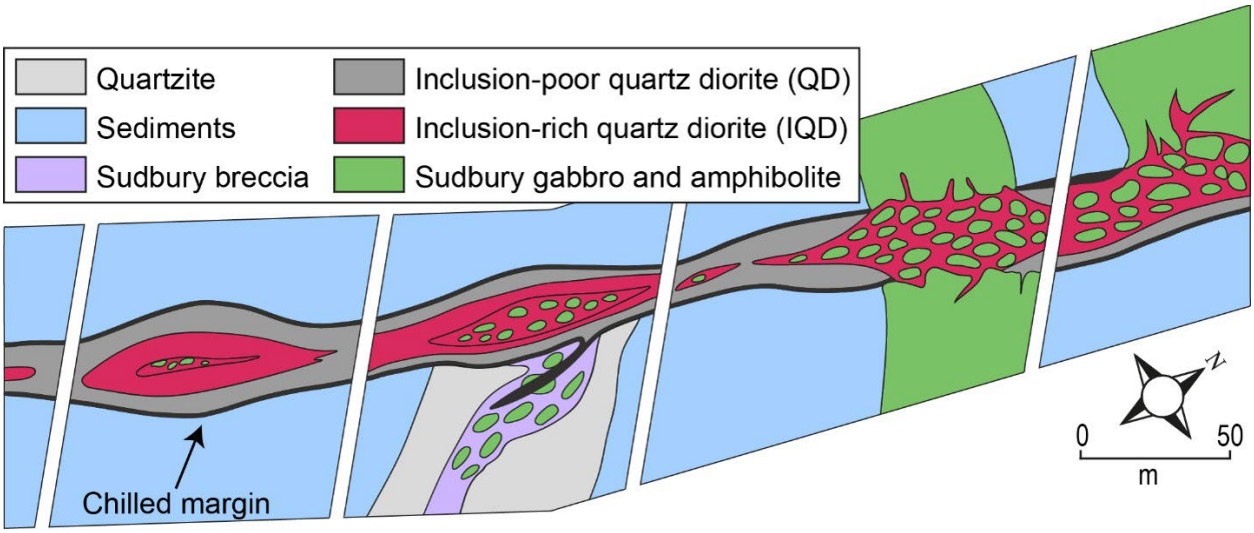


Figure 3.4: Schematic drawing of typical relationships between marginal QD and interior (mineralized) IQD) in offset dikes as well as several footwall-offset dike relationships in the Worthington offset dike. After Lightfoot and Farrow (2002).

Offset Dike Environment

Offset dikes are sub-vertical radial or concentric quartz monzodioritic lithologies (historically referred to as quartz diorite) that extend up to 20 km into the underlying footwall rocks from 300–500m-wide funnels at the contact with Main Mass (Fig. 3.1; see review by Lightfoot, 2016). The funnels typically contain SLNR ± MTBX ± IQD-QD pods and grade downward/outward into inclusion- and sulfide-poor Quartz Diorite (QD) margins with inclusion- and sulfide-rich Quartz Diorite cores (IQD; Fig. 3.4; e.g., Pattison, 1979; Grant and Bite, 1984; Lightfoot and Farrow, 2002; Murphy and Spray, 2002; Tuchscherer and Spray, 2002). Inclusions comprise angular to subrounded xenoliths derived from local country rocks, anteliths derived from the offset dike lithologies (i.e., QD clasts in IQD), and ultramafic xenoliths derived from deeper crustal lithologies (Wang et al., 2020). They vary in size from microscopic to tens of meters, and can reach up to 90% in volume (Grant and Bite, 1984). Sulfide contents vary from negligible to massive and are generally linked to the presence of (ultra)-mafic inclusions (Pattison, 1979).

Ni-Cu-PGE mineralization ranges from finely disseminated to blebby (Fig. 3.2C), semi-massive, and massive coarse-grained pyrrhotite with variable amounts of pentlandite and chalcopyrite in steeply plunging ore bodies (Cochrane, 1984; Farrow and Lightfoot, 2002). Ore tenors typically range from 3.2–6.5% Ni100, 2.6–12.8% Cu100, and 1.2–28.3 ppm (Pt100+Pd100) (Naldrett, 2004). Ore bodies are often associated with large (ultra)-mafic clasts (e.g., Totten, Podolsky), changes in strike (e.g., Copper Cliff), or cross-cutting faults (Cochrane, 1984). Endowment varies significantly between dikes located in the North Range and South Range, and with proximity to the SIC (Fig. 3.1). Most economic offset deposits (historic and current) are located in South Range dikes (e.g., Frood-Stobie, Copper Cliff). Small PGE-Cu-(Ni) occurrences in the recent discovered Rathbun offset dike indicate potential for untypical (footwall-style) mineralization in distal offset dikes (Kawohl et al., 2020). A gradual transition from Ni-rich contact mineralization to Cu-rich footwall mineralization in some offset deposits (e.g., Frood: Hawley, 1965) is consistent with fractional crystallization of Ni-IPGE-rich MSS to produce residual Cu-PPGE-rich sulfide liquid.

Ore Genesis

The ultimate sources of S and metals in the SIC are Fe ± Cu ± Ni-sulfide bearing Archean mafic gneisses, Huronian mafic volcanic rocks, East Bull Lake Suite intrusions, and Nipissing Suite intrusions (Lightfoot et al., 1997, 2001; Keays and Lightfoot, 2004), all of which contain significant amounts of sulfides and the latter two of which contain significant amounts of Ni-Cu-PGE mineralization (e.g., James et al., 2002; Sproule et al., 2007; Holwell and Keays, 2014).

However, that leaves two end-member models for the generation of the ores in the SIC (Fig. 3.5): 1) dissolution of Fe ± Cu ± Ni sulfides in the superheated impact melt followed by exsolution and settling during cooling (e.g., Lightfoot et al., 2001; Keays and Lightfoot, 2004) and 2) impact devolatilization of the majority of the S from the impact melt and incorporation of Fe ± Cu ± Ni sulfide xenomelts during thermomechanical erosion of footwall rocks (Lesher, 2019). The latter appears to be more consistent with very consistent Hf isotopic composition of the Main Mass and more heterogeneous Pb-S-Os isotopic compositions of the ores (see review by Wang et al., in press), indicating complete retention and homogenization of refractory Hf (Kenny et al., 2017) but significant loss of more-volatile Pb (O'Sullivan et al., 2016; Kenny et al., 2017; McNamara et al., 2017), Sb (O'Sullivan et al., 2016 GCA), Zn-Cd-Rb-Cs (Kamber and Schoenberg, 2020), and therefore also much/most of the highly volatile S-Se-Bi and significant amounts of moderately volatile Ag-Cu-Au-As (Lesher, 2019).

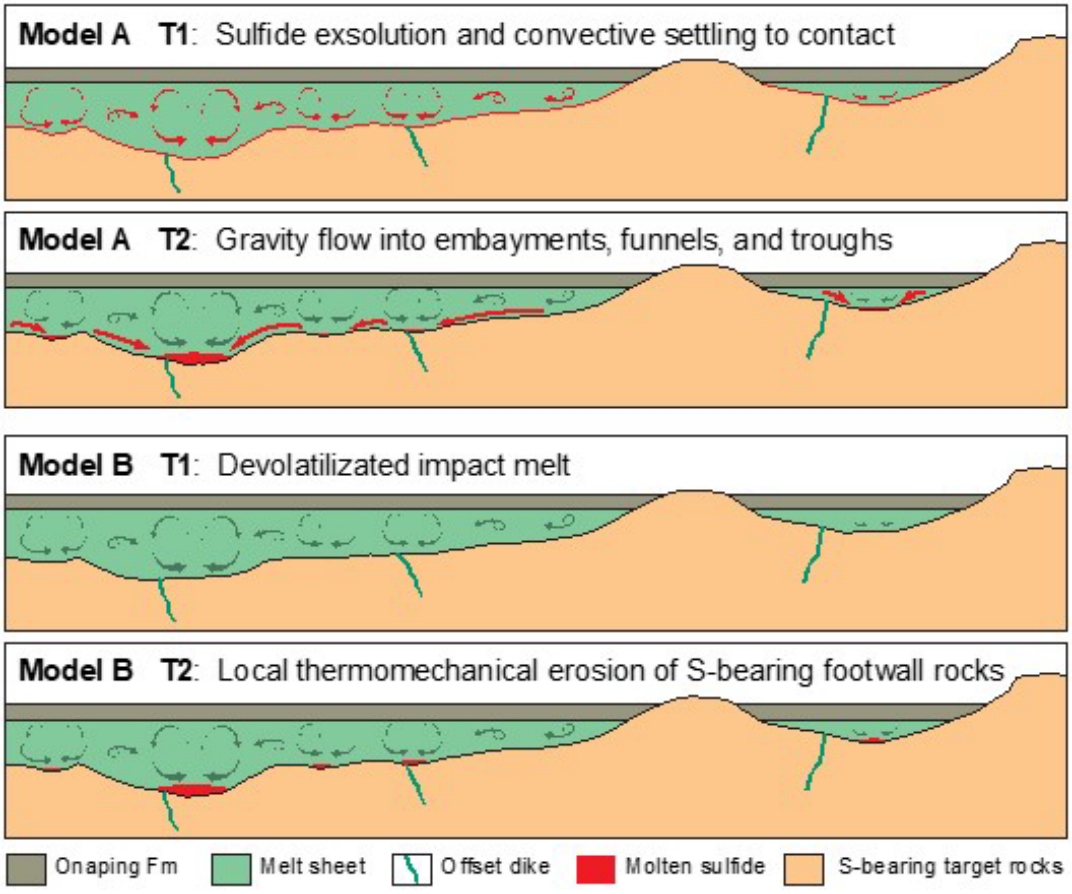


Figure 3.5: Schematic representations of end-member sulfide generation and localization models (from Wang et al., in press). **Model A:** exsolution and convective settling of molten sulfide droplets (**T1**) followed by gravity flow into embayments, troughs, and funnels (**T2**). **Model B**: volatilization of most of the S from the impact melt (**T1**) followed by generation of sulfide xenomelts by local thermomechanical erosion (**T2**).

The short interval between emplacement of inclusion- and sulfide-free QD margins and inclusion ± sulfide bearing IQD cores of nested offset dikes (<1–5 days: Wang et al., in press) provides insufficient time for exsolution and settling of sulfide droplets through the 2–5 km-thick impact melt, as dissolution, exsolution, and settling are all inherently slow processes (months: see discussion by Robertson et al., 2015, 2016), whereas the transfer of metals between the magma and sulfide droplets is much faster (hours-days: Yao and Mungall, 2021). Together, available data favor a model involving impact devolatilization of S and other volatile elements, rapid thermomechanical erosion of impact debris by the superheated impact melt, and dynamic upgrading of Fe ± Cu ± Ni sulfide xenomelts (Fig. 3.5).

4. Offset Dike Emplacement

Extensive impact melt-bearing dikes are only known from the two largest terrestrial impact structures, Sudbury and Vredefort (Dressler and Reimold, 2004; Osinski et al., 2018, and references therein). Granophyre dikes are the only remnant of an impact melt at the 2023 Ma (Kamo et al., 1996) Vredefort impact crater. They are spatially related to the centrally uplifted dome structure of the deeply eroded impact site (Reimold and Gibson, 2006). Similar to Sudbury offset dikes, the granophyre dikes have a radial and concentrical distribution, widths of 10–50m, lengths of up to 10km, crosscutting relationships with pseudotachylitic breccia, spherulitic textures, and more rarely fragment-poor granophyre margins as well as fragment-rich interiors (Reimold and Gibson, 2006; Osinski et al., 2018; Huber et al., 2022). Granophyre dikes have a homogeneous chemical composition similar to the upper continental crust which could represent the undifferentiated impact melt (Dressler and Reimold, 2004; Huber et al., 2020).

Offset Dike Characteristics

Quartz Diorite (QD) is predominantly composed of a medium-grained, homogenous matrix with acicular plagioclase, acicular sometimes radiating amphiboles (after pyroxene), variable amounts of quartz and minor biotite, granophyric quartz-alkali feldspar intergrowths, and secondary amphibole. Aphanitic and spherulitic QD margins are common in the intermediate and distal segments of offset dikes and their apophyses (e.g., Hess, Trill). Inclusion-bearing QD (IQD) is mineralogically similar to QD, but typically finer grained with a more granular texture and less pronounced acicular amphibole. Contacts between QD and IQD are often sharp, but sometimes gradational (Fig. 4.3).

QD and IQD samples from the same dike are remarkably similar in major and trace element geochemistry, indicating similar melt source and a short succession of injection (Lightfoot et al., 1997a; Pilles et al., 2017). Trace element differences between North Range and South Range dikes appear to reflect incorporation of different amounts of local footwall rocks (Lightfoot et al., 1997a).

Mechanisms of Emplacement

Two general mechanisms have been proposed to explain the geochemical, petrological, and spatial characteristics of QD and IQD. Some authors (e.g., Lightfoot and Farrow, 2002; Riller, 2005, Prevec and Büttner, 2018) prefer a multi-phase emplacement of two or more melts (Fig. 4.1), whereas others (e.g., Grant and Bite, 1984; Pilles et al, 2018) favor a single injection and flowage differentiation (e.g., Barrière, 1976) to explain the characteristics of some dikes (e.g., Foy).

In a multi-phase injection model, an initial phase of sulfide-poor, inclusion-poor QD melt is followed by injection of a second core phase of sulfide-rich, IQD melt in the center of the dikes (Fig. 4.1; e.g., Lightfoot and Farrow, 2002; Prevec and Büttner, 2018). Further injections at later stages to form the more evolved Pele and Cascaden dikes are required, but they do not contain any sulfides (Pilles et al., 2018a). This model is supported by the common presence of inclusions of QD within IQD, common sharp contacts between QD and IQD (Fig. 4.3A), and the spatial relationship between marginal QD and interior IQD.

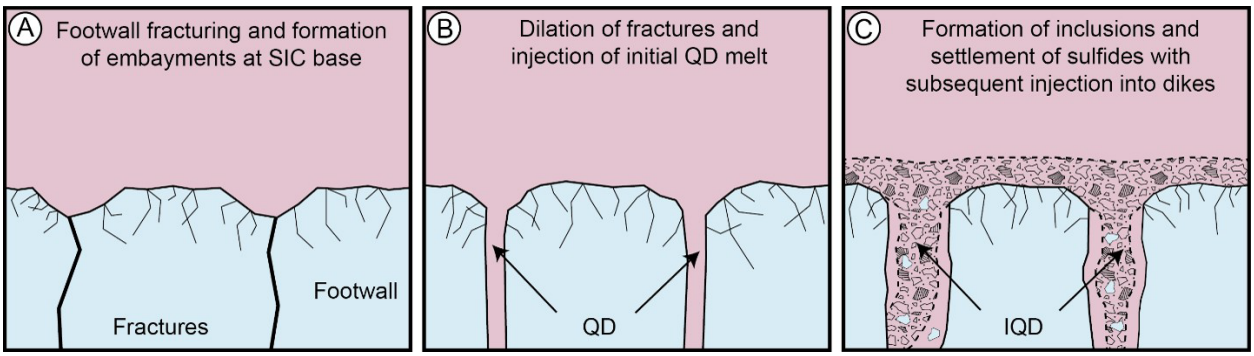


Figure 4.1: Schematic diagram of multi-stage emplacement model. Injection of an initial phase of sulfide-poor, inclusion-poor QD melt is followed by injection of a second phase of sulfide-rich IQD melt in the center of the dikes. Modified after Prevec and Büttner (2018).

In a single-stage injection model, IQD melt is injected into the fractured country rock with flowage differentiation producing marginal QD and interior IQD (Fig. 4.2). This model is supported by local gradational transitions between QD and IQD (Fig. 4.3B), rare IQD inclusions in IQD, increasing clast diameters towards the center of the dike, and clast alignments parallel to dike margins in the Foy offset dike (Pilles et al., 2018b).

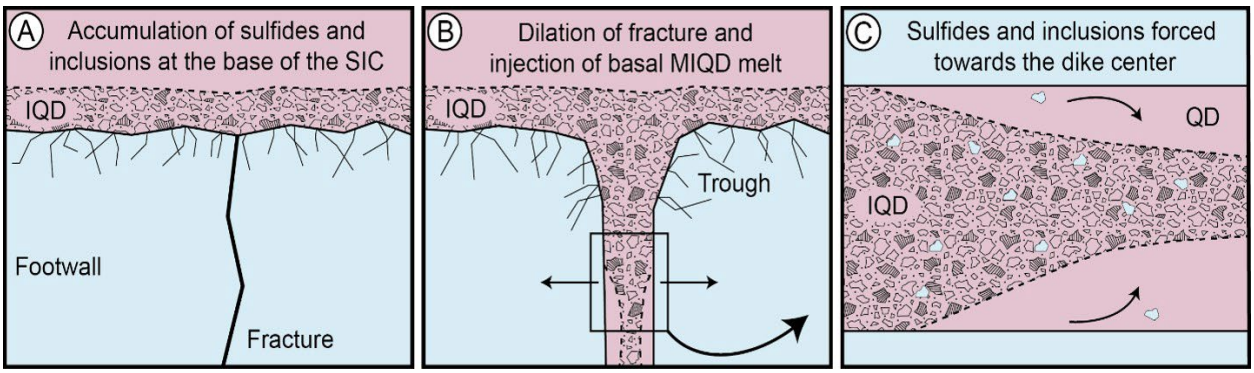


Figure 4.2: Schematic diagram of single-stage model. IQD melt is injected into dilating fractures with flowage differentiation producing marginal QD and interior IQD. Modified after Pilles et al. (2018b)

The spatial relationships between marginal QD and internal IQD can be explained by flowage differentiation or by multi-stage emplacement if IQD intruded before QD had completely solidified. However, flowage differentiation and multiple injections should produce different inclusion types and contact relationships. For example, flowage differentiation cannot easily produce the commonly observed sharp contacts between QD and IQD or the frequent inclusions of QD in IQD. In contrast, the common sharp contacts can be produced if IQD intrudes QD, the rare gradational contacts if IQD melts QD (see discussion by Huppert & Sparks, 1985), and the rare inclusions of IQD in IQD by multiple phases of injection or by radial dikes crosscutting/intruding concentric dikes. The weight of evidence presently favors a multiple injection model (Fig. 4.4 and 4.5).

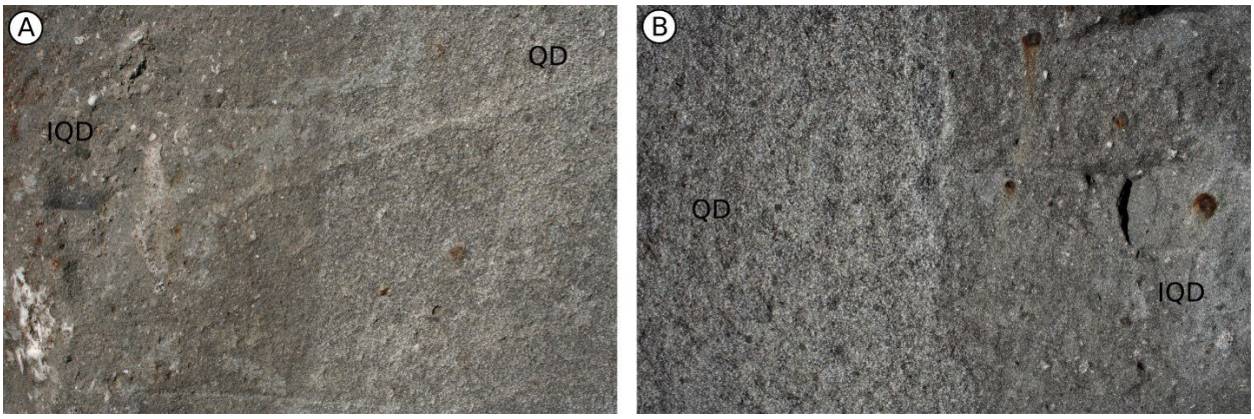


Figure 4.3: Sharp (A) and well-defined but gradational (B) contact of medium-grained QD and IQD from two stripped outcrops of the Foy offset dike. Contact is defined by grain size differences as well as abundance of inclusions and blebby sulfides. Image widths are ca. 40cm.

Timing of Emplacement

Impact cratering can be subdivided into three stages: 1) contact and compression, 2) crater excavation, and 3) crater modification (Gault et al., 1968; Osinski and Pierazzo, 2012). The timing of emplacement of the Sudbury offset dikes proposed by different workers varies from during impact excavation to tens of thousands of years after impact: 1) dilation during transient cavity formation, 1 second to 1 minute after impact (e.g., Lightfoot and Farrow, 2002; QD: Wang et al., in press); 2) dilation during rebound and central uplift, minutes to days after impact (e.g., Tuchscherer and Spray, 2002; IQD: Wang et al., in press); 3) dilation during crater wall collapse, leading to injection of melt into transfer faults (e.g., Scott and Benn, 2002); 4) dilation during isostatic uplift, up to 10,000 years after impact (e.g., Wichman and Schultz, 1993); 5) dilation after melt pressure increase as a result of a coherent roof, 1500–130,000 years after impact (e.g., Prevec and Büttner, 2018); 6) dilation during cooling and subsequent contraction of footwall rocks, >10,000 years after impact (e.g., Riller, 2005); 7) dilation during cooling of the Main Mass, 10,000s to 100,000s of years after impact (Mathieu et al., 2021); 8) dilation during readjustment and late tectonic deformation, >10,000 years after impact (e.g., Therriault et al., 2002).

1) The presence of only sparse local xenoliths and sulfides, and the presence of aphanitic, spherulitic, and radiating "quench" pyroxene textures in the distal parts of the dikes require the impact melt to have been superheated when QD was emplaced, so this precludes all of the models involving emplacement during or after crystallization of the Main Mass.
2) The nested QD/IQD relationships with no occurrences of IQD without QD (with the possible exception of the South Range Breccia Belt) require emplacement while the cores of the QD dikes were still weak, which has been modelled by to have been within 2–5 days (Wang et al., in press).
3) The presence of inclusions and sulfides in IQD require the impact melt to have reached sulfide saturation within that time interval and for IQD to have been forcibly emplaced within that interval.
4) The two events most likely to have caused rapid basin-wide sequential injection of QD and IQD are excavation and impact (Fig. 4.4).
5) The process most likely to have driven the impact melt from a superheated, sulfide-undersaturated state to a liquidus, sulfide saturated state is assimilation of fragments generated by collapse of the peak ring.

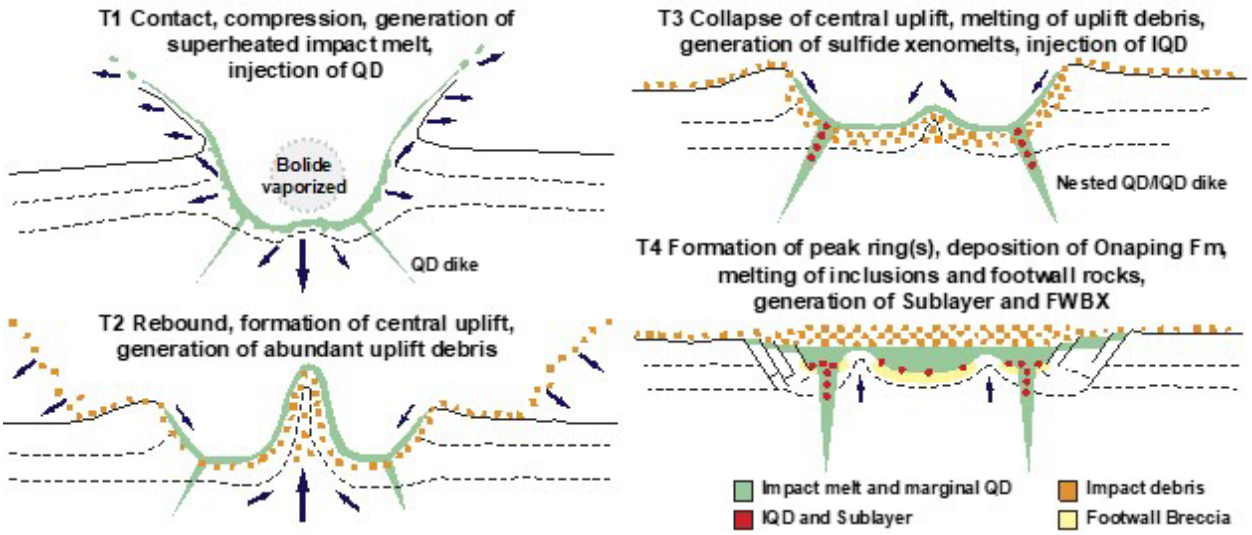


Figure 4.4: Schematic representation of complex crater-forming and -modification events leading to injection of inclusion- and sulfide-poor marginal QD (T1), generation of inclusion- and sulfide-rich IQD (T3), and melting of inclusions and generation of mineralized FWBX (T4). FWBX – Footwall Breccia, IQD – Inclusion-bearing Quartz Diorite, SLNR – Sublayer Norite, QD – Quartz Diorite. From Wang et al. (in press), as modified after Melosh (1989).

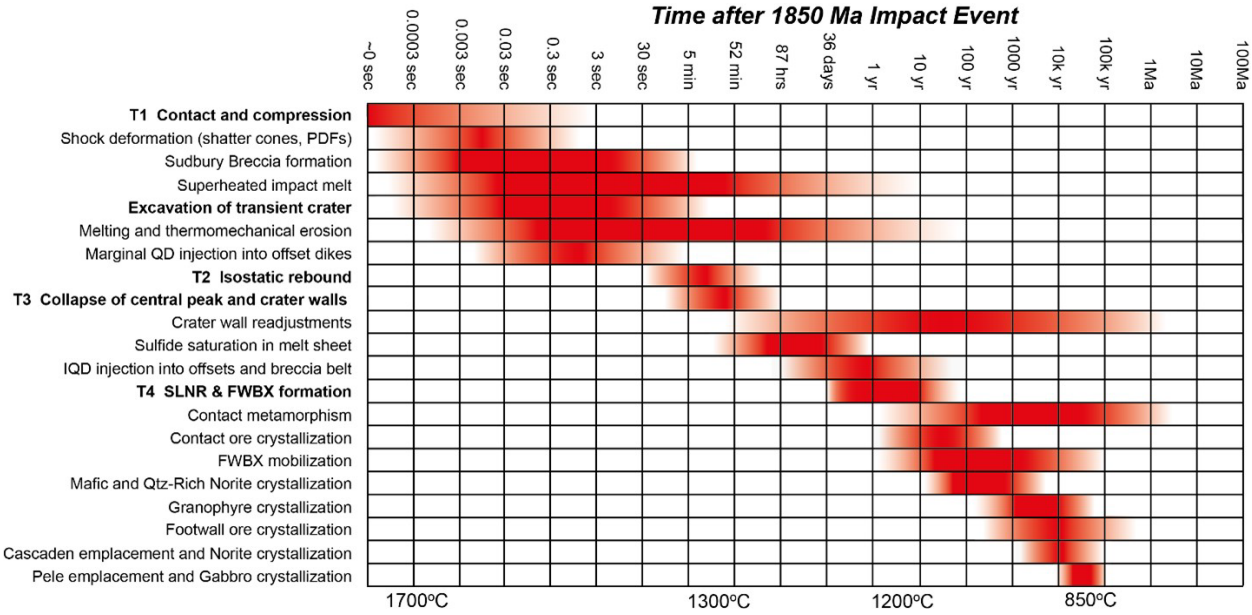


Figure 4.5: Inferred geological history of the Sudbury structure, highlighting major events related to the formation of QD, IQD and Sublayer, Footwall Breccia, and associated Ni-Cu-PGE mineralization. From Wang et al. (in press), as modified from Lightfoot, 2016).

Excursion Stops, Day 2: Ore Environments

Geology of the Whistle and Parkin Area

Footwall Rocks

The Whistle funnel and Whistle-Parkin offset dike are located at the northeastern lobe of the Main Mass of the Sudbury Igneous Complex in the Norman and Parkin townships (*see regional map* from Day 1). The footwall comprises Archean rocks of the Superior Province close to the Whistle embayment and Paleoproterozoic greenschist facies to amphibolite facies metamorphosed volcanic and sedimentary rocks of the Huronian Supergroup in the distal Parkin offset dike (Fig. 5.1; Ames et al., 2008; Lightfoot, 2016 and references therein).

With ages of 2725 to 2703 Ma (Nunes and Pykes, 1980) felsic to intermediate volcanic rocks as well as feldspar and quartz-feldspar porphyritic rocks of the Benny greenstone belt are the oldest rocks in the area. (Meyn, 1970).

Units of the Levack Gneiss Complex are not shown in Fig. 5.1, but occur as tens of meter size bodies close to the Sudbury Igneous Complex and consist of migmatitic tonalite orthogneiss, biotite paragneiss, mafic to felsic gneiss, and gabbros (Meldrum et al., 1997; Murphy and Spray, 2002). Krogh et al. (1984) established an age of 2711 Ma for leucosomes of a tonalitic orthogneiss.

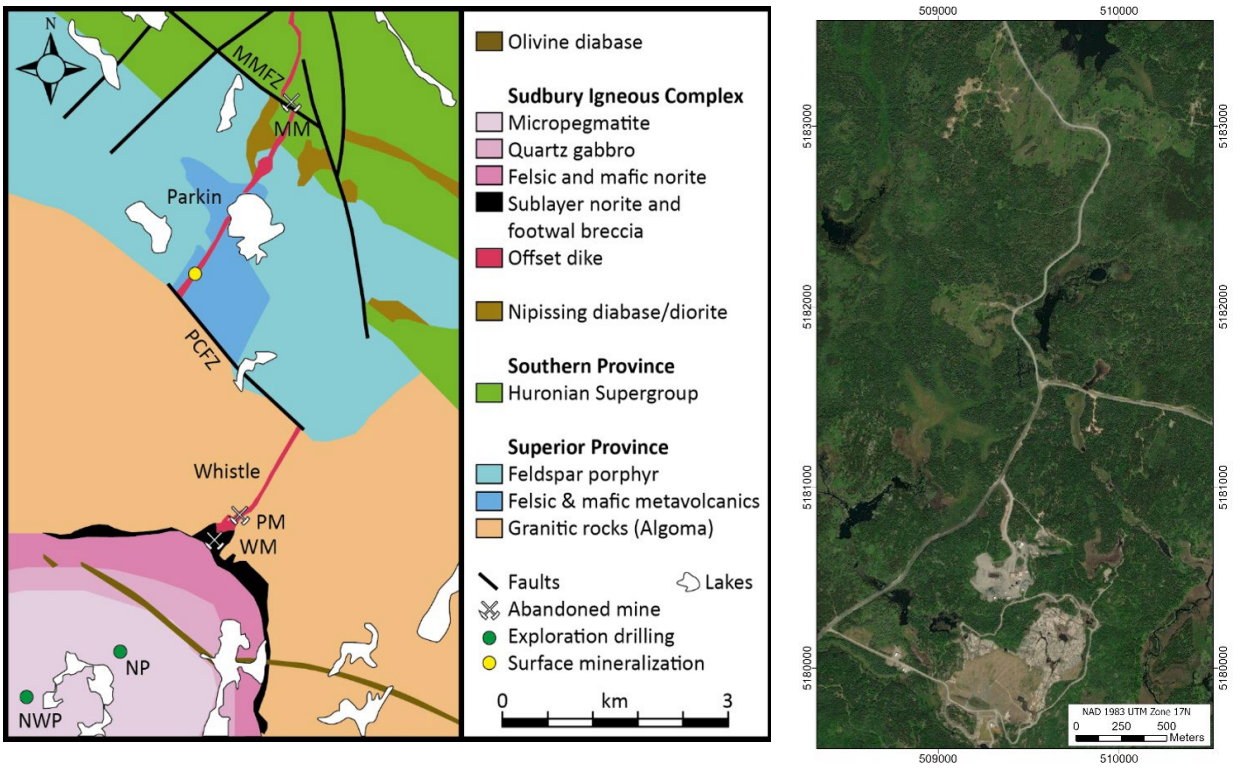


Figure 5.1: Simplified geological map of NE Sudbury after Murphy and Spray (2002). Inset: satellite image of the NE Sudbury area showing the field trip stops on the stripped outcrops at the Whistle funnel and Whistle-Parkin offset dike. MM – Milnet Mine, MMFZ – Milnet Mine Fault Zone, NP – Norman Project, NWP – Norman West Project, PCFZ – Post Creek Fault Zone, PM – Podolsky Mine, WM – Whistle Mine.

The 2642 Ma (Meldrum et al., 1997) Cartier batholith is part of the Algoma plutonic domain and consists of medium- to coarse-grained subporphyritic granite. It intruded and at least partially melted parts of the Levack Gneiss Complex (Langford, 1960; Dressler, 1984b; Ames et al., 2008), resulting in local breccia formation ("Levack breccia") which can resemble Footwall Breccia. Aphyric to plagioclase glomeroporphyritic diabase and gabbros of the 2473 Ma (Heaman, 1997) Matachewan dike swarm crosscut gneissic and granitic units (Meldrum et al., 1997).

North of the Milnet Mine fault, Huronian metasediments of the Quirke Lake and Cobalt Group dominate. In addition, diabase and gabbros of the 2.2 Ga (see Lightfoot, 2016 (p. 88) and references therein) Nipissing mafic intrusive suite occur conformable in the Huronian Supergroup.

Whistle Funnel and Contact Mineralization

The funnel-shaped Whistle embayment is located at the base of the Main Mass and is overlain by continuous layers of micropegmatite, transition quartz gabbro, and felsic norite, and a thin discontinuous layer of mafic norite (Fig. 5.2; Pattison, 1979; Lightfoot et al., 1997a; Lightfoot et al., 1997b). The dip of the contact changes from ~45 degrees south along the northern side of the funnel to ~70 degrees west along the eastern side of the funnel. First published descriptions of surface and drill core data by Pattison (1979) indicated a zonation of orthopyroxene-rich SLNR at the center of the embayment and gradually more siliceous igneous-textured SLNR matrix towards the margins. Here, the Sublayer transitions into FWBX (Fig. 5.2). In a detailed study of SLNR from the open pit, Lightfoot et al. (1997b) identified several lithologies: The central part of the embayment hosts a two pyroxenite SLNR, whereas orthopyroxene-rich SLNR, olivine norite and leucocratic norite occur at the margins or as centimeter-sized to tens of meter-sized pods within the others. Sublayer matrix is typically non-poikilitic and hosts disseminated to blebby sulfides. Inclusions can range from millimeters to meters and are described as diabase inclusions, anorthositic to gabbroic inclusions and melanorite inclusions or segregations with gradational contacts (Lightfoot et al., 1997b).

The Whistle mine was operated by INCO Limited (now Vale) from 1988–1991 and 1994–1997, producing 5.7 Mt of ore grading 1% Ni and 0.3% Cu (Carter et al., 2009). Inclusion-rich massive sulfides (2–3% Ni, >0.2% Cu, <500ppm Pt+Pd) occur at the Sublayer-Footwall breccia contact and show a fractionation to more Cu-rich ore towards the base of the embayment (Lightfoot et al., 1997b).

Whistle Dike and Footwall Mineralization

After the Whistle Mine closure, acid-generating waste rocks that had been stored to the northeast of the embayment were mechanically and hydraulically stripped during the backfilling of the open pit, allowing subsequent detailed mapping and studies of the proximal Whistle offset dike (Fig. 5.2). The main outcrop was mapped in 2003 by FNX Mining Company (now KGHM), who were exploring in the proximal Whistle dike for Footwall-type mineralization, as well as locally in more detail by Carter et al. (2009) in 2003, and the Podolsky North Zone by Lafrance et al. (2014) in ~2008–2009 during petrographic-geochemical studies of the "metabreccias" and quartz diorite lithologies.

The Whistle dike emerges from the funnel and can be traced for 2km to the northeast where it terminates in a breccia zone at the Post Creek fault (Fig. 5.1; Lightfoot et al., 1997b). MTBX is the dominant lithology with meter- to tens of meters-sized inclusion-poor and inclusion-rich quartz diorite pods more rarely occurring close to the dike margins. Magnetic lineation measurements suggest lateral emplacement with locally downward-directed flow for the dike lithologies and subsequent downward-directed sinking of massive sulfides (Giroux and Benn, 2005).

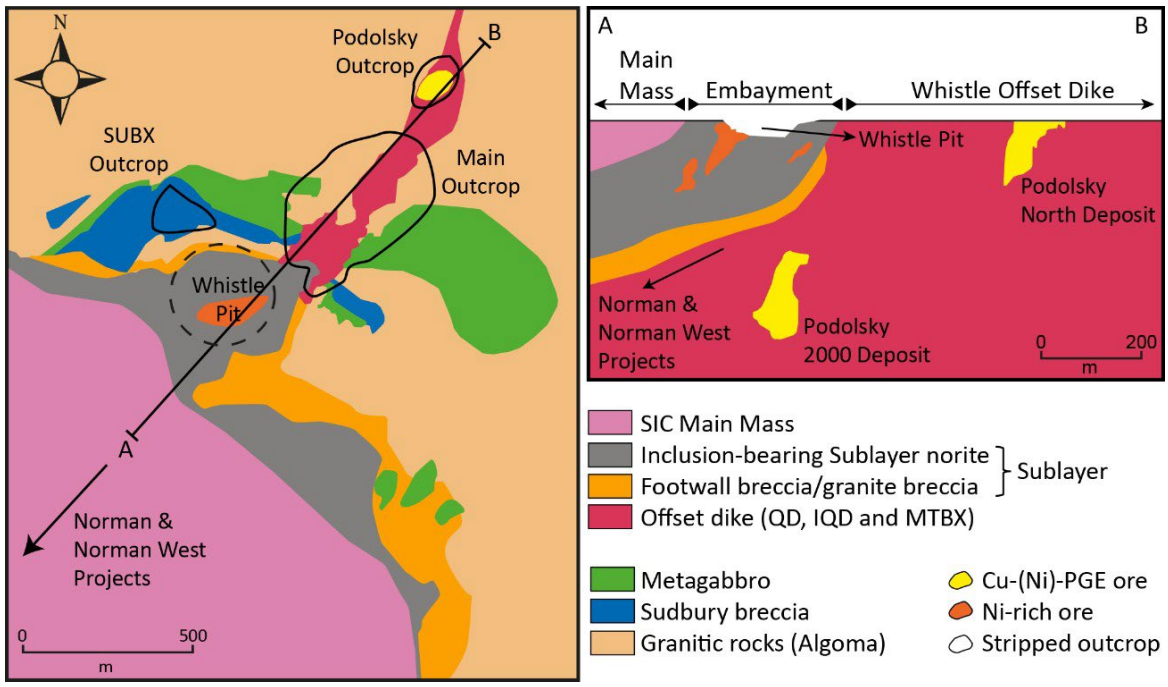


Figure 5.2: Simplified geological map and cross-section of the funnel-shaped Whistle embayment and proximal Whistle dike (after Farrow et al., 2005).

Two Footwall-style Cu-(Ni)-PGE ore bodies occur within MTBX and IQD in the proximal segment of the Whistle dike: the Podolsky 2000 Deposit, which occurs at depth near the northwestern margin of the dike, and the Podolsky North deposit, which extends to the surface northeast of the main outcrop (Fig. 5.2). Mineralization in the Podolsky 2000 deposit occurs in the form of disseminated/blebby sulfides and "low-sulfide" stockwork veins in the host rocks, as well as breccia sulfide veins and "sharp-walled" massive sulfide veins in a large gabbroic inclusion (Farrow et al., 2005). The latter mineralization appears to cut the others. Chalcopyrite and millerite are the dominant Cu and Ni ore minerals. 2.24 Mt ore at 4.2% Cu and 0.4% Ni were mined between 2008–2013 (Lightfoot, 2016).

A stripped outcrop (field trip stop 2-C1; Fig. 5.1) of the distal Whistle segment located just south of the Post Creek fault indicates that metabreccia with disseminated sulfides is still the dominant lithology.

Parkin Dike and Offset Dike Mineralization

The Parkin segment of the offset dikes appears north of the Post Creek fault, 2km displaced from the Whistle segment, and can be traced for another 12km to the northeast (Fig. 5.1). It is hosted by units of the Archean Benny greenstone belt for the first 4km and by the metasediments and metavolcanics of the Huronian Supergroup past the Milnet Mine fault.

Exploration by Wallbridge Mining in 2014–2015 led to the stripping of several large outcrops of the offset dike close to the Post Creek fault. Here, QD and IQD are the common offset dike lithologies and MTBX occurs mainly as pods or inclusions in the quartz diorite lithologies (Anders et al., 2020) or as parallel bodies adjacent to the IQD (Murphy and Spray, 2002). QD is commonly located at the margins of the dike and in sharp contact with the interior IQD.

Disseminated sulfides and sulfide stringers are mainly associated with inclusion-bearing lithologies in the dike center.

Rock Descriptions

The most common units/lithologies in the field area are listed in Table 5.1 with short descriptions, the most important of lithologies are discussed in more detail below.

Table5.1: Rock units and lithologies to be encountered on the Field Trip.

Unit/Lithology	Abbreviation	Description
Levack Gneiss Complex	MGN, IGN, and FGN	Medium- to fine-grained, weakly to strongly foliated, mafic (MGN), intermediate (IGN) and felsic (FGN) gneisses, gabbros and migmatites
Cartier Granite	GR	Medium- to coarse-grained, alkali feldspar-megacrystic granites to granodiorites and equigranular monzogranites
Matachewan Intrusives		
Gabbro	GAB	Medium- to fine-grained, green, magnetic gabbro and diorite bodies intruding other footwall rocks
Diabase	DIA	Fine- to medium-grained, northwest striking, plagioclase glomeroporphyritic dikes intruding other footwall rocks
Impact-Related Rocks		
Sudbury Breccia	SUBX	Polymictic, matrix-supported breccia with aphanitic to very fine-grained, black to dark grey groundmass that supports subrounded heterolithic footwall fragments (millimeter to tens of meters)
Quartz Diorite	QD	Medium-grained, leucocratic, homogeneous, igneous textured granodioritic matrix with tabular to acicular and sometimes radiating amphibole (after pyroxene) laths; interpreted to represent variably contaminated impact melt
Magmatic Impact-Related Breccias		
Inclusion Quartz Diorite	IQD	Polymictic, homogeneous, matrix-supported breccia with fine-grained, grey to "salt-and-pepper" igneous-textured groundmass; local and exotic (mainly ultramafic) inclusions; often with <2% disseminated sulfides
Sublayer Norite	SLNR	Polymictic, homogeneous, matrix-supported breccia with fine- to medium-grained, subophitic, noritic groundmass; abundant local and exotic (mainly ultramafic) inclusions; often with <5% disseminated sulfides
Metamorphic-Anatectic Impact-Related Breccias		
Footwall Breccia (Granite Breccia)	FWBX (GRBX)	Polymictic, heterogeneous, matrix-supported breccia with fine-grained, pinkish white, granitic groundmass; subrounded inclusions of local footwall rocks and SUBX
Metabreccia	MTBX	Polymictic, heterogeneous, matrix- to clast-supported breccia with fine-grained dark grey to pinkish-grey, recrystallized groundmass; >50% inclusions of local footwall rocks ± pods of QD and IQD
Late Dikes		
Olivine diabase	UM	East-west trending, up to 7m wide, dark, fine-grained dikes cross-cutting all other lithologies in the area

Breccia Classification

Pre-impact Archean Breccia ("Levack Breccia")

Intrusion of the 2642 Ma Cartier Batholith brecciated many parts of the Levack Gneiss Complex (Card and Innes, 1981; Dressler, 1984b; Meldrum et al., 1997). Levack Breccia is characterised by a medium- to coarse-grained, locally alkali feldspar-megacrystic, granodioritic to monzogranitic matrix with angular to rounded, cm- to tens of meter-sized mafic, gneissic, and migmatitic inclusions (Fig. 5.3A). The distribution and abundance of Levack Breccia is not well known, but there are several areas on the Whistle outcrop where this breccia is present.

Sudbury Breccia

Pseudotachylitic SUBX appears to be the earliest formed impact breccia as it is crosscut by all other impact-related lithologies. Shock compression and cataclasis during crater formation led to shattering, pulverization, and frictional melting of footwall rocks, forming centimeter- to meter-sized veins and more irregular bodies of intense brecciation, often at structurally weakened zones and at lithology boundaries (e.g., Speers, 1957; Dressler, 1984a; Rousell et al., 2003; Lafrance et al., 2008; Lafrance and Kamber, 2010). It is characterized by a black to dark grey, aphanitic to very fine-grained matrix with (sub)-rounded centimeter- to meter-sized clast of immediate country rock lithologies (Fig. 5.3B). Sudbury breccia occurs frequently in the vicinity of the SIC but has been described up to distances of 50 km or more (Dressler, 1984a). Close to the SIC, contact metamorphism led to a grain size increase (often accompanied by lighter matrix colours), growth of biotite porphyroblasts, and partial melting of felsic mineral clasts.

Inclusion-Bearing Quartz Diorite

IQD is a polymictic breccia constraint to the offset dikes. It typically consists of a homogeneous, equigranular, fine- to medium-grained groundmass with a "salt-and-pepper" appearance (Fig. 5.3C). Inclusion sizes and abundancies vary from millimeter to tens of meters and <10% up to 90%, respectively. The matrix typically consists of tabular plagioclase with lesser alkali feldspar, quartz, and amphiboles. Radial amphiboles, which are typical for QD, occur less frequent in IQD.

Sublayer Norite

SLNR forms a discontinuous layer at the base of the Sudbury Igneous Complex normally occurring within funnels, troughs, and embayments. It is a variably mineralized polymictic breccia with centimeter- to meter-size inclusions of Ol melanorite anteliths, local xenoliths, and exotic ultramafic xenoliths set in a fine- to medium noritic matrix (Fig. 5.3D; e.g., Pattison, 1979; Naldrett et al., 1984; Wang et al., 2018, 2020).

Footwall Breccia

FWBX (also termed "leucocratic breccia" and "late granite breccia") is a polymictic breccia containing inclusions of local footwall rocks, SUBX, Main Mass norite, and exotic ultramafic inclusions (e.g., Pattison, 1979; Coats and Snajdr, 1984; Dressler, 1984a; Lakomy, 1990; McCormick et al., 2002; Wang et al., 2020). Grain sizes, compositions, and textures of the matrix are highly variable and dependent on proximity to the SIC contact and mineralization: the matrix is coarser-grained, igneous-textured and of dioritic to tonalitic composition closer to the SIC (Fig. 5.3E) and finer-grained, metamorphic-textured (recrystallized) and of granitic composition further into the footwall (Lakomy, 1990; McCormick et al., 2002). In proximity to mineralization, FWBX is often characterized by a grey colour (Greenman, 1970).

Several modes of formation have been proposed, including a) contact metamorphism of impact-brecciated footwall during cooling of impact melt (Dressler, 1984a), b) partial melting and anatexis of felsic layers in the Levack Gneiss Complex due to heat conduction of cooling impact melt (Coats and Snajdr, 1984), and c) contact metamorphism during injection of Sublayer norite (Pattison, 1979).

Metabreccia

The term "metabreccia" was first used by INCO mining geologists in the 1970's to describe thermal metamorphism of SUBX in proximity to the superheated melt sheet (E.F. Pattison, 2019, pers. comm.), which has been adopted by some of the other mining companies (e.g., Poulin et al., 2009). Other workers have interpreted MTBX to represent variable amounts of impact melt (QD/IQD) and partially melted footwall rocks (FWBX) mobilized into the dike structure (Murphy and Spray, 2002; Giroux and Benn, 2005; Lafrance et al., 2014; Carter et al., 2009; Anders et al., 2020).

Farrow et al. (2005) were the first to introduce "metabreccia" as a term in the published literature, but different terminology exists by further authors, such as diatexite, metatexite (Lightfoot, 2016), radial breccia, mafic sulfide-bearing breccia (both Murphy and Spray, 2002), metamorphic leucocratic breccia/Footwall Breccia (Carter, 2005; Carter et al., 2009), and recrystallised Footwall Breccia (Grant and Bite, 1984). Research on metabreccia (or its synonyms) is restricted to the Whistle funnel and Whistle-Parkin offset dike, Ministic, Foy and Trill offset dikes – all located in the North Range.

Metabreccia is a heterogeneous, grey pinkish, polymictic breccia with abundant millimeter- to centimeter-sized and lesser meter-sized footwall inclusions (Fig. 5.3F). Textures and mineralogy vary from igneous-textured with similar modal abundancies as QD and IQD (Lafrance et al., 2014) to dynamically recrystallized with a fine-grained quartz-feldspar-rich matrix (Anders et al., 2020). MTBX is the dominant lithology in the Whistle dike, whereas it occurs as pods in quartz diorite in Parkin.

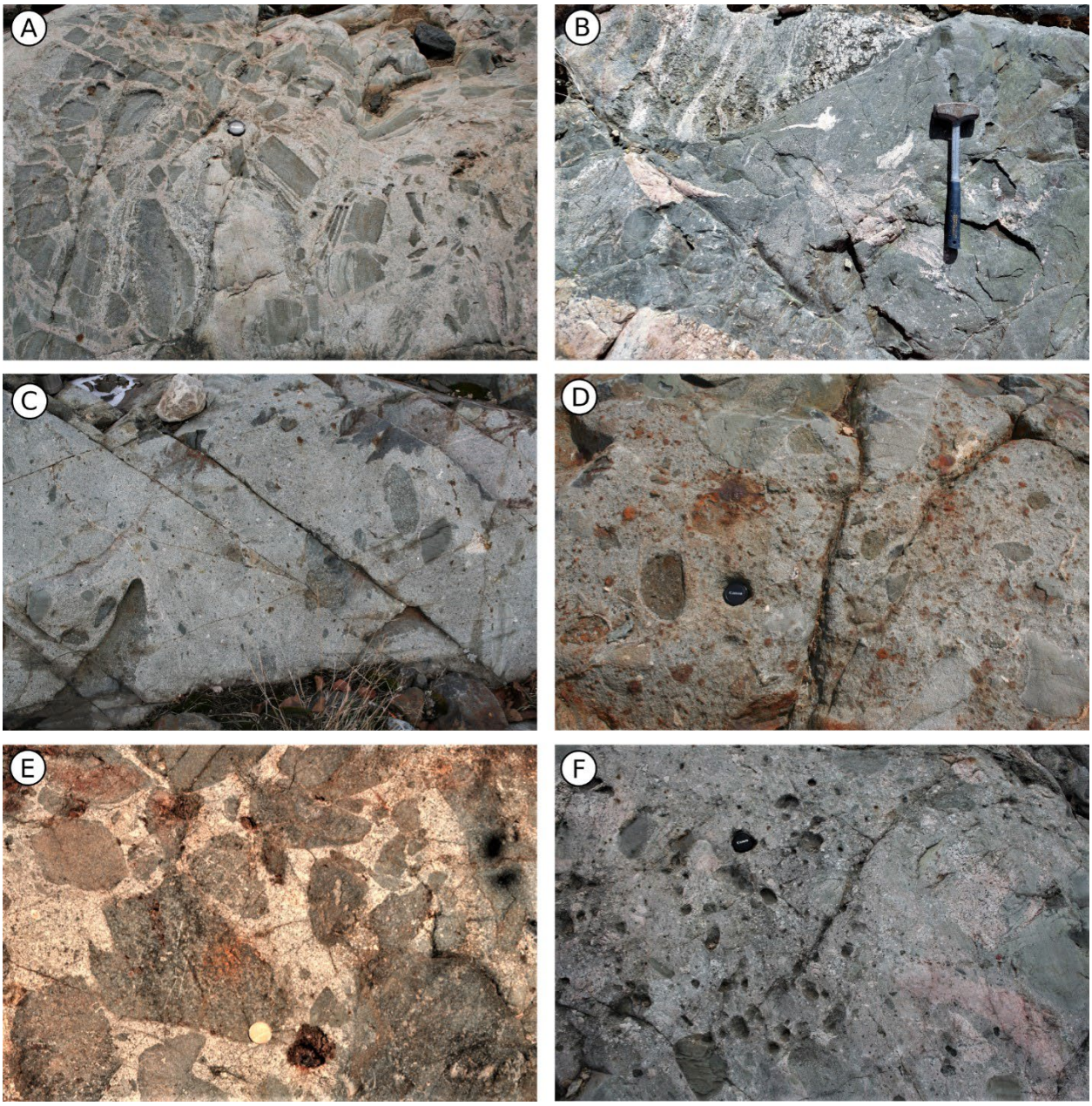


Figure 5.3. A) Levack breccia near Whistle funnel. **B)** Metamorphosed SUBX near Whistle funnel (field trip stop 2-A2. **C)** IQD in Whistle funnel (field trip stop 2-A4). **D)** SLNR in Whistle funnel (field trip stop 2-A3). E) FWBX near Coleman Mine**. F)** MTBX in Whistle offset dike.

A. Whistle Embayment and Proximal Offset Dike

The stripped outcrops of the Whistle funnel provide excellent exposure and a wealth of interesting field relationships and textures. Key aspects are highlighted on the satellite image (Fig. 5.4) and on the geological map (Fig. 5.5), and are described in more detail below.

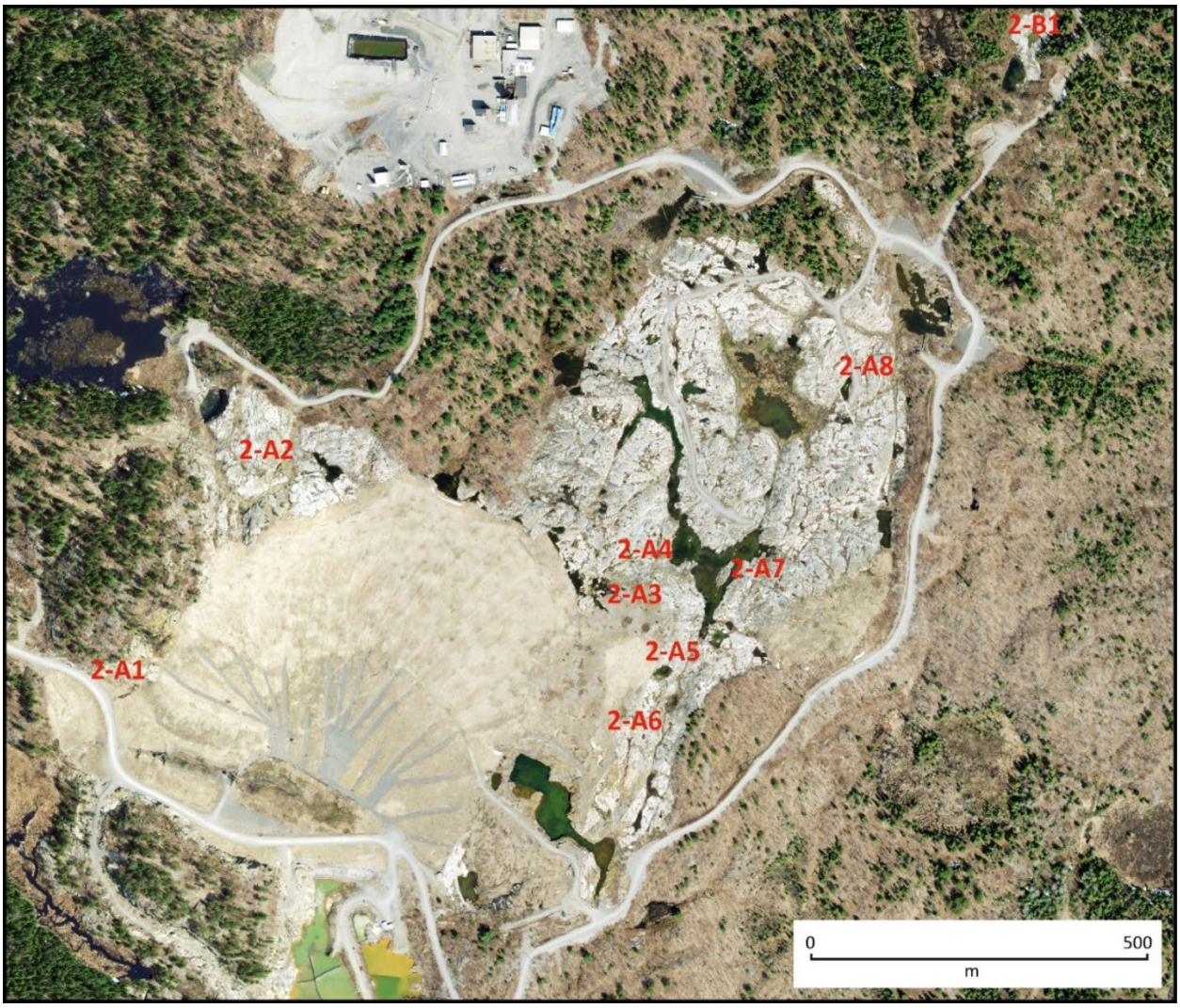


Figure 5.4: Satellite image of the filled and reclaimed Whistle open pit (prior to re-greening stage), showing field trip stops on the associated stripped outcrops. The Podolsky Mine is in the upper right part of the image.

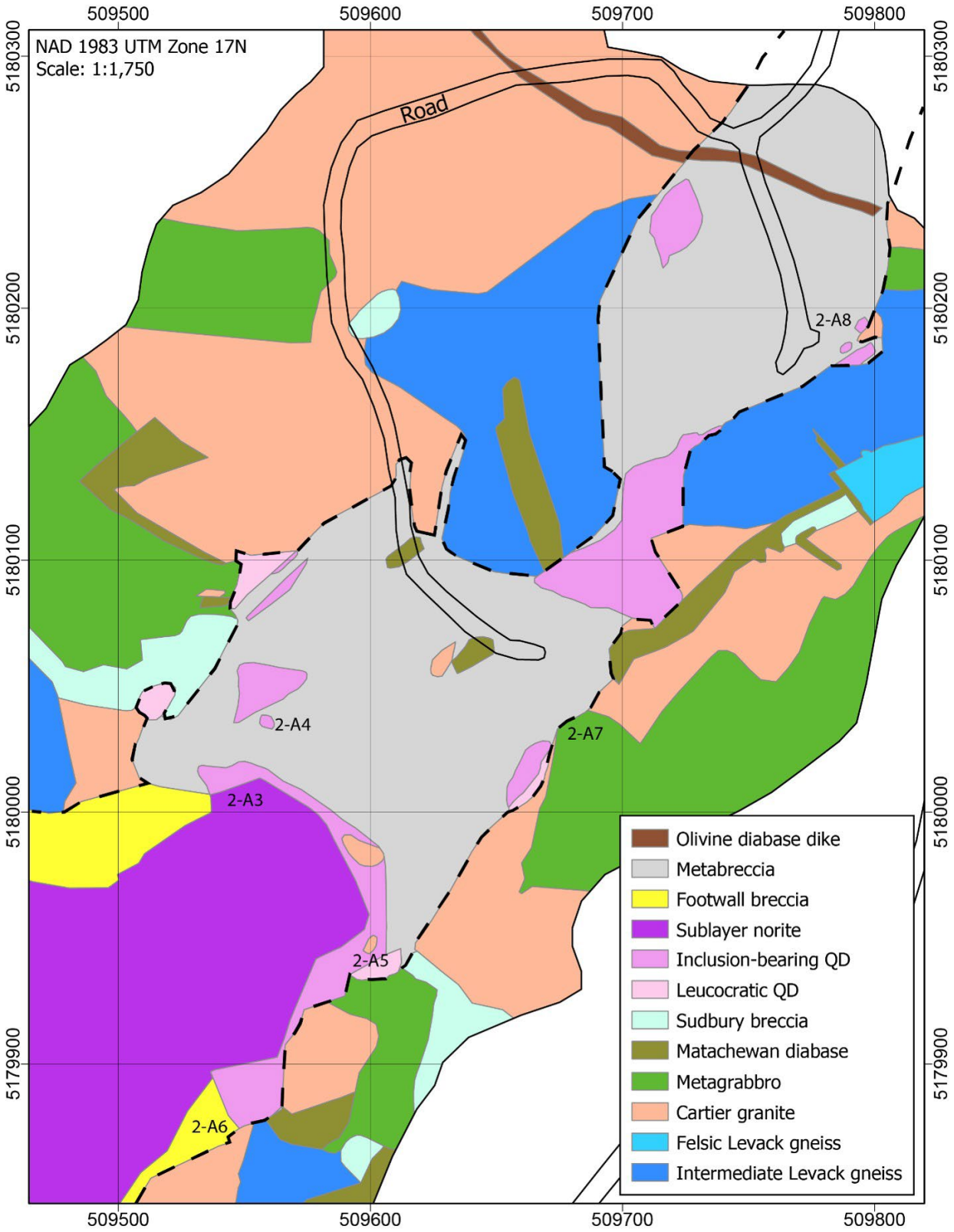


Figure 5.5: Map of the Whistle outcrop (simplified from FNX Mining Company) showing field trip stops. Dashed lines delineate the footwall-funnel/dike contact.

2-A1) SLNR-GRBX contact (NAD83 17N, 509053, 5179920)

The first stop is located at the western flank of the embayment near the reclaimed open pit. SLNR is in contact with FWBX over a few centimeters. SLNR has a fine- to medium-grained, dark grey to black matrix with minor rugged sulfide blebs and some subrounded inclusions (Fig. 5.6A–C). FWBX has a pinkish grey, fine- to medium-grained, heterogenous matrix with abundant strongly intergrown quartz-feldspar and variable amounts of interstitial green amphibole. Inclusions are typically subangular and of mafic to intermediate composition (Fig. 5.6D–F).

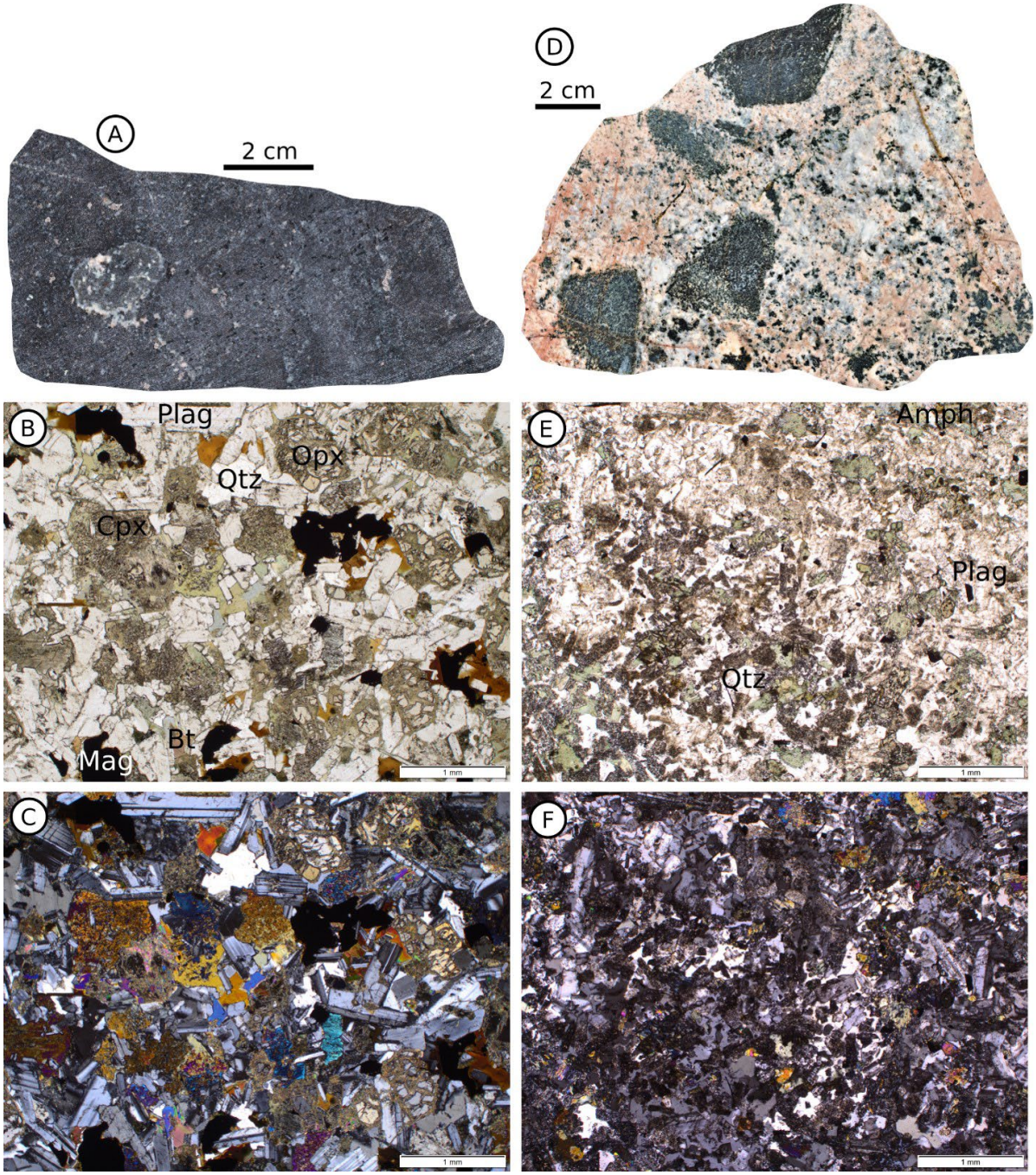


Figure 5.6: High-resolution sample scans, plane-polarized and cross-polarized images of SLNR (**A-C**) and FWBX (**D-F**) from the eastern funnel flank.

2-A2) SUBX and footwall rocks (NAD83 17N, 509289, 5180141)

This 170m x 120m stripped outcrop is located in the footwall just west of the former open pit. Meter- to tens of meters-scale blocks of various country rocks are enveloped and truncated by cm- to dm-wide pseudotachylitic SUBX veins and bodies (Fig. 5.7A). Walking from the NW end of the outcrop towards the embayment (ESE direction), a slight gradational change of color, grain size and clast angularity are noticeable in the SUBX veins and might reflect a baking of the breccia closer to the former impact melt. The matrix changes from a black/dark grey to a medium grey, coarsens slightly, biotite porphyroblasts are observable and small felsic clasts are getting wispier closer to the former impact melt.

Small SUBX bodies are intruded by a pinkish medium- to coarse-grained feldspar-rich granitoid in some areas, which might indicate the formation of localized feldspar-rich FWBX (Fig. 5.7B).

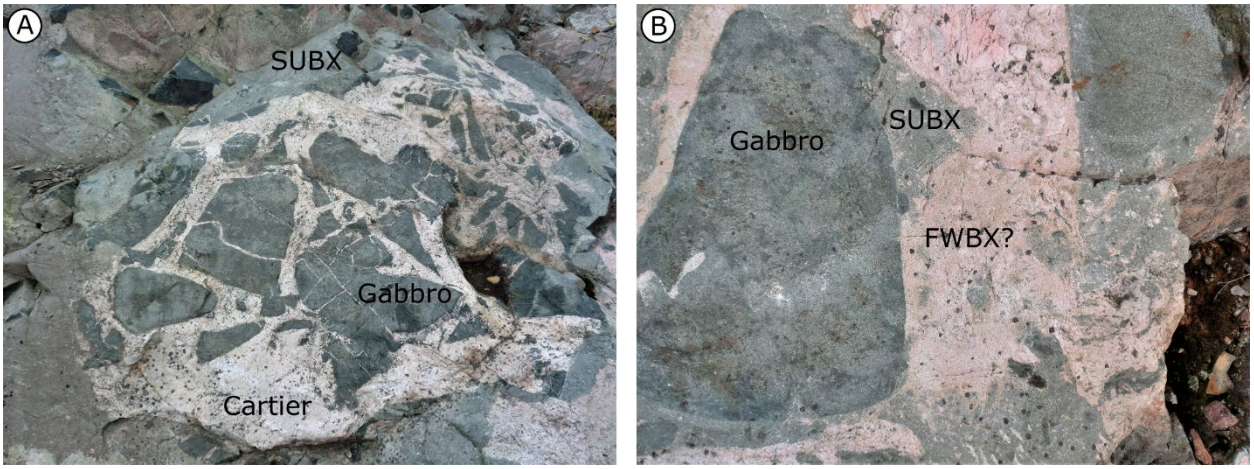


Figure 5.7: A) Clast of a pre-impact Levack Breccia consisting of sub-angular to -rounded dm-sized gabbro in a granitic matrix, surrounded by SUBX. **B)** SUDBX fragment (?) wrapped around subrounded, 30cm large gabbro inclusion itself intruded by feldspar-rich granitoid (FWBX?).

2-A3) Sublayer norite (NAD83 17N, 509540, 5180011)

SLNR is only exposed at the SW end of the 300m x 300m main outcrop. It typically displays a rusty-brown weathering surface sulfide oxidation (Fig. 5.3). The noritic matrix is typically homogeneous, fine- to medium-grained, and displays equigranular textures. It contains abundant country rock inclusions, minor dark green/brown ultramafic inclusions, and QD-type inclusions. Globular and ragged sulfide blebs up to 2cm in size are common.

2-A4) IQD-MTBX (NAD83 17N, 509563, 5180046)

To the NE, the SLNR transitions into FWBX/MTBX in the central part of the main outcrop, indicating the beginning transition from an embayment environment to an offset environment. At the margins of the FWBX/MTBX, bodies or pods of leucocratic QD and IQD occur (Fig. 5.5), often in sharp contact with FWBX/MTBX as shown here. The IQD is characterized by a homogeneous fine-to medium-grained matrix with abundant amphiboles needles and less than 20% (sub)-rounded inclusions. FWBX/MTBX is appearing more heterogenous and rich in inclusions.

2-A5) SLNR-IQD-MTBX-QD (NAD83 17N, 509600, 5179940)

This area on the southeastern flank of the embayment, which was mapped in detail by Carter et al. (2009), features excellent exposure of contact relationships between leucocratic QD, MTBX/FWBX, IQD, and SLNR. It also highlights the complexity and variability in breccia matrix composition in a small area (Fig. 5.8 and Fig. 5.9).

Footwall gabbros and granitoids are crosscut by decimeter wide SUBX veins (e.g., location 4 in Fig. 5.9). Abundant subrounded inclusions from local footwall rocks are set in an aphanitic to very fine-grained groundmass which is, independent of the host unit, composed of predominantly chlorite and epidote.

Leucocratic QD occurs as a large body in contact with the footwall (location 1 in Fig. 5.9). It is medium-grained with abundant amphibole and plagioclase laths as well as oikocrystic quartz, feldspar, and granophyric quartz-feldspar intergrowth (Fig. 5.8B–C).

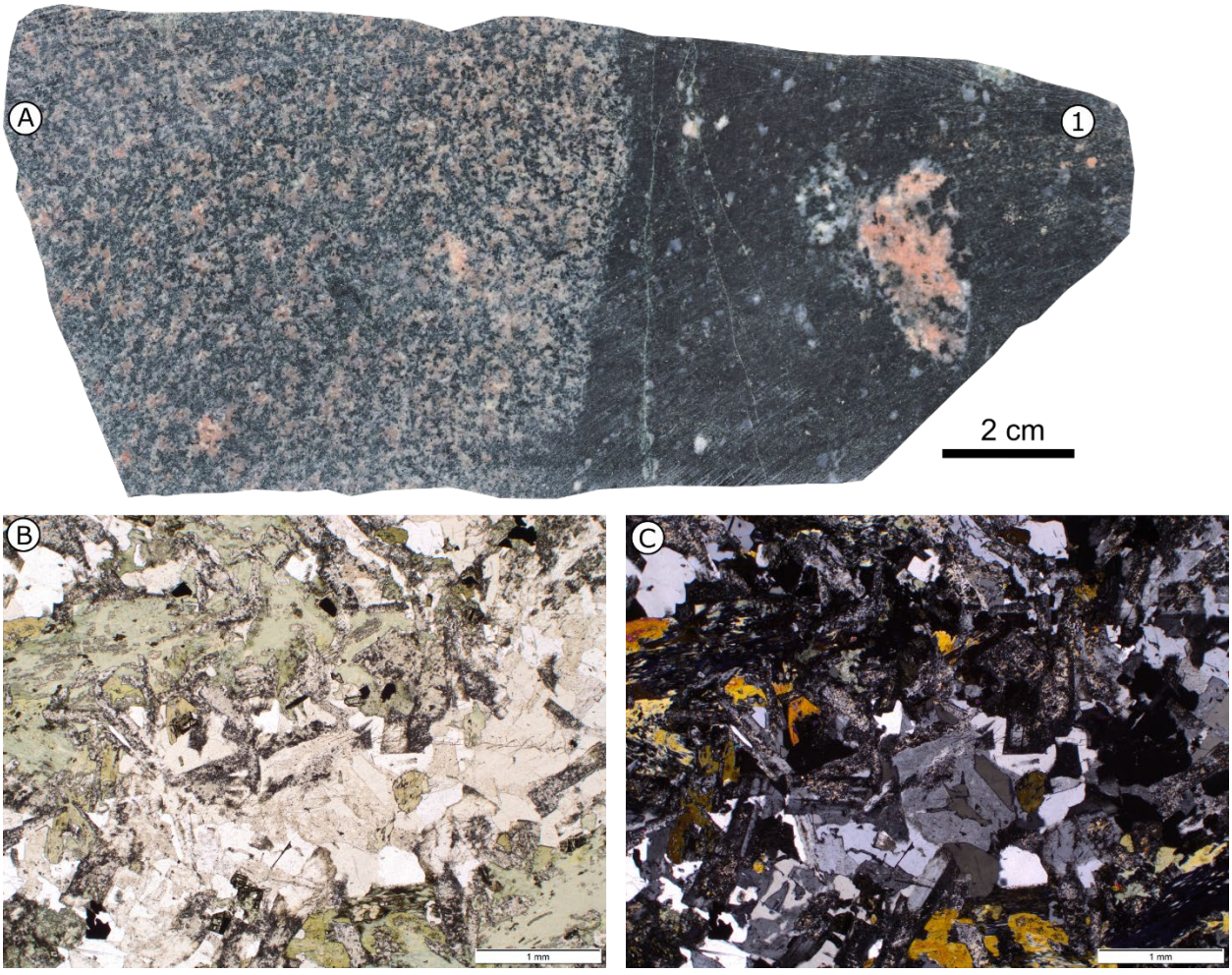


Figure 5.8: A) Sharp contact between LQD (left) and polymictic MTBX/FWBX (right). See Fig. 5.9 for location. Plane- **(B)** and cross-polarized **(C)** images of the leucocratic QD matrix.

MTBX/FWBX (depending on the map) is fine-grained, equigranular, polymictic, and in sharp contact with leucocratic QD at location 1. The polymictic breccia is typically inclusion-rich (>50%) with inclusion sizes ranging from millimeters to several meters. Feldspar with lesser quartz and interstitial amphibole are the dominant minerals in the dark grey, fine-grained breccia groundmass. In contrast, MTBX/FWBX at location 3 (Fig. 5.9), 5m away from location 1, displays typical grey-pinkish colors, is inclusion-rich and appears recrystallized on fresh surfaces. The poikilitic groundmass consists of highly altered equant plagioclase and tabular amphibole chadacrysts in coarser, oikicrystic quartz-(feldspar).

IQD at location 2 (Fig. 5.9) is characterized by a homogeneous matrix with a typical "salt-and-pepper" color on fresh surfaces. It is polymictic with <25% inclusions and minor sulfide blebs. The matrix is fine- to medium-grained and poikilitic. Variably altered, equant plagioclase and rare amphibole needles are set in oikocrystic quartz with minor feldspar and granophyric intergrowth.

Macroscopically, the different breccia units appear to be quite different and easily distinguishable. SUBX adjacent to the funnel/offset dike displays distinct grain size differences, is strongly altered and of a more mafic matrix composition. Groundmass compositions of the other breccia units are remarkably similar (see also Lafrance et al., 2014).

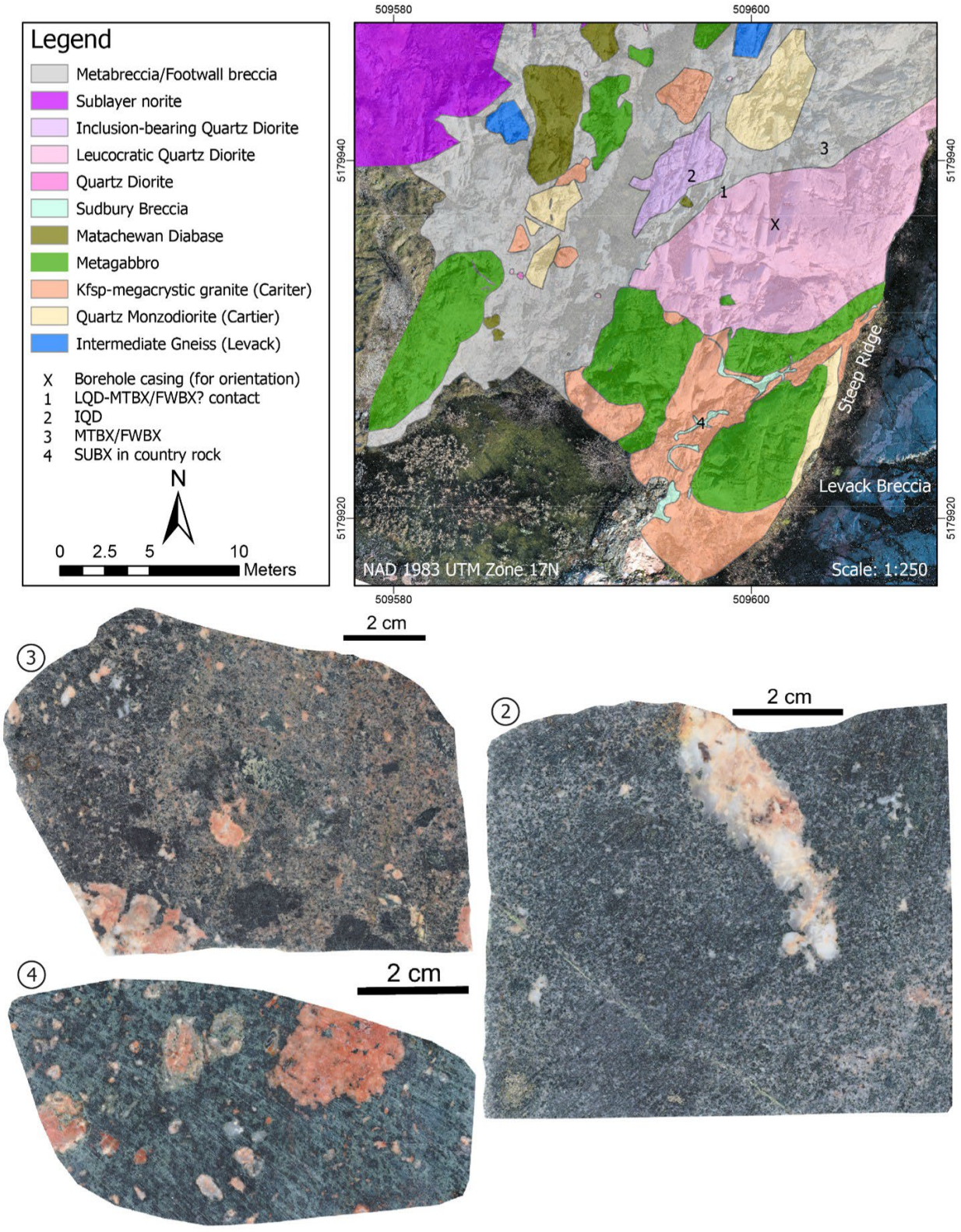
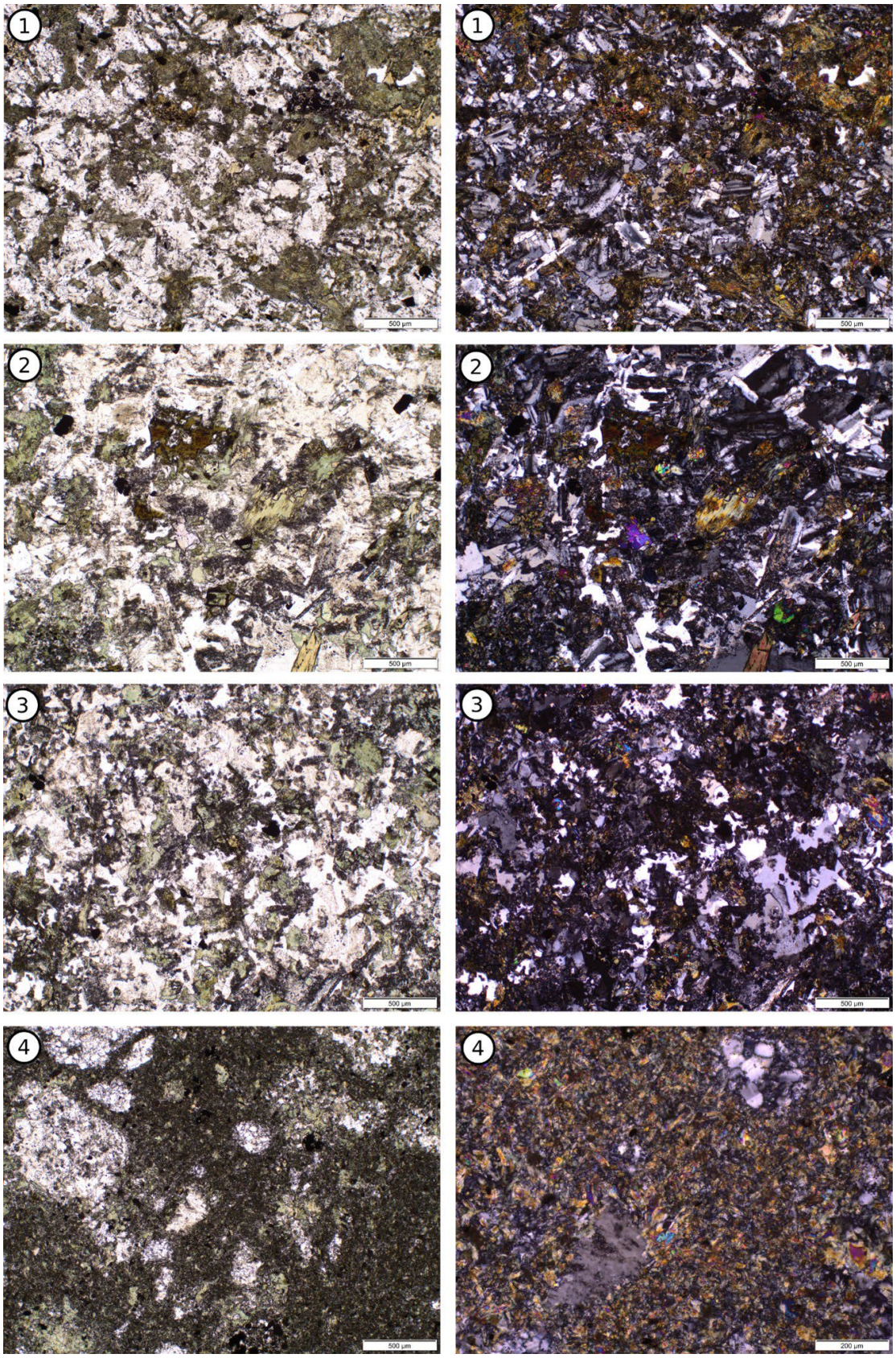


Figure 5.9: Detailed map of the southeastern flank with SUBX, SLNR, IQD, MTBX/FWBX, leucocratic QD, and footwall rocks. Locations of breccia samples are shown in the map. The leucocratic QD-MTBX/FWBX contact is shown in Fig. 5.8. Images of slabs and thin sections are described in the text.


1
1
2
2
3
3
4
4
500 µm
500 µm
500 µm
500 µm
500 µm
500 µm
500 µm
200 µm

2-A6-1 GRBX-Cartier Batholith contact (NAD83 17N, 509538, 5179867)

Located at the SE flank of the embayment just meters away from the covered and re-greened open pit. FWBX is in a well-defined contact over 1–2cm with Cartier granite.

2-A6-2 IQD with QD Inclusions (NAD83 17N, 509552, 5179875)

Sharp contact between QD and IQD (SLNR?). QD is leucocratic, medium-grained and displays well developed amphibole needles <1cm. IQD is characterized by a finer-grained, homogeneous matrix with salt-and-pepper texture, 10–40% inclusions and disseminated blebby sulfides. Several subrounded QD pods or inclusions are visible and highlighted by their more leucocratic appearance (Fig. 5.10A).

2-A7 Leucocratic QD with local footwall clasts (NAD83 17N, 509693, 5180036)

Coarse-grained, leucocratic QD with up to 2cm long amphibole needles and angular, greenish-altered gabbro inclusions (Fig. 5.10B). Located just 1m away from the SE dike margin right at the contact between Cartier batholith and gabbro, this area indicates only a short lateral transport of the gabbro inclusions. Country rock inclusions at QD margins is also documented in other offset dikes, such as Foy, Hess, and Worthington.

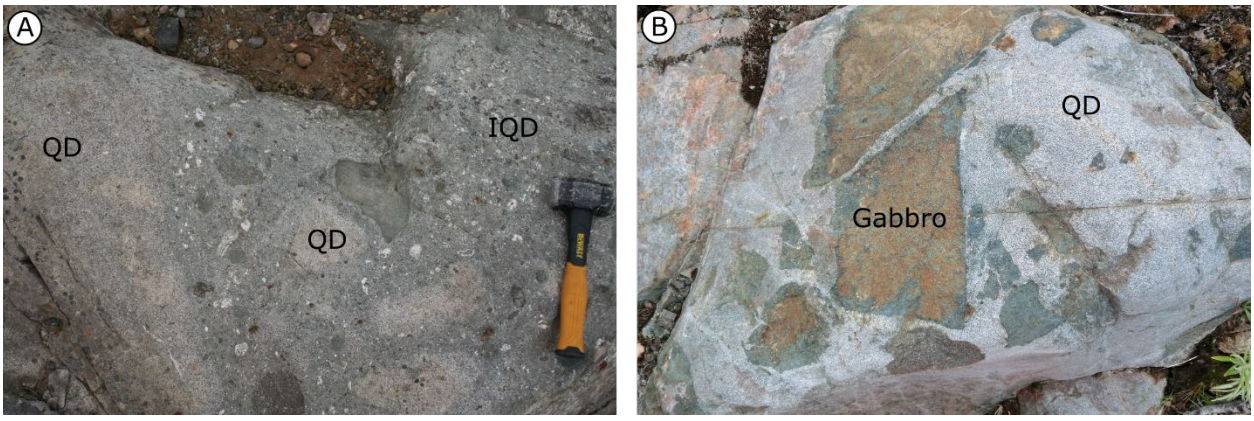


Figure 5:10: A) IQD with three (leucocratic) QD pods/inclusions in sharp contact with (leucocratic) QD at **2-A6-2**. **B)** QD with large amphibole needles and local clasts at **2-A7**. Image width is approximately 75 cm.

2-A8 IQD-MTBX (NAD83 17N, 509799, 5180205)

This area has been mapped in detail by Carter et al. (2009) and shows some excellent IQD-MTBX/FWBX features on glacially polished surfaces. The IQD-MTBX/FWBX contact can be traced over several meters and is generally sharp. MTBX/FWBX is dark grey on fresh surfaces, inclusion-rich (ca. 50%), and closely associated with large (>5m) Cartier granite clasts. Small (<3cm), rounded, red feldspar inclusions are common with mafic inclusions typically being smaller and less common. IQD is leucocratic, medium-grained with well-developed amphibole needles and typically less than 10% inclusions.

B. Intermediate Whistle Dike

2-B1 Podolsky IQD-MTBX-Sulfides (NAD83 17N, 509937, 5180490)

The Podolsky outcrop located 500m NE of the Whistle embayment (Fig. 5.4), features the surface exposure of sharp-walled Cu-PGE-rich sulfide veins from the Podolsky North Zone. Steeply dipping chalcopyrite-rich veins are locally up to 3m wide but range more commonly on a cm- to dm-scale. They follow lithological boundaries as well as crosscuts IQD and MTBX (Fig. 5.11B–C). Flow laminations can be observed in the veins (Lafrance et al., 2014).

IQD predominates in the southern part of the outcrop and MTBX in the northern part. QD pods or inclusions (typically <3m) are more abundant in contact with IQD than with MTBX and display several centimeter long radiating amphibole needles (after pyroxene). Contacts between QD and the other units can be sharp or gradational over a few centimeters (Fig. 5.11A). IQD and MTBX have similar clast compositions, but the latter is generally more enriched. Detailed descriptions of the lithologies can be found in Lafrance et al., (2014).

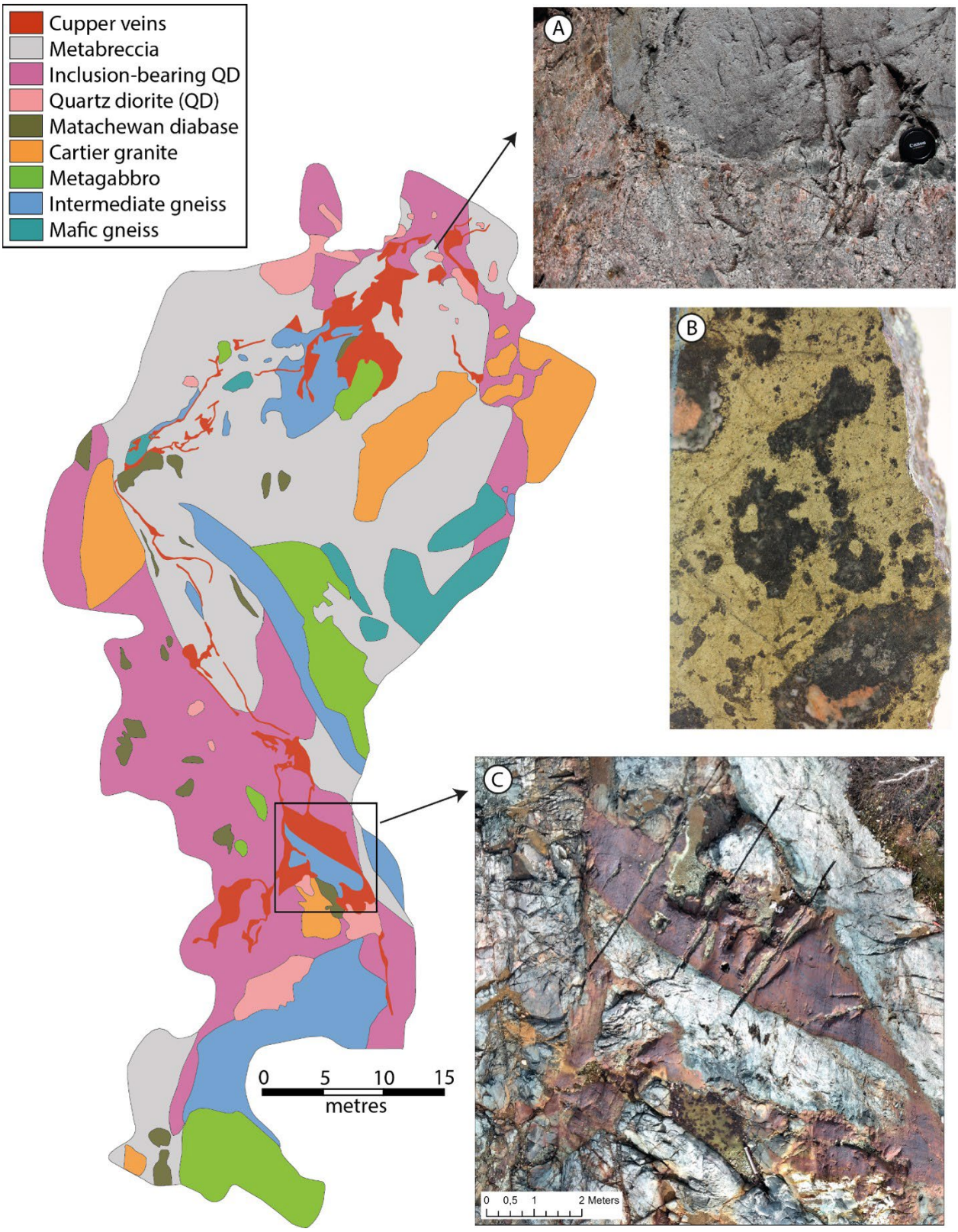


Figure 5.11: Geological map of the Podolsky outcrop (simplified after Carter et al., 2009). **A)** Sharp contact between MTBX and QD. **B)** Chalcopyrite-rich sample with inclusions. **C)** Drone image of the Cu-vein with gneissic clast.

C. Distal Whistle Dike

2-C1 (NAD83 17N, 510293, 5181352)

This stripped outcrop 1.5km northeast of the Whistle embayment (Fig. 5.1) belongs to claims of North American Nickel Inc. and displays the northern most segment of the Whistle dike before it is offset by the Post Creek fault. The eastern part of the outcrop consists of Cartier granite. MTBX is the dominant lithology in the central part. It is very fine-grained, grey pinkish with abundant small feldspar inclusions and <2% disseminated sulfides (Fig. 5.12A–B). Meter-sized inclusions of subangular to subrounded diabase, granite and QD inclusions are common (Fig. 5.12C).

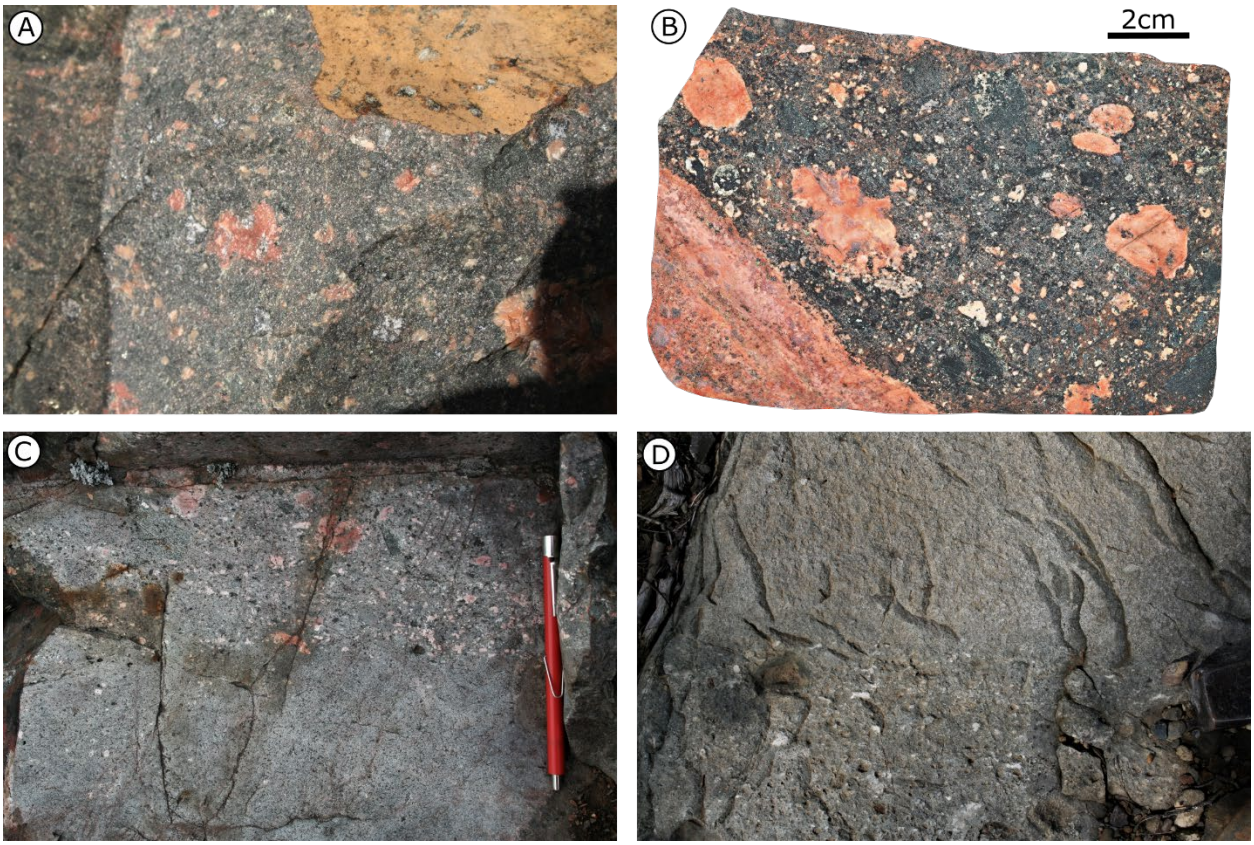


Figure 5.12: A) close-up image of MTBX matrix. **B)** polished slab of MTBX with abundant feldspar clasts. **C)** QD clast in contact with MTBX at the distal Whistle outcrop **2-C1**. **D)** Well defined contact between inclusion-poor and inclusion-rich quartz diorite at the proximal Parkin outcrop **2-D1**. Hammer head for scale

D. Proximal Parkin Dike

Wallbridge Mining´s proximal Parkin properties are located just north of the Post Creek Fault (Fig. 5.1). Mechanical stripping, mapping, and drilling of 63 drill holes as well as geophysical surveys were conducted in 2015 and 2016. Outcrops maps are shown in Fig. 5.13 and Fig. 5.14.

2-D1 Northern Part (NAD83 17N, 509199, 5183212)

The northern stripped outcrops show large sections of the central portion of the Parkin offset dike. IQD is the dominant lithology and incorporates millimeter to tens of meter sized country rock inclusions as well as disseminated and stringer sulfides. QD can locally be observed at the margins of the dike and the outcrops. There are several areas where a sharp contact between QD and IQD can be observed (Fig. 5.12D).

MTBX can be observed as inclusions and irregular pods in QD and IQD, often with well-defined contacts over <1mm and more rarely, with gradational contacts as indicated by changes in inclusion abundance (see Anders et al., 2020).

2-D2 Southern Part (NAD83 17N, 509104, 5183025)

The southern stripped outcrop displays similar features as the northern outcrops. In addition, two prominent large (>10m) felsic gneiss inclusions occur in the central portion of the dike. These potential Levack Gneiss inclusions are not from the immediate country rocks, thus indicating the high energy of the quartz dioritic melt during injection. Similar observations were made by Murphy and Spray (2002) 1km to the northeast of this outcrop.

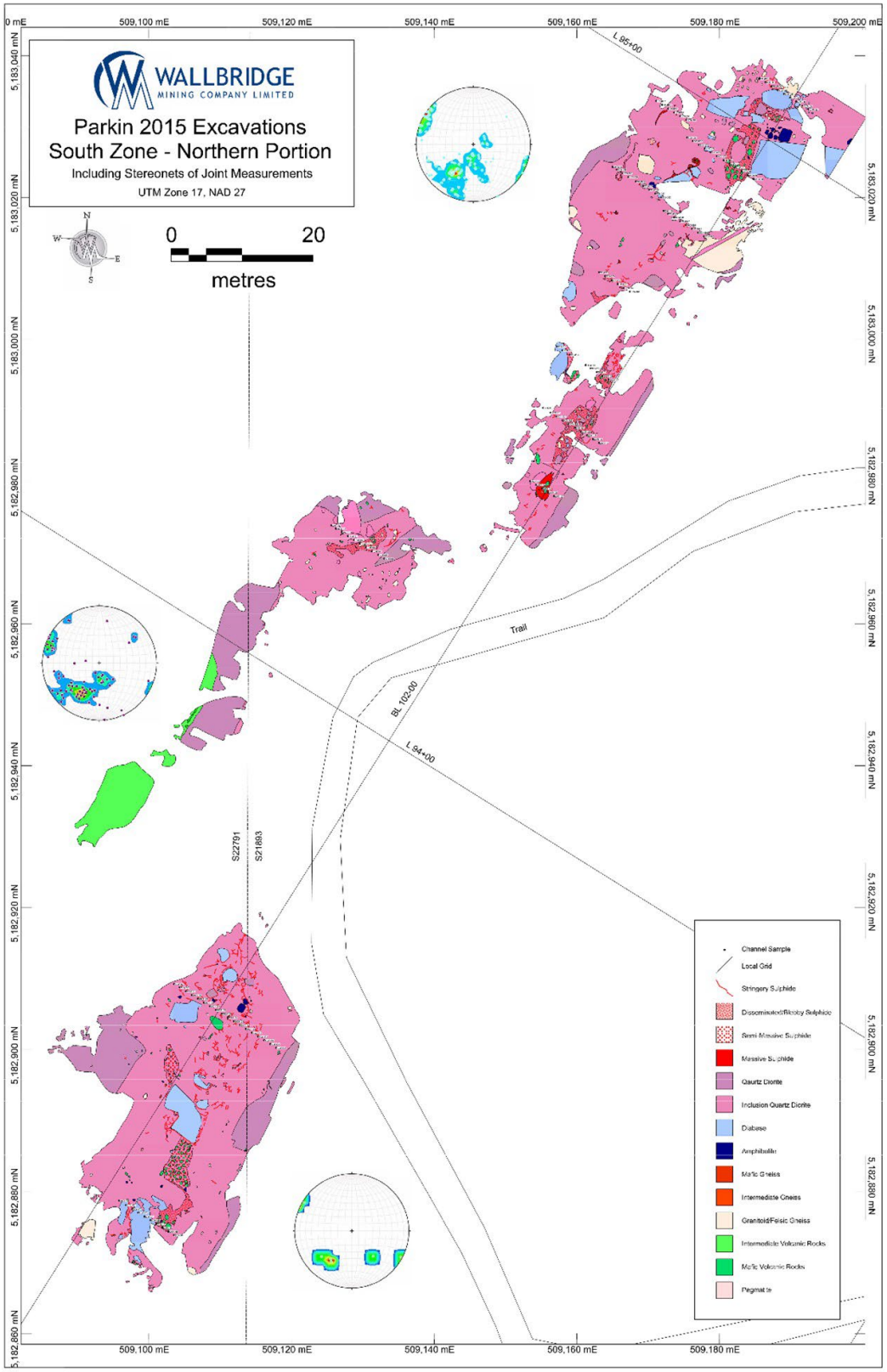


Figure 5:13: Map of Wallbridge Mining´s proximal Parkin stripped outcrops (northern part, from Wallbridge Mining Assessment Report on the Parkin Property 2015-16).

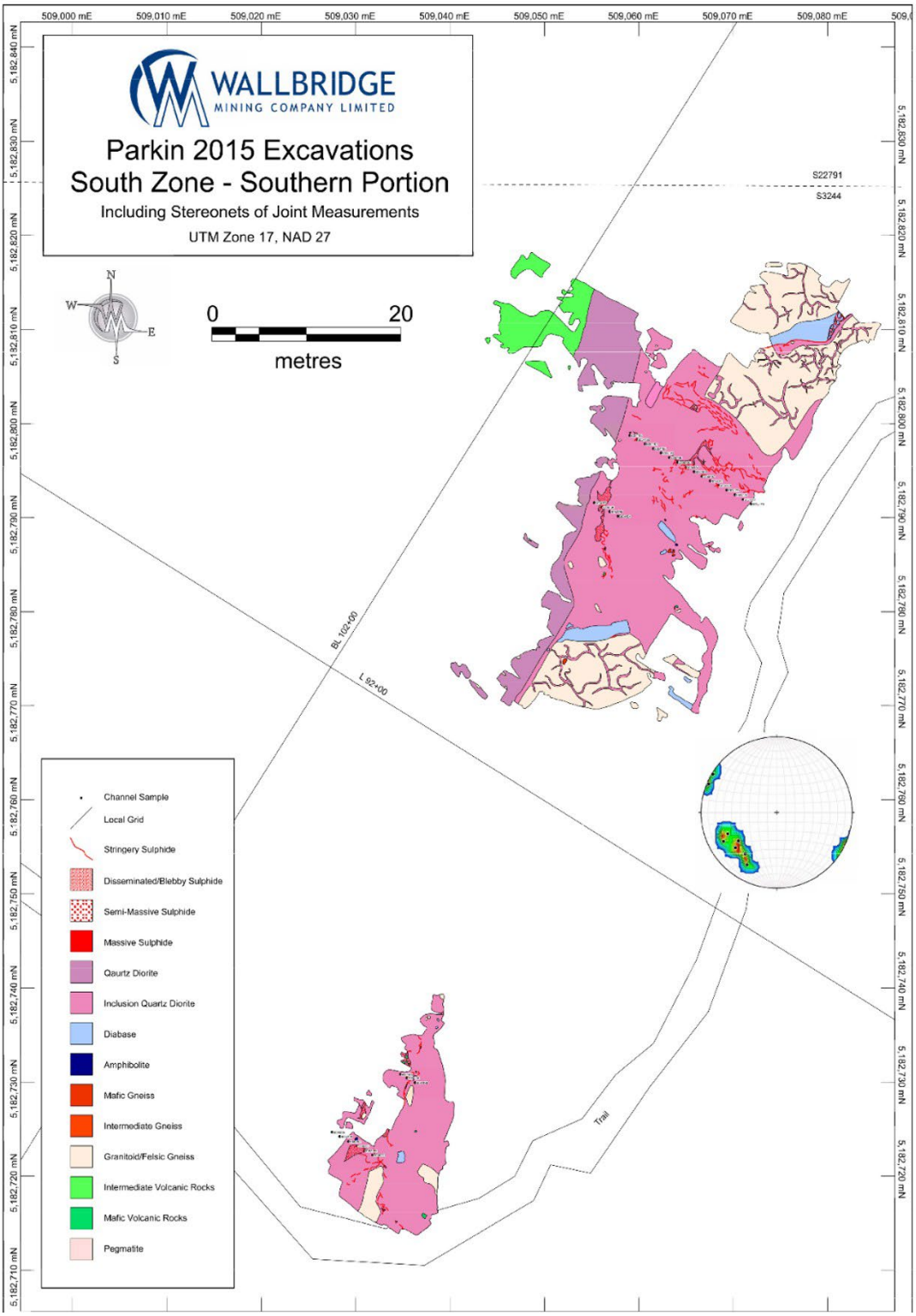


Figure 5:14: Map of Wallbridge Mining´s proximal Parkin stripped outcrops (northern part, from Wallbridge Mining Assessment Report on the Parkin Property 2015-16).

Alternative Field Trip Stops: Worthington, Trill East, and Hess Ermatinger

Worthington

The Worthington offset dike is located at the SW lobe of the SIC and is connected to the Main Mass of the SIC through the Victoria embayment (Fig. 2.1). The dike crosscuts and incorporates inclusions of metasediments of the Huronian Supergroup as well as Nipissing diabase.

Several historical mines targeted small ore bodies close to the surface over the last century. Vale´s Totten mine, in operation since 2014, is the latest addition to Sudbury operations and the only active mine at the Worthington dike. Nevertheless, there are several exploration projects in the area. KGHM´s Victoria project is located close to the embayment and SPC Nickel is conducting an exploration and drilling program on their AER-KIDD property (Fig. 6.1), located between Victoria to the northeast and Totten to the southwest.

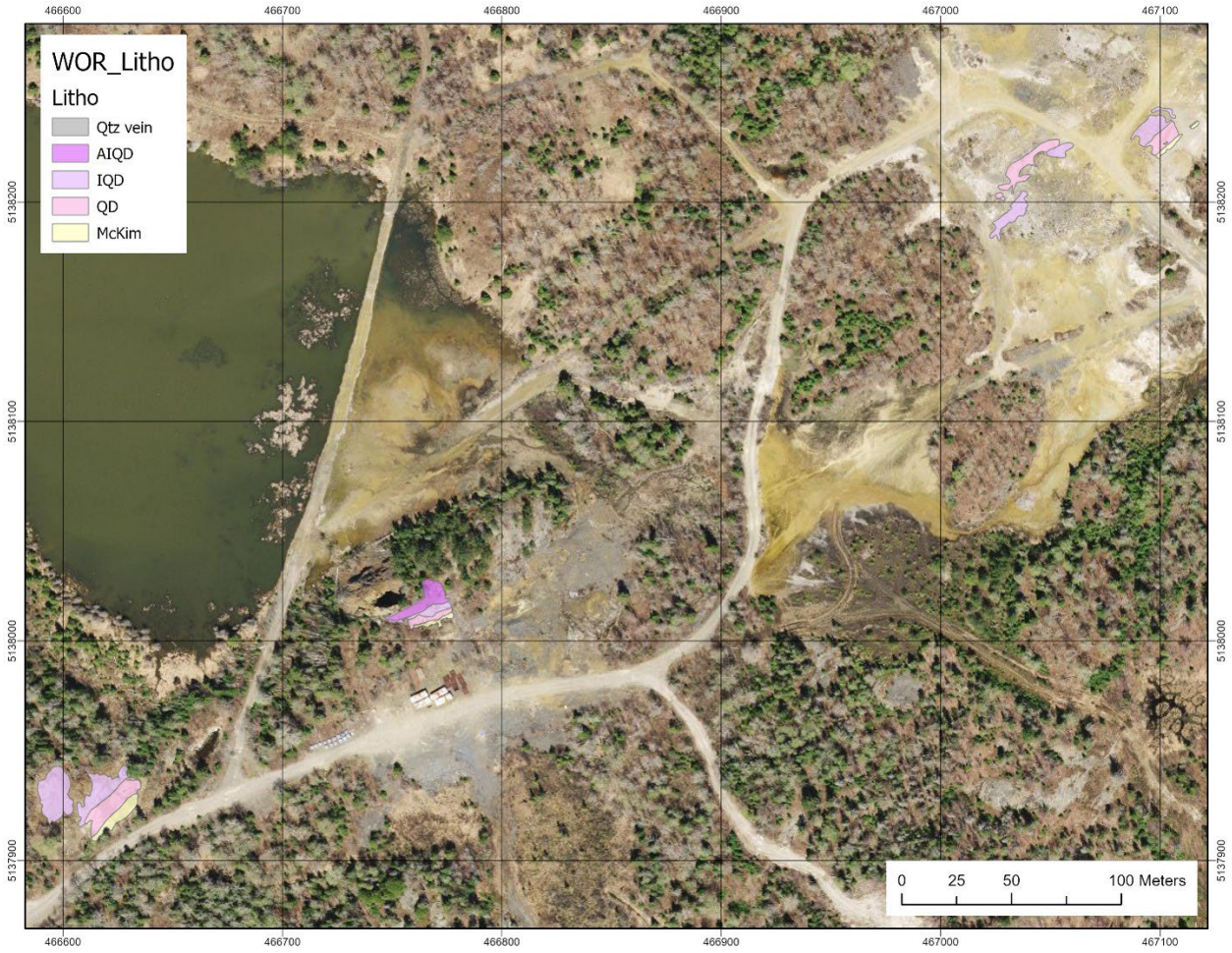


Figure 6.1: Satellite image with outcrop locations of SPC Nickel's AER-KIDD exploration project. Outcrop 1 is located in the southwest, outcrop 2 in the center and outcrops 3 and 4 in the northeast of the map.

AER-KIDD Outcrop 1

Outcrop 1 displays the eastern contact between the Worthington offset dike and Footwall metasediments of the McKim Formation, which are in sharp contact with very fine-grained to spherulitic QD (Fig. 6.2A–B). QD matrix grain size gradually increases over 2–3m towards the center where it is in a well-defined contact with finer grained IQD (Fig. 6.2A–B). The contact is characterized by a) the grain size difference, b) inclusions, and c) disseminated sulfides. Centimeter- to decimeter-sized inclusions are generally aligned parallel to the strike of the dike and are comprised of footwall rock assemblages, i.e., gabbros and amphibolites presumably of the Nipissing mafic suite and McKim Formation metasediments. Subrounded, meter-sized inclusions of QD in IQD are common.

AER-KIDD Outcrop 2

Outcrop 2 is adjacent to the former small Robinson mine and displays similar field relationships as outcrop 1, although less well defined due to abundant sulfide oxidation. In addition to QD and IQD, the most central part of the dike (and most westerly part of the outcrop before the fenced-off mine) is comprised of an amphibolite inclusion-rich quartz diorite (AIQD) which can grade into amphibolite-bearing sulfide matrix breccia. Amphibolite inclusions are typically larger in AIQD than in IQD.

AER-KIDD Outcrop 3 and 4

Outcrops 3 and 4 are located 300m northeast of outcrop 2 and display similar features as outcrop 1 (Fig. 6.3). QD-IQD contacts are generally well defined over a few millimeters and characterized by grain size differences as well as inclusion and sulfide abundances (Fig. 6.3A). In some areas, rounded inclusions of McKim Formation are incorporated into QD close to the footwall contact. Also, “mushroom-like” bulging of McKim Formation into QD can be observed (Fig. 6.3B).

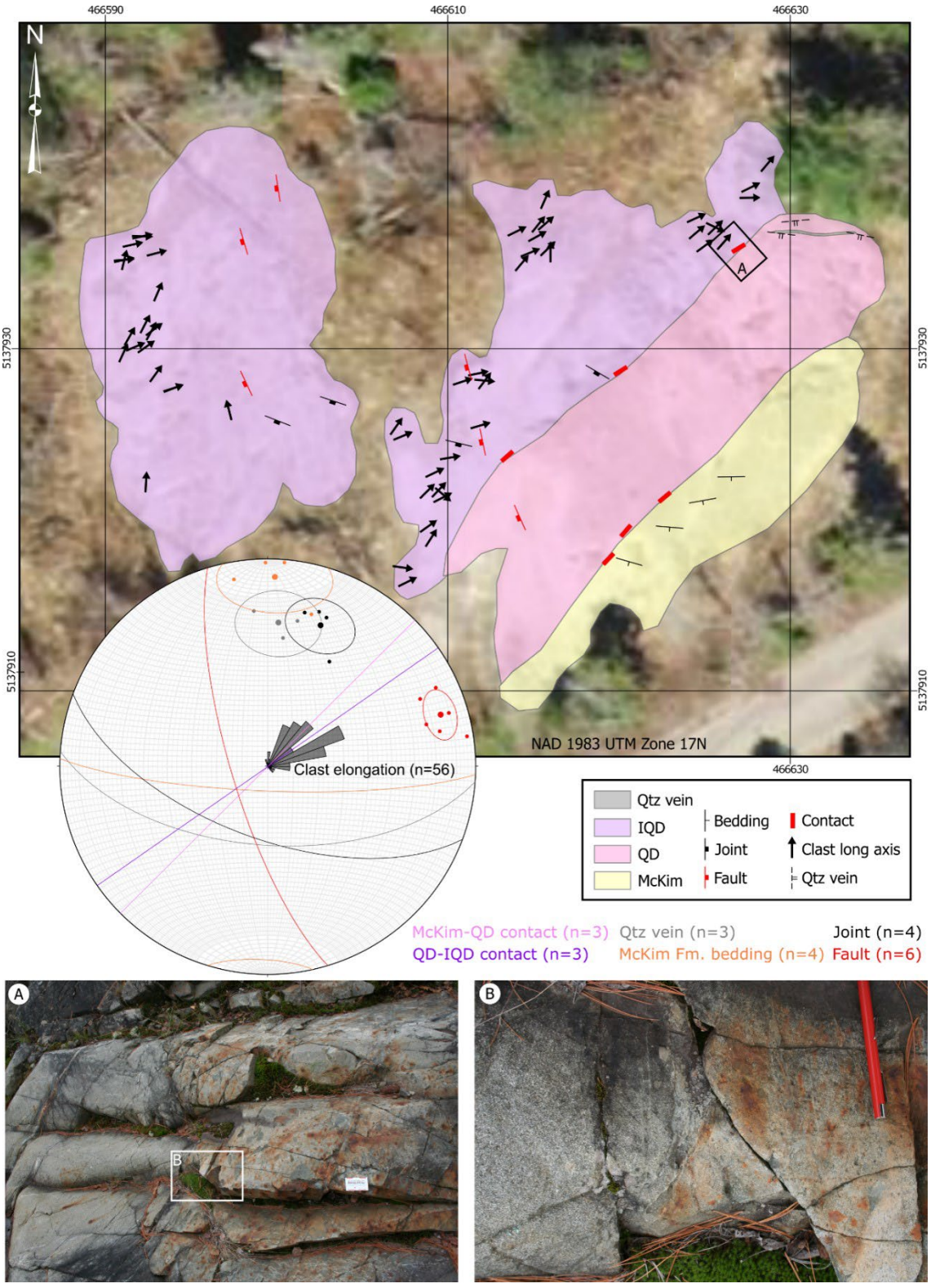


Figure 6.2: Geological map, measurements, and field images of the AER-KIDD outcrop 1. Most country rock clast in the IQD are rotated parallel to the IQD-QD and QD-Footwall contact. A-B) QD-IQD contact is well-defined and characterized by grain size differences as well as inclusion and sulfide content.

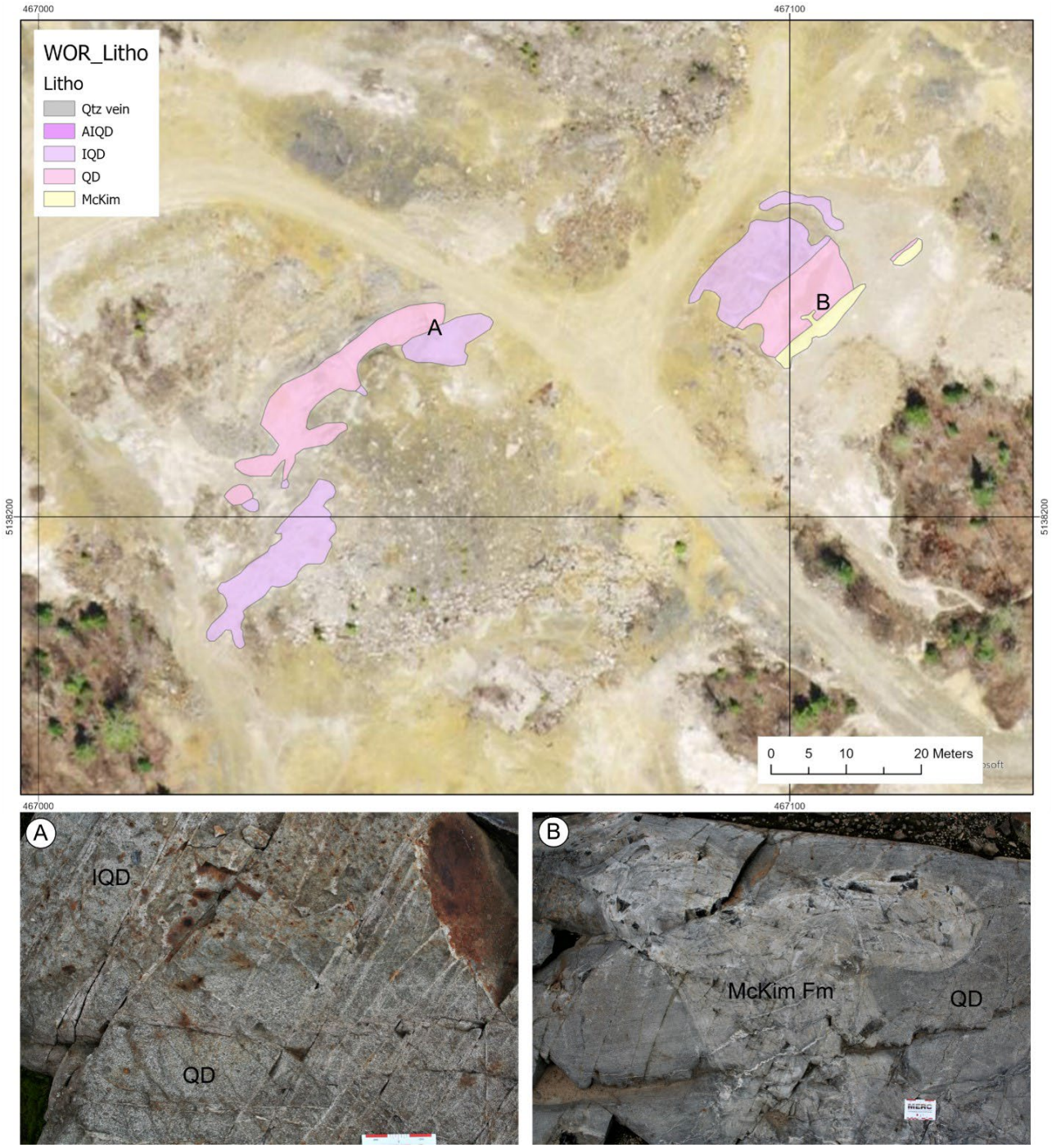


Figure 6.3: Geological map and field images of AER-KIDD outcrops 3 and 4. **A)** Sharp QD-IQD contact defined by grain size differences, inclusions and disseminated blebby sulfides. **B)** McKim Fm. "intruding" QD.

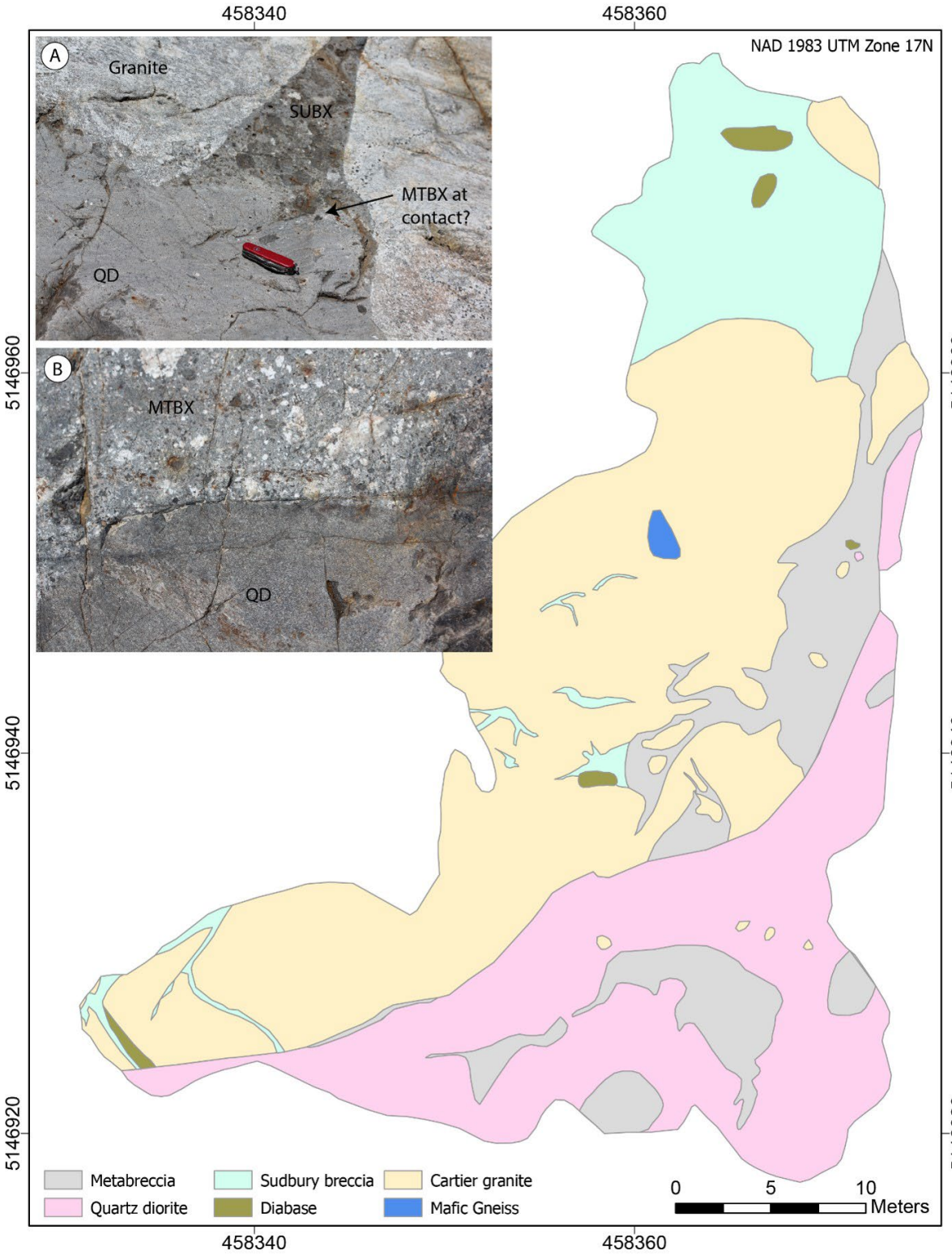


Figure 6.4: Geological map of Trill East outcrop (simplified after Wallbridge Mining Assessment Report on the Trill Property 2014). **A)** SUBX vein crosscut by QD. MTBX might be present at the contact. **B)** Close-up of MTBX inclusion/pod in QD.

Trill East

The 3–20m wide, radial Trill offset dike is located southwest of the Sudbury Igneous Complex, just north of the boundary between Superior province and Southern province (Fig. 2.1). Exploration by Wallbridge Mining in 2005 led to the discovery of a 65m x 5m long sulfide mineralization 4km west of the SIC. Mechanical stripping of several outcrops revealed several offset dike units: QD, IQD, spherulitic QD and glassy QD. Common QD is typically found at the dike margin and IQD in the center. Glassy and spherulitic QD are restricted to thin apophyses in the footwall (Klimesch, 2009).

The Trill East outcrop, ca. 500m west of the SIC, features the northern dike-footwall contact (Fig. 6.4). SUBX crosscuts the Cartier granite in fine veinlets as well as in a thicker zone at the eastern end of the outcrop and is itself crosscut by the offset dike (Fig. 6.4A). Dark grey QD is the only offset dike unit present. Several inclusions or pods of MTBX occur within QD (Fig. 6.4B) or between the QD-footwall contact. MTBX is fine-grained, has a light grey appearance and is enriched in small (mm–cm) plagioclase-quartz, alkali feldspar mineral, and mafic clasts (Anders, 2016).

Hess Ermatinger

The concentric Hess offset dike follows the northern outline of the SIC in ca. 15km distance. It has been traced from E of the Foy-Hess intersection to the Ermatinger township WNW of Sudbury (Fig. 2.1). Exploration efforts by Wallbridge Mining in 2010–2012 led to the mechanical stripping and drilling of several outcrops with one in particular displaying excellent exposures of offset dike-footwall interactions, QD and IQD.

The footwall contact of the steeply SE dipping offset dike can be traced for 40m and is characterized by a decimeter-wide zone of QD chilled against Cartier granite (Fig. 6.5). The contact is often sharp but can be irregular in zones of QD bulging into the footwall (Fig. 6.5C). Granite inclusions are abundant in the chilled QD, vary in size from cm-dm and are sometimes partially digested (Fig. 6.5D). Several smaller and larger apophysis are cutting through the footwall granite and diabase dikes.

QD grain sizes gradationally increase from chilled to medium-grained over 5m. The contact between medium-grained QD and fine-grained IQD is predominantly defined by the abrupt change in grain size (Fig. 6.5A–B). IQD is characterized by a fine-grained, sulfide-poor (<2%), inclusion-poor (<10%) IQD. In comparison to the local inclusions in the chilled QD, inclusions in IQD are of gabbroic, dioritic and ultramafic composition not directly associated with the local footwall. Disseminated blebby sulfides occur in patches and are composed of a typical magmatic sulfide assemblage.

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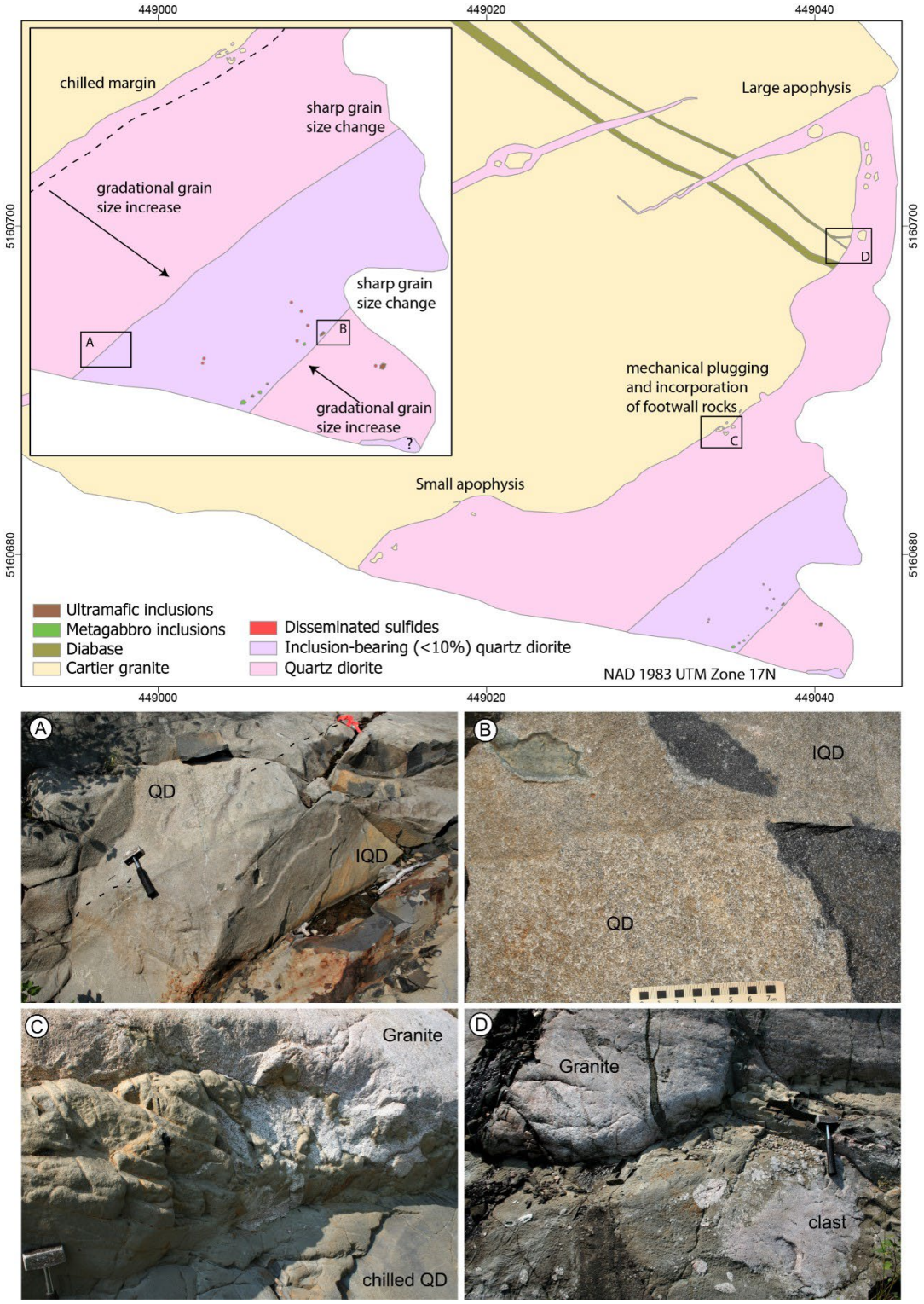


Figure 6.5: Geological map of the Hess Ermatinger outcrop. **A)** Northern contact (dashed lines) between medium-grained QD and fine-grained IQD. Rusty spots in IQD indicate presence of sulfide blebs. **B)** Southern QD-IQD contact with fresh and weathered surfaces. Gabbro clast in IQD is not from the local footwall. **C)** Chilled QD intrudes into Cartier granite. **D)** Dm-sized granite clast showing partial digestion. Diabase dikes are crosscut by QD.

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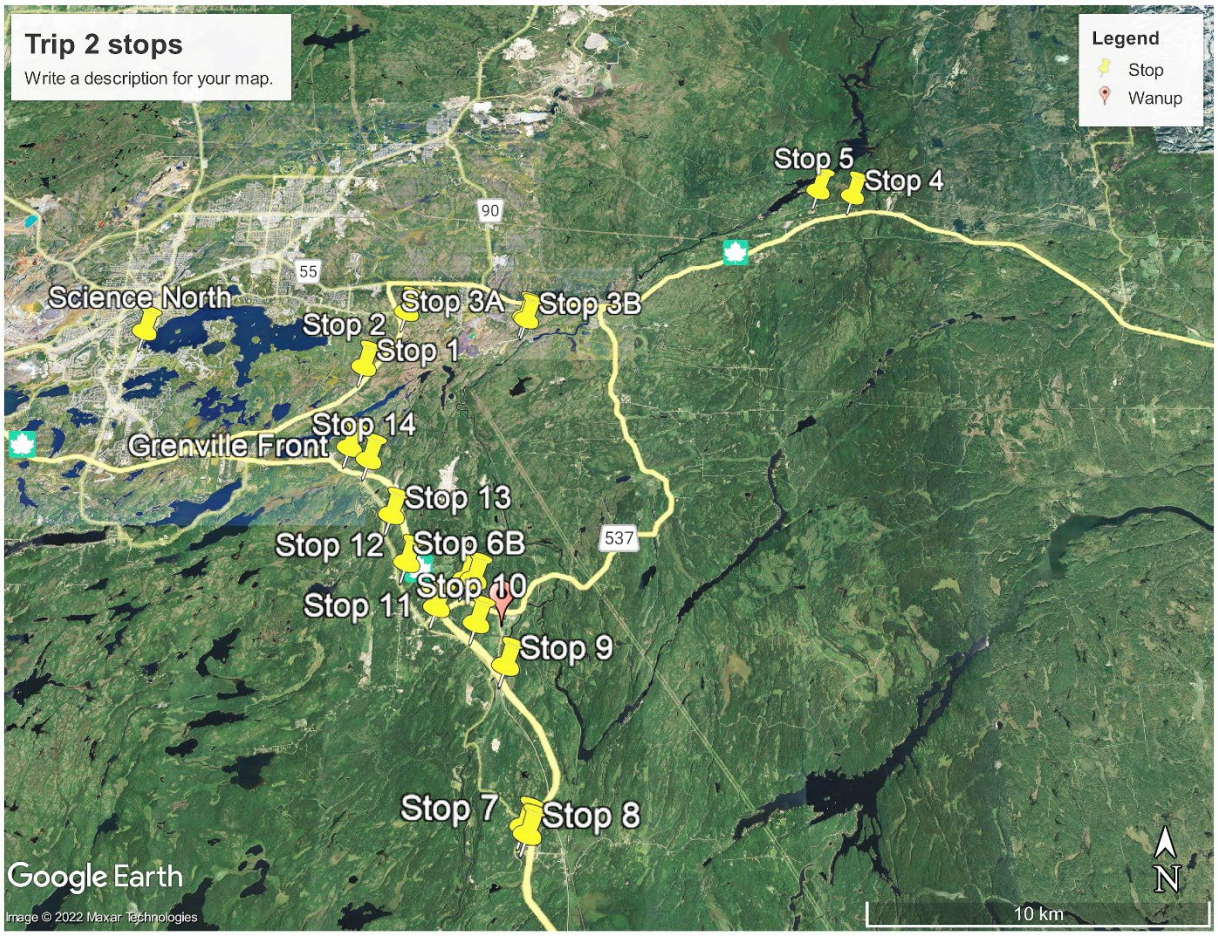
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Location of stops for ILSG Field Trip 2. Trip starts and leaves from Science North (upper left). Stops 1 to 5 are accessed along Highway 17 and the Highway 17 bypass. Stops 6 is on Highway 537. Stops 7 to 14 are accessed from Estaire Road (formerly Highway 69).

Field Trip 2 – Geology of the Grenville Front and the Grenville Front Tectonic Zone in the Sudbury area

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Introduction

The field trip uses road accessible outcrops. All of the road stops can be accessed using a 2-wheel drive vehicle. Unless otherwise stated, all UTM co-ordinates are in Zone 17, datum NAD 83.

Safety

Many of the field trip stops are located on highways that are especially busy during the summer season. Care should always be exercised when parking, exiting vehicles, and crossing the roads. Use of safety vests and/or bright clothing is recommended, in order to improve your visibility to motorists.

Most of the trip routes are on Crown land or public roadways, but access is on or near private property in some cases. As in all such situations, please respect the property rights of others, so as to maintain good relationships, so that future access for geologists is not adversely affected.

Terminology

A number of terms used in this report are outlined below.

Rock Classification

Layering thickness terms used in this report are listed below. These terms apply to bedded, layered and gneissic rocks.

Very thinly layered	<3 cm
Thinly layered	3 to 10 cm
Medium layered	10 to 30 cm
Thickly layered	30 to 100 cm
Very thickly layered	1 to 3 m
Extremely thickly layered	>3 m

Terminology for plutonic rocks follows that of Streckeisen (1976) and LeMaitre et al. (2002).

For metamorphic rocks, mineral prefixes are listed in order of relative abundance, starting with least abundant first. Mineral abbreviations follow Whitney and Evans (2010). The following conventions are used regarding descriptive adjectives. A *gneissic granite* is a meta-igneous rock of granitic composition. A *granitic gneiss*, a *granite gneiss*, or a *gneiss of granitic composition* may be either a meta-igneous or a metasedimentary rock. Similarly, a *tonalitic gneiss* or a *tonalite gneiss* is a gneiss of tonalite modal composition but may be of either meta-igneous or metasedimentary origin. A *gneissic meta-arkose* is a metasedimentary gneiss of overall granitic composition. The term metamorphic grade is used where bulk-rock composition or other factors prevent a more detailed assignment of metamorphic conditions. Where metamorphic conditions can be outlined, metamorphic facies terminology is used.

Many rocks in the Grenville Province were subjected to extreme ductile deformation and subsequently recrystallized, and can be described either as tectonites or gneissic mylonites. Several field-based terms have been proposed to describe these gneissic mylonites including the terms straight gneiss, block gneiss, and porphyroclastic gneiss (e.g., Davidson et al. 1982; Hanmer and Ciesielski 1984).

A migmatite is a heterogeneous rock composed of two or more components, one generally quartzofeldspathic in composition (leucosome or neosome) and the other more mafic in composition (paleosome or mesosome). Within the field trip area, such rocks are commonly layered, and in many instances, are formed by partial melting during high-grade regional metamorphism.

Descriptive terminology for these rocks follows Sawyer (2008) and Mehnert (1971). Migmatites collectively display a wide variety of features depending on the degree of partial melting and deformation during development. The first-order division of migmatites, based on morphology and proportion of leucosome, results in 2 types: metatexite and diatexite. The division between the 2 is based on the relative amount of melt (leucosome) in the rock. The Ontario Geological Survey uses a boundary of 20% leucosome between metatextite and diatextite, which is near the minimum value suggested by Sawyer (2008) but does not require the same precision in estimating leucosome content as the use of 16% would require. The 20% boundary also accounts for the fact that initial bulk-rock composition of the protolith is a factor in the amount of partial melt that can be produced, and thus is better suited for a wide range of bulk-rock compositions.

Purpose

The purpose of the trip is two-fold. First, is to examine the nature of the Grenville Front and the associated Grenville Front tectonic zone using a variety of exposures in the Sudbury area. Second, is to examine outcrops mapped in the fall of 2021 between Wanup and Estaire. This new mapping indicates that Nepewassi domain rocks occur closer (by approximately 5-10 km) to the Grenville Front then previously recognized. Also, although not specifically part of the field trip route, the new field data also suggests the presence of a major north-northwest-trending structure along the Wanapitei River near Wanup, with the character of the rocks in the Grenville Front tectonic zone being very different in lithology and structural character on either side of the structure (*see* section on "The GFTZ zone near Wanup").

Geological Setting

Proterozoic Rocks in the Sudbury area

Proterozoic rocks in the Sudbury area are assigned to either the Paleoproterozoic Southern Province or the Mesoproterozoic Grenville Province (cf. Wynne-Edwards 1972; Easton 1992). The Southern Province in Ontario comprises Paleoproterozoic metasedimentary and metavolcanic rocks of the Huronian Supergroup and gabbroic intrusions of the Nipissing gabbro suite. Also included in the Southern Province are the Sudbury Igneous Complex and the Whitewater Group; plutonic and minor volcanic rocks of the Killarney Magmatic Belt; and rocks of the Sudbury diabase dike swarm (Figure 1) (Bennett, Dressler and Robertson 1991).

The Huronian Supergroup (Figure 2) was deposited unconformably on Archean plutonic and supracrustal rocks of the Superior Province. The lowest unit, the Elliot Lake Group, consists of both metavolcanic and metasedimentary rocks (Figure 2). In the Sudbury area, the metavolcanic units include tholeiitic basalts of the Elsie Mountain Formation, evolved tholeiitic basalts, dacites, and metasedimentary rocks of the Stobie Formation, and dacites and rhyolites of the Copper Cliff Formation (*circa* 2460 Ma). The latter are likely coeval with the Murray and Creighton granites (Bennett, Dressler and Robertson 1991; Bleeker et al. 2015). The metavolcanic units interfinger with, and are overlain, by the Matinenda Formation in the west and the McKim Formation in the east. Geochemical data reported by Innes (1972, 1977), Easton (1998), Gordon (2021) indicate a tholeiitic affinity for the Stobie Formation, whereas felsic metavolcanic rocks of the Copper Cliff Formation and the Murray and Creighton granites show calc-alkalic signatures.

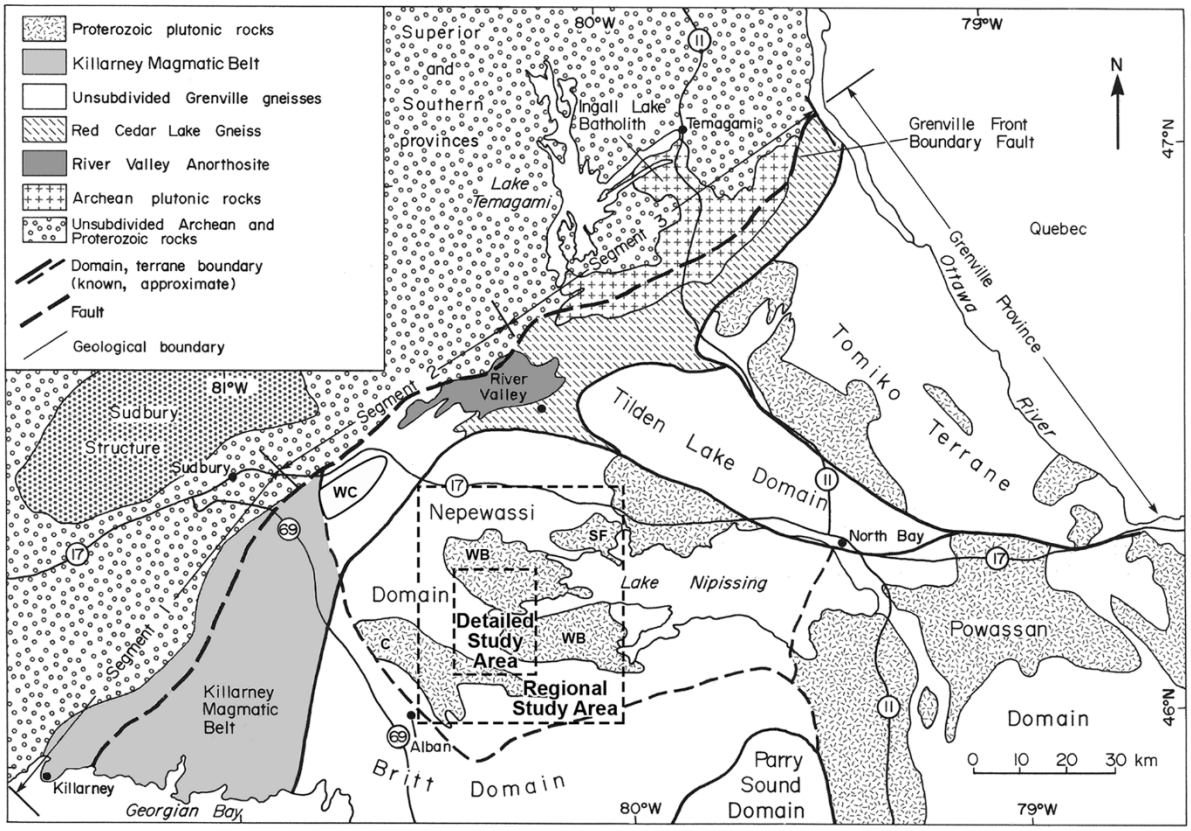


Figure 1. Geology of the northern Central Gneiss Belt of the Grenville Province and the Grenville Front region in Ontario. Locations of mapping areas described by Easton (2014) and Van de Kerckhove (2014) are indicated by dashed-line boxes. Abbreviations: C, Cosby pluton; SF, Sturgeon Falls batholith; WB, West Bay batholith; and WC, Wanapitei complex. Figure *modified from* Easton (1992, p.755).

At the base of the Huronian Supergroup in the Elliot Lake, Agnew Lake and Sudbury areas are several layered gabbro to anorthosite intrusions referred to as the East Bull Lake intrusive suite (Peck et al. 1993; James et al. 2002a, 2002b). These bodies have ages of *circa* 2475 Ma (Clough and Hamilton 2017; Krogh et al. 1984) and appear to be slightly older than the rocks of the Elliot Lake Group.

Each of the 3 groups overlying the Elliot Lake Group consists of sedimentary cycles of conglomerate, mudstone, siltstone or carbonate, capped by crossbedded sandstone (Bennett et al. 1991) (Figure 2). Conglomerate units (e.g., Ramsey Lake, Bruce and Gowganda formations) in each of the cycles have been interpreted as being glaciogenic in origin, likely deposited in a marine environment adjacent to an ice shelf. The siltstone and sandstone units are interpreted to represent deposition during warmer intraglacial or post-glacial periods in either fluvial or marine environments (Junnila and Young 1995; Fralick and Miall 1989). Huronian Supergroup deposition was complete by 2217 Ma, the age of the Nipissing gabbro (Davey et al. 2019; Corfu and Andrews 1986; Noble and Lightfoot 1992).

The Huronian Supergroup has been interpreted to represent a Wilson cycle, starting from a rifting phase represented by the Elliot Lake, Hough Lake and Quirke Lake groups; followed by a passive margin sequence (Cobalt Group); and concluded by a continent-arc collision between the Superior-Southern provinces and the Wisconsin Magmatic Arc Terrane (e.g., Young 1983; Hoffman 1989; Bennett et al. 1991).

This collisional event at 1870 to 1835 Ma, termed the Penokean Orogeny, is believed to be responsible for most of the metamorphism and deformation present in the Huronian Supergroup. The scale and intensity of the Penokean Orogeny remains a subject of debate (Davidson et al. 1992; Card 1992; Raharimahefa et al. 2014; Holm et al. 2018; Zi et al. 2022), in part because the Penokean Orogeny has no associated plutonism in Ontario. In contrast, Riller et al. (1999) attributed deformation and peak metamorphism of the Huronian Supergroup to the Blezardian Orogeny (2470-2220 Ma), with subsequent transpressional deformation during the Penokean. These divergent views reflect the lack of constraints on the age of Huronian Supergroup metamorphism and deformation.

The Sudbury Igneous Complex was emplaced at 1850 Ma (Krogh et al. 1984; Davis 2008) and consists of a lower, ore-bearing sublayer, a main mass of norite, and an upper granophyre (e.g., Dressler et al. 1991). Associated with the Sudbury Igneous Complex are brecciated rocks, termed the Sudbury breccias (e.g., Dressler et al. 1991), consisting of randomly oriented blocks of country rock in a fine-grained, pseudotachylite matrix. The breccias occur up to 200 km from Sudbury but are most abundant near Sudbury. The Sudbury Igneous Complex and related rocks have been variously interpreted as originating from meteorite impact, impact-induced plutonism and volcanism, and volcanism (*see* reviews in Pye et al. 1984). The southern part of the Sudbury Igneous Complex was weakly metamorphosed by an event that also retrograded metamorphosed rocks of the Huronian Supergroup. Regional sodium and potassium metasomatism and silicification have intensely altered rocks locally within the Huronian Supergroup, especially along faults, at *circa* 1700 Ma (Meyer et al. 1990; Gates 1991; Schandl et al. 1994; Easton et al. 1996; Fedo et al. 1997). Significant magmatism occurred again at 1750 to 1730 Ma and at 1500 to 1450 Ma in the Killarney Magmatic Belt (van Breemen and Davidson 1988; Davidson and van Breemen 1994; Krogh 1994).

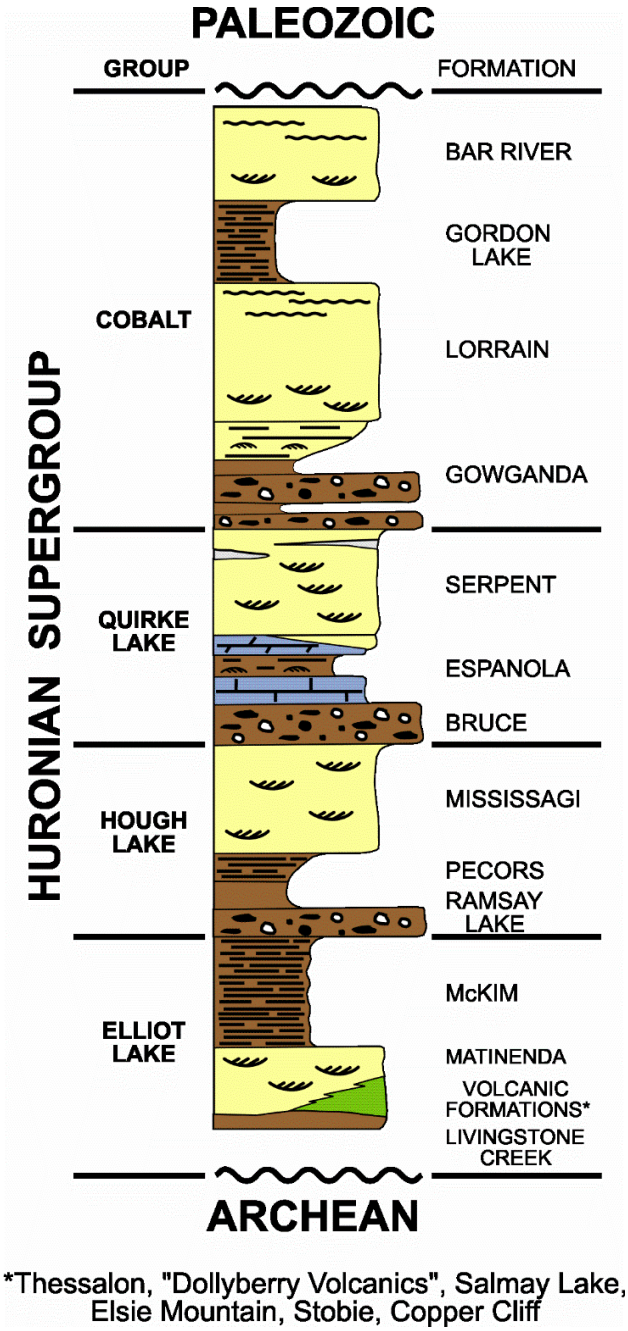


Figure 2. Idealized Huronian Supergroup stratigraphy as utilized by the Ontario Geological Survey based on Robertson, Card and Frarey (1969). Yellow units are sandstone dominated, blue units are carbonate rocks (limestone or dolostone), brown units are mudstones and/or turbidites (lined) or conglomerates. Green units are volcanic rocks.

Table 1. Timing of major geological events and summary of age constraints on the main rock units present in the Sudbury area.

Event and/or Map Unit	Age Constraint (Ma)	Comment and/or Source
Grenville dike swarm	586±4	Kamo, Krogh and Kumarapeli (1995)
Pegmatite vein emplacement	989±2	Corfu and Easton (2000)
Age of peak metamorphism in the hanging-wall of the Grenville Front tectonic zone	1000 to 990	Corfu and Easton (2000), this study
Age of peak Grenvillian metamorphism in the Central Gneiss Belt	1040 to 1030	Carr et al. (2000)
Sudbury dike swarm	1238±4	emplaced in or along northwest-trending faults in the Southern Province, deformed and metamorphosed in the Grenville Province, Krogh et al. (1987).
Killarney magmatic belt second-stage magmatism, coincident with magmatism in the Eastern Granite Rhyolite Province and in the Central Gneiss Belt	1471±3	van Breemen and Davidson (1988)
Regional albitization metasomatic event	1701±4	U/Pb monazite, Schandl, Gorton and Davis (1994); fluid focussed along northwest faults
Killarney magmatic belt volcanism and high-level plutonism	1740, 1747±3, 1749±12	van Breemen and Davidson (1988); Sullivan and Davidson (1993); Davidson and van Breemen (1994)
Northwest-trending regional faults	Pre-1700, post-1850	Faults cut Sudbury Structure
Penokean orogeny (folding and metamorphism of Huronian Supergroup rocks?)	1775±10 ~1835	Peak deformation. Zi et al. (2022) Peak metamorphism. Holm et al. (2001)
Impact event and formation of Sudbury breccia	1850±1	Krogh, Davis and Corfu (1984); Davis (2008)
Penokean arc formation	1880-1870, 1845-1830	Zi et al. (2022)
Thrust faulting	post-F_2 pre-regional faulting	Sudbury breccia localized along these faults, suggesting they are pre-Sudbury Structure
F_2 folding	post-2200, pre-1700, pre 1850?	Pre-regional faulting, Nipissing sills axial planar to folds
F_1 folding	pre-2200	Nipissing sills folded or intruded into early folds
Emplacement of Nipissing gabbro sills	2217±4	Davey et al. (2019); Corfu and Andrews (1986); Noble and Lightfoot (1992)
Huronian Supergroup sedimentation	>2220 but <2460	Youngest detrital grains in Bar River Fm are 2306 Ma (Hill, Davis and Corcoran 2018)
Huronian Supergroup felsic volcanism and related plutonic rocks, including the Matachewan dike swarm	~2477 to 2375 (2450±25, 2460±20, 2477±9, 2415±5	Krogh, Davis and Corfu (1984), Heaman (1997); Corfu and Easton (2000), Krogh, Kamo and Bohor (1996), Smith (2002); Bleeker et al. (2015)
Emplacement of East Bull Lake intrusive suite rocks	2475±2	Heaman (geochronologist, University of Alberta, personal communication, 1999); Clough and Hamilton (2017)
Emplacement of orthopyroxene hornblendite bodies (East Bull Lake suite)	2468±5	Corfu and Easton (2000)
Emplacement of alkali feldspar granite and megacrystic granodiorite near River Valley	2660 to 2665	Bodies intrude Crerar and Pardo gneiss, Easton (2003)
High-grade Archean metamorphism and migmatization	2647±4	Krogh, Davis and Corfu (1984); Wodicka and Card (1995); Ames et al. (2005)
Emplacement ages of Archean units in the Sudbury area	2711±7 to 2642±1 see also Table 5	Krogh, Davis and Corfu (1984); Wodicka and Card (1995); Chen, Krogh and Lumbers (1995); Meldrum et al. (1997); Ames et al. (2005)

The last major magmatic activity in the Southern Province occurred at *circa* 1240 Ma with the emplacement of the northwest-trending Sudbury diabase dike swarm (Krogh et al. 1987). This event is noteworthy, as rocks of this dike swarm can be traced across the Grenville Front into the Grenville Front tectonic zone, providing an important marker horizon (e.g., Bethune 1997). (Figure 3).

Significant changes in thickness within the Huronian Supergroup occur east and west of a line roughly coincident with the trace of the north-trending Upper Wanapitei River fault. Debicki (1990) estimated the thickness the Huronian Supergroup at Sudbury to be approximately 10,350 m, 85% of which consists of the lower 3 groups. In contrast, east and northeast of Wanapitei Lake, the thickness of the Huronian Supergroup is approximately 6,250 m, 75% of which consists of the Cobalt Group. The Elliot Lake, Hough Lake and Quirke Lake groups are all considerably thinner east of Wanapitei Lake, and the McKim and Ramsey Lake formations are apparently absent (Easton and Murphy 2002).

Major Fault Systems

Upper Wanapitei River fault

The Upper Wanapitei River fault has had a protracted deformation history, exhibiting at least 7 to 8 km of left-lateral movement between 2170 and 1850 Ma (Buchan and Ernst 1994), and at least 3 km of left-lateral movement post-1040 Ma (Easton and Murphy 2002). According to Easton (2000), the north-trending Upper Wanapitei River fault apparently divides the Archean rocks in the Elliot Lake to North Bay area into two domains, with the boundary between these domains passing through Street Township. The eastern domain, which includes the River Valley–Hagar area consists of supracrustal and metaplutonic rocks, with deeper levels in the crust being exposed to the south, likely due to Grenville orogenesis. In contrast, the western domain is pluton-dominated, with deeper levels of the crust, being exposed to the east. The amount of vertical movement across the fault is unknown. Significant vertical movement would be necessary to juxtapose granulites of the Levack gneiss complex against higher-crustal-level rocks to the east, however, if the Upper Wanapitei River fault re-activated an older listric fault system, then considerably less relative uplift across the fault may be present.

Murray fault system

Significant changes in stratigraphic thickness within the Huronian Supergroup occur across the Murray fault system (e.g., Bennett et al. 1991, and references therein). North of the Murray fault, the McKim Formation is tens of metres thick, whereas south of the fault, it probably exceeds 1000 m. The thickness and facies variations across the Murray fault system suggest that the faults represent south side down, syn-sedimentary, growth faults that were reactivated during compression attributed to the Penokean Orogeny (e.g., Zolnai et al. 1984).

In addition to stratigraphic thickness variations, the Murray fault system also marks profound changes in structural style, metamorphic grade and magmatic associations (Card et al. 1972). Deformation is more complicated and of greater intensity south of the fault. Likewise, metamorphic grade is higher immediately south of the fault (amphibolite facies transitional southward to greenschist facies) than to the north (greenschist to subgreenschist facies). South of the fault, there are several 1750 Ma and younger granitoid complexes (e.g., Cutler batholith) (Davidson and van Breemen 1994), whereas, north of the fault, there are no such intrusions.

Northeast of Coniston, the Grenville Front boundary fault and the Murray fault system are thought to merge into the Wanapitei fault (Davidson 1997), which can be traced into Street Township. This fault is then offset to the north by the Upper Wanapitei River fault and continues eastward as the Ess Creek and Grenville Front boundary faults along the trend of the Kabikotitwia and Sturgeon rivers (Easton and Murphy 2002). Thus, Huronian Supergroup strata located west and northwest of the Wanapitei and Ess Creek faults occur north of the Murray fault system, whereas

any Huronian Supergroup rocks preserved within the Grenville Province would have originally been deposited south of the Murray fault system.

The Grenville Province

Introduction

Rocks of the Grenville Province in Ontario range in age from circa 2690 to 990 Ma. All rocks older than 1300 Ma are pre-Grenvillian, whereas those younger than 1300 Ma are Grenvillian. With respect to nomenclature, a variety of subdivisions are in use for the Grenville Province in Ontario and fall into 2 broad groups: those that are lithologically based, commonly with a long history of usage (e.g., Wynne-Edwards 1972); and those that are more tectonic or interpretative in character, generally of more recent vintage (e.g., Rivers et al. 1989; Carr et al. 2000). Geological domains and their boundaries between the different types do not always coincide from one scheme to another (e.g., the Central Gneiss Belt contains para-autochthonous and allochthonous rocks), however, both approaches are valid, and usage is based on needs (e.g., lithologic- and historic-based terminology may be used more on detailed maps (<1:50 000 scale), tectonic-based terminology may be used on regional maps and in academic literature). Key divisions of the Grenville Province are listed in Table 2.

The field trip route mostly lies within the northernmost part of the Grenville Front tectonic zone (GFTZ), but also includes some rocks in northernmost Nepewassi domain. Both are part of the parautochthonous belt or the Laurentian Margin, which is defined as that part of the Grenville Province in which the rocks, although thoroughly reworked during the Grenvillian and/or earlier orogenies, can be reasonably equated with rocks of older Shield provinces to the northwest (Rivers et al. 1989; Carr et al. 2000).

These rocks are considered parautochthonous in the sense that they were once part of the autochthon, having been subsequently transposed and uplifted northwestward. It is difficult to recognize such rocks directly in many places along the Grenville Front This is due not only to the effects of intense reworking within the Grenville orogen, but also to juxtaposition, on opposing sides of the front, of rocks that were originally at different crustal levels, thus exhibiting different states of deformation and metamorphism, and not necessarily representing the same lithologic units. Geochronology has been of inestimable value in making broad correlations across the front. Lack of identification of deformed and metamorphosed equivalents in the Grenville Province of the flat-lying supracrustal rocks (e.g., Huronian Supergroup) northwest of the front is interpreted to be due to uplift and erosion of these successions, so that only their substrate is preserved. This field trip specifically examines these issues.

The extent of Superior and Southern province rocks within the Grenville orogen can be documented by Nd depleted mantle model ages. Dickin and McNutt (1989) found that Archean and Paleoproterozoic model ages of gneisses are restricted to northwest of a line extending from Key Harbour to Timiskaming, well southeast of the GFTZ and some 60 km from the Grenville Front. It can be argued, however, that meta-sedimentary rocks younger than the Huronian Supergroup (>2.2 Ga) could equally well have had Archean provenance. Gneisses southeast of this line have distinctly younger Nd model ages (*circa* 1.9 Ga).

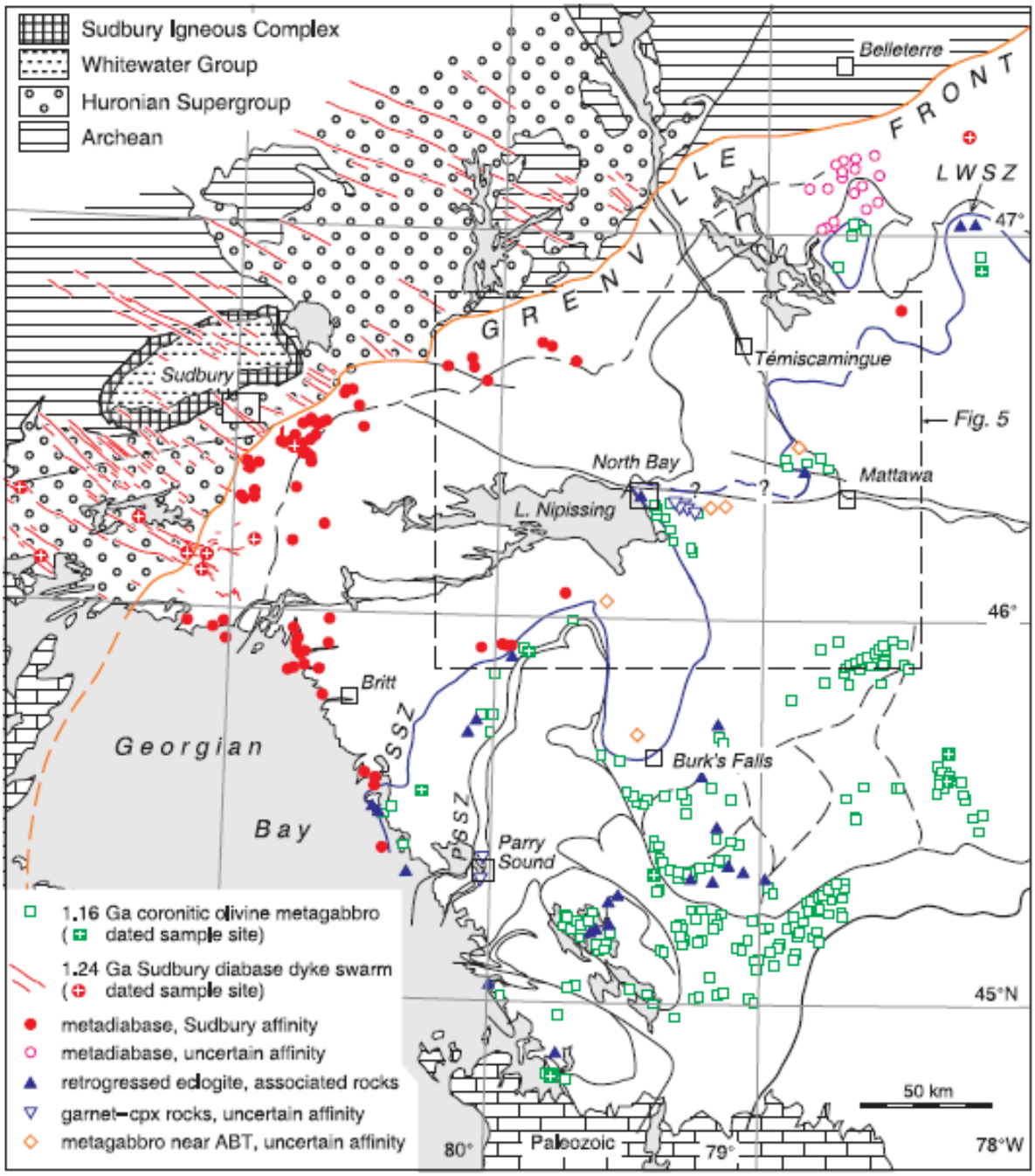


Figure 3. Distribution of 1.24 Ga Sudbury diabase and metadiabase, -1.17–1.15 Ga coronitic olivine metagabbro, and eclogitic rocks in the Central Gneiss Belt, Ontario and westernmost Quebec. The broken line near the Grenville Front is the southeast margin of the Grenville Front tectonic zone (Wynne-Edwards 1972). Figure *from* Ketchum and Davidson (2000).

Table 2. Key divisions and boundaries within the Grenville Province in Ontario.

Historic/Lithologic	Regional Tectonic	Local Tectonic/Historic
Grenville Front Tectonic Zone (GFTZ)	Para-autochthonous belt (Rivers et al. 1989) or Laurentian margin 1 (Carr et al. 2000)	Segments 1, 2, 3
Central Gneiss Belt (CGB) (Wynne-Edwards 1972; Easton 1992)	Para-autochthonous and/or allochthonous belt (Rivers et al. 1989), Laurentian margin 2 and 3 (Carr et al. 2000)	Parry Sound, Algonquin, Tomiko, Beaverstone terranes Britt, Fishog, Go Home (lower), Go Home (upper), Huntsville, Kiosk, McCraney, McClintock, Moon River, Nepewassi, Novar, Powassan, Shawanaga, Sequin, Tilden Lake domains
Central Metasedimentary Belt (CMB) (Wynne-Edwards 1972; Easton 1992)	Composite Arc Belt (CAB) and Frontenac-Adirondack Belt (FAB) (Carr et al. 2000)	Bancroft, Elzevir, Frontenac terranes (Elzevir contains Anstruther, Belmont, Grimsthorpe, Mazinaw, Sharbot Lake domains), Adirondack Lowlands and Highlands
Important Boundaries		
Grenville Front (Wynne-Edwards 1972; Easton 1992)	North limit of Grenville metamorphism and penetrative deformation (locally migmatite front)	Grenville Front boundary fault (GFBF)
Allochthon Boundary Thrust (ABT) (Rivers et al. 1989)	Separates para-autochthonous and allochthonous rocks (Rivers et al. 1989; Carr et al. 2000)	a.k.a. central Britt shear zone, Shawanga shear zone
Laurentian Margin - Composite Arc Belt boundary (Carr et al. 2000)	Composite Arc boundary zone (CABZ) (Carr et al. 2000)	Central Metasedimentary boundary zone (CMBBZ), a.k.a Central Metasedimentary Belt boundary thrust zone (CMBbtz)
Composite Arc Belt – Frontenac-Adirondack Belt boundary (Carr et al. 2000)	Frontenac-Adirondack boundary zone (FABZ) (Carr et al. 2000)	a.k.a. Maberly shear zone, Sharbot Lake-Frontenac boundary

Nepewassi domain

The Nepewassi domain (Easton 1992) was discriminated from its neighbours on the basis of structural trends as well as rock types. The area underlain by the Nepewassi domain in the field trip area was mapped by Lumbers (1975) at 1:126 720 scale. The Nepewassi domain is underlain by compositionally heterogeneous migmatitic gneisses which have a polycyclic history. Plutonic rocks in the domain form 2 suites which are less deformed than their typically migmatitic host rocks (Lumbers 1975): an older, granite-monzogranite suite *circa* 1740 Ma that includes the migmatitic West Bay and the Sturgeon Falls batholiths, and a younger, non-migmatitic suite, *circa* 1450 to 1420 Ma, that includes the Cosby pluton. Near Alban, the Cosby pluton intruded a thick sequence of quartzite, known as the French River quartzite (cf. Lumbers 1975). Both plutonic suites represent an eastward continuation into the Grenville Province of the main igneous components of the Killarney Magmatic Belt, which is straddled by the Grenville Front in the Killarney area. Table 3 summarizes the ages from plutons of both suites in the Killarney Magmatic Belt, the Grenville Front tectonic zone, and the Nepewassi domain. Leucogabbro to anorthosite of the St. Charles and Mercer intrusions cut the West Bay batholith and were emplaced at *circa* 1225 Ma (Prevec 2004).

Limited geochronological data are available from the Nepewassi domain (Table 3). Tonalite-trondhjemite gneisses exposed between Hagar and Warren yielded U/Pb zircon ages of 2678 to 2683 Ma, with titanite indicating that regional metamorphism of these same gneisses occurred between 996 and 975 Ma (Chen et al. 1989). A grey gneiss located on Highway 535 north of Noelville yielded a similar age of 2680±11 Ma by laser

ablation on zircon (Van de Kerckhove 2016). These zircon ages are consistent with the Nd/Sm and Pb/Pb model ages of Dickin (1998a, 1998b) which indicate the presence of Archean crust throughout much of Nepewassi domain.

Along the western margin of the Nepewassi domain and the Grenville Front tectonic zone, a layered gneiss unit, the French River "paragneiss", yielded a zircon population with an age of 1744±11 Ma (Krogh 1989). The homogeneous nature of the zircon population (Krogh 1989) suggested that the French River "paragneiss" is not a typical clastic metasedimentary rock, but rather that it may have been derived from metamorphosed volcanic and/or volcaniclastic rocks. Alternatively, it may be a highly strained orthogneiss. In contrast, the French River quartzite contained only Archean zircons, with monazite giving a metamorphic age of 1062±15 Ma (Krogh 1989). Quartzites examined by Van de Kerckhove (2016) northeast of Noelville had detrital zircon populations ranging from 2563 to 2962, with peak populations between 2686-2702 Ma, consistent with detrital zircon populations from Huronian Supergroup rocks in the Southern Province (*see* summary in Easton 2019). Van de Kerckhove (2016) also reported metamorphic zircon and monazite ages of 1755±11 and 1761 Ma, respectively, from the same area, suggesting a regional metamorphic event coincident with Killarney belt magmatism.

The age of the major plutonic units in Nepewassi domain is poorly known, but many probably have affinities to the Killarney magmatic suite (*circa* 1740) (Easton 2014). The migmatitic, French River granite, located along the western boundary between the Nepewassi domain and the Grenville Front tectonic zone, and intruded into the French River "paragneiss", gave a robust Rb/Sr age as well as a U/Pb zircon age of *circa* 1700 Ma (Krogh and Davis 1969, 1972). The age of the texturally similar West Bay batholith near Lavigne has not been determined reliably, although a poor-quality laser U/Pb zircon age of *circa* 1255 was reported by Aldis (2016). A U/Pb zircon age of *circa* 1420 Ma has been reported from the Cosby pluton (Lumbers 1975, p.98), but neither the details on the age determination nor the location of the sample are available. Aldis (2016) reported a laser U/Pb zircon age of *circa* 1434 Ma from the Cosby pluton near Noelville, similar to the age reported by Lumbers (1975).

New observations from the northwestern Nepewassi domain, which will be seen during the field trip. include the identification of granulite-facies, green and pink, garnet-bearing, potassium feldspar megacrystic granodioritic gneiss of the Estaire pluton and incorporated pods of hypersthene-bearing gneissic diorite, a likely co-magmatic phase (Stop 7, 8). The Estaire pluton granodiorites are characterized by high Ba (>2000 ppm) and high Zr (>500 ppm) contents similar the those found in the West Bay pluton, south of Verner. In addition, quartzite (Stop 9), possibly correlative with the French River quartzite, occurs as a 2.2 km long, up to 300 m wide, belt on the south side of the Wanapitei River, only 2.2 km south of the boundary with the GFTZ.

The Grenville Front

The Grenville Front itself is a zone of southeast-dipping faults and mylonites and has generally been placed at the southeast limit of recognizable Southern Province rocks (e.g., Lumbers 1975; Davidson 1997). Locally there are complications that have led to many debates concerning the identity of the Grenville Front and its distinction from other faults that intersect, merge with or are parallel to the Front (*see discussion* in Davidson 1997). In central Street Township the Grenville Front (which is coincident with the Wanapitei fault) has been displaced to the north by the younger, north-trending, Upper Wanapitei River fault by at least 850 m of sinistral and west-side-up movement (Easton and Murphy 2000, 2002). This displacement likely occurred after *circa* 590 Ma, as a Grenville swarm diabase dike in northern Henry and Loughrin townships is also displaced by north-trending faults.

Table 3. Summary of geochronological data for Killarney area magmatic rocks, the Wanapitei complex and the Nepewassi domain.

Age (in Ma)	Unit	Comment	Method	Source
Killarney Area Magmatic Rocks				
$1749^{+12}/_{-8}$	biotite granodiorite	North of Chief Lake granite	U/Pb TIMS zircon	Davidson and van Bremen (1994)
1747±3	granodiorite	Eden Lake complex	U/Pb TIMS monazite	Sullivan and Davidson (1993)
1744±29	granodiorite	Eden Lake complex	U/Pb LA-ICP–MS zircon	Raharimahefa, Lafrance and Tinkham (2014)
1704±13	granitic dike	Cuts fabric and shear zone in Huronian Supergroup rocks near McFarlane Lake	U/Pb LA-ICP–MS zircon	Raharimahefa, Lafrance and Tinkham (2014)
1596±39	granitic dike	Cuts fabrics in Eden Lake complex	U/Pb LA-ICP–MS zircon	Raharimahefa, Lafrance and Tinkham (2014)
$1746^{+16}/_{-6}$	granodiorite	Cutler batholith	U/Pb TIMS zircon	Davidson, van Breemen and Sullivan (1992)
1742±1.4	granite	Killarney granite	U/Pb TIMS zircon	van Breemen and Davidson (1988)
1464±2	granite	Chief Lake granite	U/Pb TIMS zircon	Davidson and van Bremen (1994)
1467±18	granite	Chief Lake granite	U/Pb LA-ICP–MS zircon	Raharimahefa, Lafrance and Tinkham (2014)
1429	pegmatite dike	South of GF, Chief Lake area	U/Pb TIMS zircon	Krogh (1994)
1447	pegmatite dike	South of GF, Chief Lake area	U/Pb TIMS titanite	Krogh (1994)
1464	pegmatite dike	South of GF, Chief Lake area	U/Pb TIMS monazite	Krogh (1994)
Wanapitei Complex				
$1746^{+12}/_{-6}$	quartz monzonite dike	Wanapitei complex, cuts metagabbro	U/Pb TIMS zircon	Davidson, p.39 *in* Easton, Davidson and Murphy (1999)
1707±17	quartz monzonite dike	Wanapitei complex, cuts metagabbro	U/Pb LA-ICP–MS zircon	Rousell et al. (2012)
$1746^{+6}/_{-5}$, 996	hornblende metanorite	Wanapitei complex, lower intercept age of metamorphism	U/Pb TIMS zircon	Prevec (1993, 1992)
1735±3	garnet metagabbro	Wanapitei complex	U/Pb LA-ICP–MS zircon	Rousell et al. (2012)
1694±7, 1640±10	garnetiferous mafic dike	Wanapitei complex, cuts other units, 2 populations	U/Pb LA-ICP–MS zircon	Rousell et al. (2012)
Nepewassi Domain				
1744±11	French River "paragneiss"	GFTZ near western boundary of Cosby subdomain, 2 sample sites, single population	U/Pb TIMS zircon	Krogh (1989)
circa 1700, 1689±16	French River "granite"	GFTZ near western boundary of Cosby subdomain, migmatitic, cuts French River "paragneiss"	U/Pb TIMS zircon, Rb/Sr whole rock	Krogh and Davis (1969, 1972)
1420	Cosby pluton	Cosby subdomain, no location given	U/Pb TIMS zircon	Lumbers (1975)
1062±15	French River quartzite	Cosby subdomain, quartzite only had Archean zircons	U/Pb TIMS monazite	Krogh (1989)
1245±48	Mercer anorthosite	Southern subdomain, 1222±2 Ma *in* Prevec (1992)	U/Pb TIMS zircon	Prevec (2004, 1992)
1244±100	St. Charles anorthosite	Southern subdomain, 1206±36 Ma *in* Prevec (1993)	U/Pb SHRIMP zircon	Prevec (2004, 1993)
2678 to 2683	tonalite, granodiorite	Northern subdomain, range from 7 sample sites	U/Pb TIMS zircon	Chen, Krogh and Lumbers (1995); Van de Kerckhove (2016)
975 to 996	tonalite, granodiorite	Northern subdomain, age of metamorphism, range from 6 sample sites	U/Pb TIMS zircon, lower intercept	Chen, Krogh and Lumbers (1995)

Abbreviations: *GF, Grenville Front; GFTZ, Grenville Front tectonic zone; LA-ICP–MS, laser ablation inductively coupled plasma mass spectrometry; SHRIMP, sensitive high-resolution ion microprobe; TIMS, thermal ionization mass spectrometry.*

At the time of formation, the Grenville Front was probably equivalent to the Main Boundary thrust of the current Himalayan orogen, marking the boundary between lowlands to the north and high-standing mountains to the south.

Along most of its length in Ontario, the Grenville Front is characterized by a major, intense, moderately southeast-dipping mylonite zone a few metres to tens of metres thick. The mylonitic rocks have a dip-parallel lineation and kinematic indicators show a northwestward thrust sense. Rocks in the immediate foreland adjacent to front-parallel faults, which are generally steeper than the front mylonite zone, show cataclastic deformation. Gneissic and protomylonitic rocks southeast of the front show penetrative ductile deformation with the same northwest-directed thrust sense as the front mylonites. Thus, the Grenville Front marks a transition from brittle to ductile deformation toward the southeast.

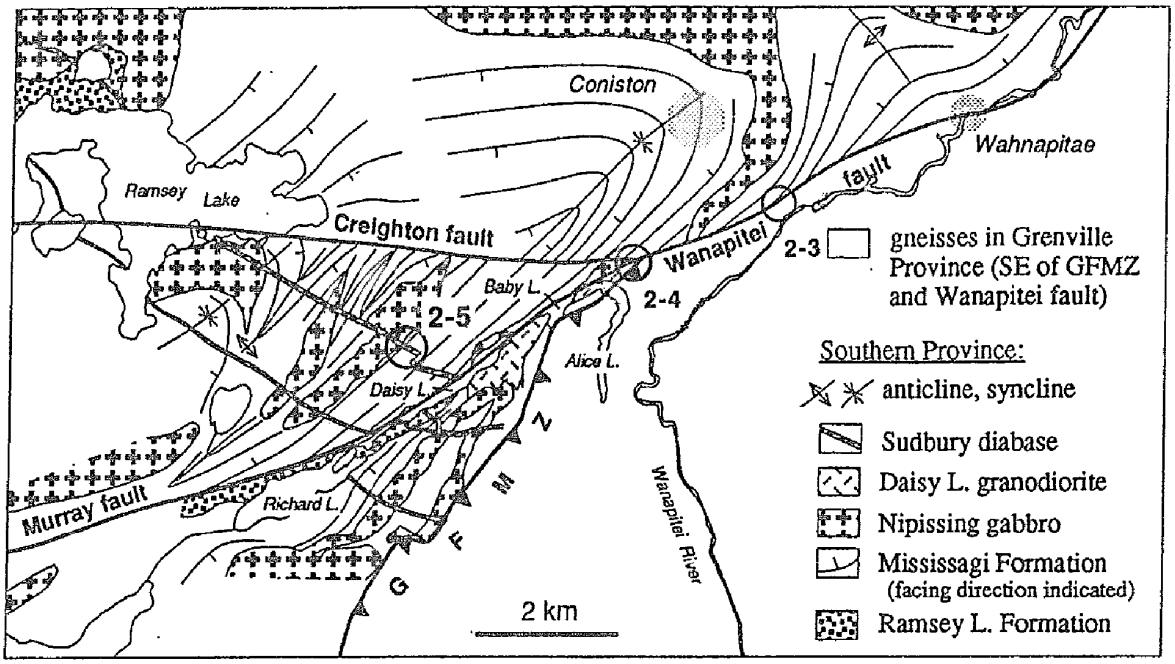
In many places, front-parallel mylonite zones occur within the gneisses southeast of the front, and both these and the front mylonitic rocks exhibit local, superimposed cataclasis. This demonstrates the changing nature of deformation, from ductile to brittle, during the time taken for the orogen to rise and, with accompanying exhumation, to cool.

With respect to the time taken for uplift, it is noteworthy that nowhere along the length of the Grenville Front is there any evidence that the foreland was ever the site of a basinal depression that received detritus from an elevated Grenvillian mountain belt (it is probably significant that sediments of suitable age that are part of the Midcontinent Rift fill have Archean or Penokean and not Grenvillian provenance). This can be explained through a combination of factors such as the length of time taken for exhumation, and the effect of crustal thickening within the orogen that may have allowed the foreland to remain isostatically buoyant (e.g., Jamieson and Beaumont 1989). Lack of a foreland basin would have allowed only ephemeral deposition of coarse detritus adjacent to the orogen, and finer detritus to bypass the foreland and to be spread widely farther afield, perhaps contributing to the fill of late Mesoproterozoic basins in the western and northern parts of North America (e.g., Hoffman and Grotzinger 1993; Rainbird et al. 1997).

In the field, the Grenville Front has generally been placed at the southeastern limit of recognizable Southern Province rocks (Lumbers 1975; Davidson 1997, 1998). Locally, however, there are complications that have led to many debates concerning the identity of the Grenville Front and its distinction from other faults that intersect, merge with, or are parallel to it (*see* discussion in Davidson (1995, 1997, 1998)). In the west, the Grenville Front can be traced from Georgian Bay across the Killarney Magmatic Belt until Coniston where it intersects the Murray and the Creighton faults.

The timing of isotopic closure in a variety of mineral systems adjacent to the Grenville Front in the Sudbury area was examined by Corfu and Easton (2000). Zircon closed at 995 to 987 Ma, similar to ages reported all along the Grenville Front from Killarney to Labrador by Krogh (1994); this age represents some of the youngest Grenvillian activity in Ontario. Titanite and monazite from the same localities record only slightly younger ages between 989 to 977 Ma, consistent with the slightly lower closure temperatures for these minerals. Rutile and apatite ages from the same samples record ages of 973 to 971 Ma and 959-932 Ma, respectively, consistent with slow cooling along the Grenville Front after last movement at *circ*a 995 Ma.

As mentioned, the Murray fault (Figure 1, 4, 5) is a major west-trending lineament, which locally separates weakly metamorphosed Huronian Supergroup rocks to the north from strongly metamorphosed and deformed, Mesoproterozoic-granite-bearing, Huronian Supergroup rocks to the south. Metamorphic contrast across the Murray fault is most pronounced in proximity to plutonic rocks (e.g., the Eden Lake and Cutler plutons). Northeast of Coniston (Figure 4, 5), the Grenville and Murray faults are thought to merge into the Wanapitei Fault (Davidson 1997). In central Street



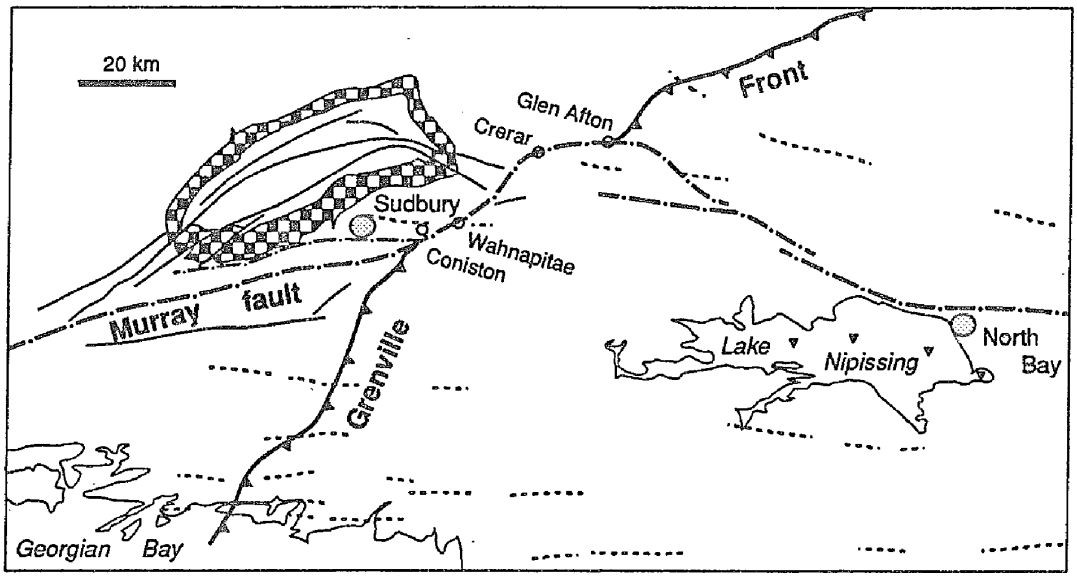


Figure 4. A) The juncture between the Grenville Front mylonite zone and the Murray–Wanapitei fault south of Coniston. Stop 2-5 is the same location as Stop 1 in this guide, Stop 2-3 is the same as Stop 3. Abbreviations: GFMZ, Grenville Front mylonite zone; L, lake.
B) Regional relationship between the Grenville Front, the Murray fault, and faults extending westward from the Ottawa-Bonnechere rift system. Dashed lines are Neoproterozoic Grenville swarm dikes, inverted triangles in Lake Nipissing are alkalic complexes associated with Neoproterozoic rifting. Both figures *from* Davidson (1995).

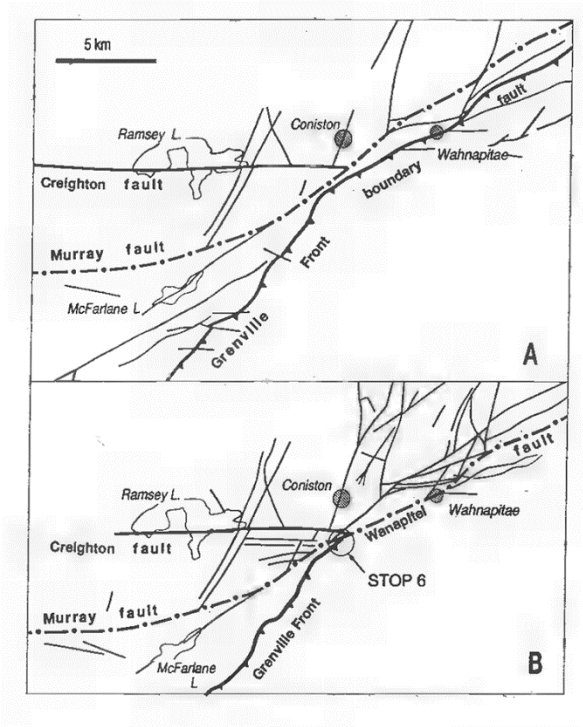


Figure 5. Fault interpretations in the Coniston area. **A)** *after* Lumbers (1975); **B)** *modified after* Dressler (1984). Stop 6 on the figure is located just southwest of Stop 3 in this guidebook. Figure *from* Davidson (1997).

Township, the Grenville Front (Wanapitei Fault) is displaced to the north along the younger, north-trending, Upper Wanapitei River fault by at least 850 m of sinistral and west-side-up movement (Easton et al. 1996; Easton and Murphy 2002).

The Grenville Front tectonic zone (GFTZ)

The Grenville Front tectonic zone (GFTZ) is a region up to 30 km across lying between the Grenville Front and the Central Gneiss Belt of the Grenville Province (Figure 1). Easton (1992) divided the Grenville Front tectonic zone in Ontario into 3 lithologic segments. The western segment between Killarney and Wahnapitae comprises rocks equivalent in age, geophysical signature, and rock type to the adjacent Killarney Magmatic Belt. The central segment stretches from Wahnapitae to River Valley and contains mafic intrusive rocks of both the East Bull Lake and the Nipissing intrusive suites, as well as several types of migmatitic gneisses, likely of Neoarchean age (Easton 2000). The eastern segment, between River Valley and the Ottawa River, includes rocks mainly derived from the adjacent Superior Province. The field trip area straddles the boundary between the central and eastern segments.

The GFTZ can be envisaged broadly as an anastomosing network of higher-strain rocks, moderately inclined and shallowing southeastward, surrounding lower-strain pods and lenses that are presumably elongate parallel to the prevalent southeast-plunging stretching lineation. The same style of structure can be seen in many places at outcrop scale and, if one considers porphyroclasts in mylonite, also at microscopic scale. Its southeast margin is ill-defined; as the front-parallel layered structure becomes shallower to the southeast; it also becomes progressively warped (buckled) about gentle, predominantly southeast-plunging axes so that the structural grain expressed at the surface changes from northeast to southeast. Map-scale enveloping surfaces, however, maintain a generally northeast trend.

In the Street Township area east of Sudbury, a rapid southeastward increase in metamorphism is present, with garnet-staurolite-kyanite developed within 800 m of the front and sillimanite-potassium feldspar rocks within 2 km. (Easton and Murphy 2002; Easton et al. 1999). Metamorphism and progressive disruption of the Sudbury swarm dikes occurs much closer to the front than to the southwest, and successive orthopyroxene-clinopyroxene-garnet coronas are fully developed in non-deformed cores of dismantled lenses within a kilometre of the front. Northwest of the front, shaly interbeds in the Mississagi Formation (predominantly cross-bedded feldspathic arenite) and shales of the underlying Pecors Formation are low-grade phyllites (muscovite-chlorite-albite-quartz) 800 m from the Grenville Front. Nipissing gabbro contains the assemblage epidote-actinolite-chlorite-albite ± quartz.

Despite its proximity to Sudbury, no rocks that can be related to the Sudbury impact event (e.g., metamorphosed Sudbury breccia; Offset dikes, etc.) have been reported from the GFTZ.

The GFTZ zone near Wanup

A key observation from the 2021 mapping program by the author is that the Grenville Front tectonic zone in the Sudbury area displays different structural styles whether one is west, or east, of the Wanapitei River. Most of the stops on the field trip are in the area west of the Wanapitei River, primarily for logistical reasons.

West of the Wanapitei River

West of the river, lithological units consist mainly of highly-strained gneisses, typically migmatitic (Photo A, B), that form thin, near continuous belts interlayered with migmatitic amphibolite, amphibolite and garnet amphibolite. Possible metasedimentary units, including schists containing aluminosilicate minerals, calc-silicate gneisses, and minor impure marble (Photo C), form thin, lenticular, discontinuous units that occur locally within the package of highly-strained gneissic and amphibolitic units. Rocks of the East Bull intrusive suite are locally present, but form thin, discontinuous units. Sudbury swarm dikes are large, with strike lengths of 50 to 100 m. In addition, the boundary with the Nepewassi domain is only 8 km from the Grenville Front.

Pegmatite dikes are common west of the Wanapitei River and are predominantly granitic. Some are deformed and concordant or near-concordant, with gneissosity, whereas others are highly discordant. Large, late, niobium-yttrium-fluorine (NYF), variably-zoned, discordant, granitic pegmatite dikes (e.g., Stop 6a, 6b) are common in the area west of the river, and many have been quarried in the past, mainly for feldspar and/or mica (Vos et al. 1981). Narrow (0.5 to 3.0 m wide), garnet-baring, fine-grained leucogranite pegmatites are abundant in the GFTZ near the boundary with Nepewassi domain (Photo D, E) .and were not observed by the author east of the river. The late NYF pegmatites are younger than the Sudbury dike swarm. There is no relative age constraint on the leucogranite pegmatites, although they are metamorphically recrystallized.

East of the Wanapitei River

In contrast, east of the river, several large roundish or ovoid plutonic bodies are present, likely related to the Killarney Magmatic Belt, including the Wanapitei Complex (6 x 2 km in size) and the Cleland stock (3 x 3 km in size). East Bull Lake intrusive suite rocks former larger, more continuous, and better-preserved bodies, including the Red Deer Lake intrusion (11 km long, up to 1 km wide) and a body present along the southeast margin of the Wanapitei Complex. Host gneisses are dominantly quartzofeldspathic and contain. garnet-in both melanosomes and leucosomes (Photos F, G, H, I), and are similar to many of the gneisses present in Street Township only a few kilometres to the north-northeast. Also present are deformed, lenticular granitoid bodies ("meta-arkoses" of Lumbers 1975), which in map pattern appear to define broad folds (see maps of Lumbers 1975; Dressler 1984). Similarly, the kyanite-bearing paragneiss units west of Wahnapitae (Grant et al. 1962; Pearson 1959; Easton and James 1997) are more continuous, and better preserved, than possible correlative schistose rocks to the west of the Wanapitei River (Stop 13). Calc-silicate gneisses and impure marbles have not been reported from the area east of the river. Sudbury swarm dikes are pod-like, with minimal strike lengths. Late granite pegmatites seem to be less abundant in the area east of the river. Finally, the boundary with the Nepewassi domain is approximately 12-15 km from the Grenville Front.

Structure along the Wanapitei River

The feature causing the observed lithological differences across the Wanapitei River has a linear, north-northwest trend and appears as a weak linear magnetic feature in the low-resolution magnetic data available for the area (Figure 6, upper). The course of the Wanapitei River coincides with this structure from Coniston to Estaire, which obscures direct examination of the rocks immediately



Photo A. Rocks west of the Wanapitei River. Folded granitic leucosome in migmatitic, garnet-bearing, gneissic diorite at north end of the roadcut. West side of Highway 69/400, 1.8 km from the Grenville Front (UTM 509315E 5141309N).



Photo B. Rocks west of the Wanapitei River. Most of rock face is relatively massive, leucosome-poor migmatitic, garnet-bearing, gneissic granodiorite, however there is a zone that is much more leucosome-rich in the lower, centre part of the photo. West side of Highway 69/400, 1.8 km from the Grenville Front (UTM 509366E 5141236N).



Photo C. Rocks west of the Wanapitei River. Impure dolomitic marble and calc-silicate layers in southerly of two calc-silicate and marble bands on the west side of Highway 69/400, northside of Old Wanup Road overpass. Scale card is 9 cm long (UTM 510571E 5140422N).



Photo D. Rocks west of the Wanapitei River. Most of rock face is relatively massive, leucosome-poor migmatitic, garnet-bearing, gneissic granodiorite, however there is a zone that is much more leucosome-rich in the lower, centre part of the photo. West side of Highway 69/400 (UTM 5111057E 5137014N).



Photo E. Close-up of white, garnet-bearing, near concordant white pegmatite dike shown in Photo D. Scale card is 9 cm long.



Photo F. Rocks east of the Wanapitei River. Close-up view of layered felsic metatexite gneiss at this station. Outcrop is on the west side of St. Cloud Road. Scale card is 9 cm long (UTM 515775E 5139250N).



Photo G. Rocks east of the Wanapitei River. Layered felsic metatexite gneiss at this station. Outcrop is on the west side of St. Cloud Road. Hammer for scale, handle is 33 cm long (UTM 515544E 5138949N).

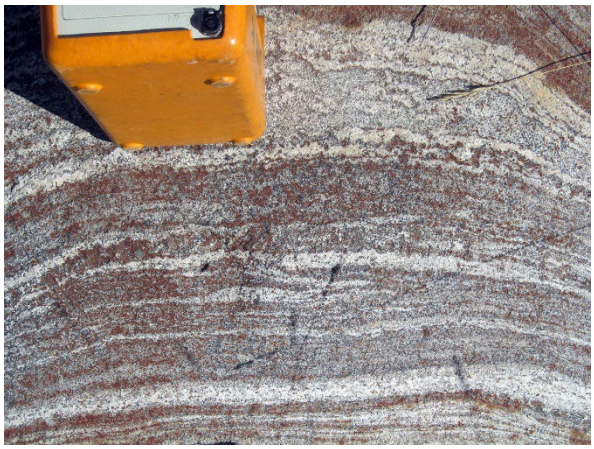


Photo H. Rocks east of the Wanapitei River. Layered garnet-bearing felsic metatexite gneiss at this station. Outcrop is on the west side of St. Cloud Road. Scintillometer for scale, instrument is 23 cm long, back of scintillometer is 10 cm square (UTM 515544E 5138949N).



Photo I. Rocks east of the Wanapitei River. Layered felsic metatexite gneiss at this station. Outcrop is on the west side of St. Cloud Road. Hammer for scale, handle is 33 cm long.

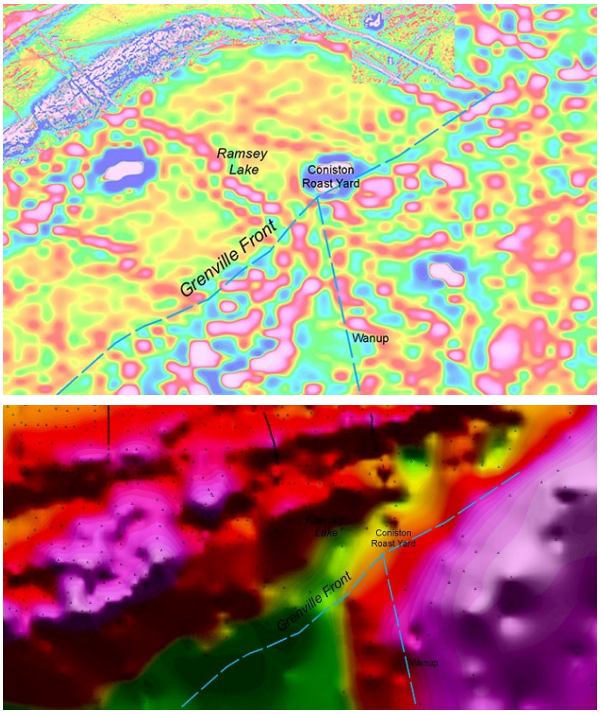


Figure 6. Upper. First vertical derivative of the residual magnetic intensity of the Wanup area showing the location of Wanup, the Grenville Front, and the north-northwest structure along the Wanapitei River. Northwest linear magnetic highs west of the Grenville Front are Sudbury diabase dikes. **Lower.** Bouguer gravity field.

adjacent to the structure. The feature is also apparent in the Bouguer gravity data, separating a broad gravity high on the east side of the river from an area of lower density rocks to the west (Figure 6, lower). Which is interesting, as many of the mafic and granulite facies rocks west of the river have specific gravity values of 3.0 and 3.2 g/cm^3.

The north-northwest-trending structure is located where there is a flexure in the trace of the GFTZ, from north-northeast from Killarney to Wanup, to northeast from Wanup eastward.

A preliminary explanation for the lithological and structural differences across the north-northwest-tending structure would be that different structural levels are exposed on either side, with a likely deeper, and possibly hotter level on the west compared to the east. The greater degree of migmatization, and the presence of granulite facies rocks in the Nepewassi domain west of the Wanapitei River are consistent with higher temperatures. Regardless, more work needs to be done on rocks from both sides of the north-northwest-trending structure to understand its origin and tectonic history.

Granite Pegmatites in the GFTZ

Ercit (1999) demonstrated that most granitic pegmatites in the northern Grenville Province, regardless of age, are classified as niobium-yttrium-fluorine (NYF) type pegmatites with no parental granite, indicating an anatectic origin and post-kinematic emplacement.

Lumbers (1975) recognized two types of granitic pegmatite intrusions in the field trip area. Gneissic, deformed, granitic pegmatite intrusions composed mainly of quartz and potassium feldspar with minor mica and amphibole that occur as discontinuous dikes and sills. Zircon from one of these dikes in Cleland Township, east of the Wanapitei River, yielded an age between 1600 and 1700 Ma (Krogh and Davis 1970; Lumbers 1975), suggesting an affiliation with the Killarney magmatic suite.

More common are late, post-metamorphic granitic pegmatite intrusions (*circa* 1000 Ma) that are distinctly zoned, commonly with quartz-rich cores (Stop 6b). Some are locally radioactive because of the presence of allanite (Stop 6a). Many have been quarried in the past, mainly for feldspar and/or mica (Vos et al. 1981).

Acknowledgements

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FIELD TRIP DETAILS

Geological Maps

Geological compilation maps covering all or parts of the area of the field trip include Ames et al. (2005), Dressler (1984), Lumbers (1975) and Card and Lumbers (1973).

ROAD LOG

Note: Caution should be taken when parking vehicles on the shoulder of the highway and when examining outcrops located along Highway 17 and on other roads along the field trip route. All UTM co-ordinates are given in NAD 83 datum, zone 17.

Leave from Science North at the junction of Paris Street and Ramsey Lake Road in Sudbury. Head south on Paris toward highway 69.

0.0 km — Junction of Highway 69 and the southeast and southwest bypass. Turn right onto the eastbound ramp and proceed east on Highway 17 after merging on to the Highway.

6.3 km — pull over on the right shoulder by the 1552.0 kilometre sign. Walk ahead (east) onto the roadcut on the south (right) side of the road. This locality is Stop 2-3 in Davidson (1995), Stop 1 in Davidson (1997); Stop 1-1 in Davidson et al. (2002).

Stop 1. Mississagi Formation sandstone, Nipissing gabbro sill and a Sudbury swarm olivine-diabase dike

UTM co-ordinates 508085E, 5144868N

This stop lies within the Southern Province 900 m northwest of the Murray fault and approximately 3 km west-southwest of Stop 3. It is only two kilometres from the Grenville Front (*see* Figure 4a). Here a southeast-trending, vertical, 65-m-thick olivine diabase dike of the 1235 million-year-old Sudbury dike swarm cuts across the contact between Mississagi sandstone (south side of the road) and Nipissing metagabbro (north side).

Crossbedding in the sandstone indicates that the steeply dipping beds face north-northwest. In fact, we are on the same limb of the Coniston syncline that we will be on at Stop 3 (*see* Figure 4a).

Cross the road and walk slightly west to the roadcut on the north side. The roadcut and rounded outcrops north of the road consist of Nipissing metagabbro that is massive but metamorphosed, containing secondary green amphibole, epidote and chlorite. The metagabbro also has low magnetic susceptibility. This metamorphic assemblage is typical of Nipissing gabbro throughout the Southern Province and thus does not appear to be a function of proximity to the Grenville Front.

Cross the road again back to the south side. Here we see the northeast contact of the Sudbury diabase dike. Blocks containing its chilled contact with Mississagi sandstone can be found at the roadside. Near this contact the diabase is fine grained and contains xenocrysts of plagioclase which in turn include earlier-crystallized olivine crystals — a common feature at the margins of Sudbury dikes (Bethune 1997; Bethune and Davidson 1997). There is absolutely no evidence in thin section of any metamorphic reaction between these two minerals at this location. Note that this dike is unmetamorphosed and has high magnetic susceptibility. The Sudbury dikes are chemical distinct compared to other dike swarms in the region, and are characterized by TiO_2 >2.5 wt.%, >700 ppm barium and >300 ppm Zr (Ketchum and Davidson 2000). This distinctive chemistry allows for these dikes to be recognized south of the Grenville Front, where they serve as important markers of deformation and metamorphic history (Figure 3).

In the same dike that we see here, but farther to the south on the southeast side of the Murray fault, plagioclase xenocrysts become clouded with fine epidote, and rims of fine actinolite appear between olivine and plagioclase grains. Where Sudbury dikes have been identified in the immediate hanging wall of the Grenville Front, olivine in plagioclase xenocrysts has reaction coronas of pale orthopyroxene with outer rims of pargasite-spinel symplectite, and Ti-Fe oxide grains are surrounded

by Ti-biotite and garnet symplectite. Near the Grenville Front to the southwest, Sudbury dikes cut across and are chilled against a pre-existing mylonitic fabric that is developed in granitoid rocks as young as 1470 Ma (Bethune 1997; Davidson and Ketchum 1993), pointing to the existence of some kind of pre-Grenvillian tectonic front roughly coincident with the Grenville Front *sensu stricto*. In this regard it is pertinent that, in the hanging wall within a few kilometres of the front, metamorphic monazite in pelitic gneiss (Dudas et al. 1994) and zircon from pegmatitic leucosomes (Krogh 1994) record an age of *circa* 1445 Ma.

Return to vehicles and continue east on the bypass.

9.0 km — pull over on the right shoulder by just after the curve on the way up the hill. Examine outcrops on the east (right) side of the road.

Stop 2. Shattercones in Mississagi Formation sandstone

UTM co-ordinates 509418E, 5147073N

Sudbury is famous for its shatter cones, which are well exposed in units of the Huronian Supergroup, especially the Mississagi Formation. The roadcut exposes numerous large shatter cones developed in quartz arenite of the Mississagi Formation. This outcrop is best visited in late afternoon, where the evening sun provides excellent lighting. Note that the shatter cones are not isolated individuals. The whole outcrop is full of shatter cones, something that is not revealed on a polished glaciated surface.

The shatter cone collar around the Sudbury structure forms a near continuous ring extending up to 20 km distant from the contact between the footwall rocks and the Sudbury Igneous Complex. Prior to regional deformation and folding, most shatter cones pointed upward, as their impact source origin was from above. Despite statements to the contrary, no volcanic blast has ever formed a shatter cone collar, and nuclear blasts have at best produced small (cm-size) poorly defined cones (Spray et al. 2007).

Return to vehicles and continue east on the bypass.

11.2 km — Junction of the bypass and Highway 17, take the right merge lane onto Highway 17 and head east toward Coniston.

14.1km — Junction to Coniston (traffic light), continue east on Highway 17.

16.8 km — Junction with Highway 17 and the Coniston Hydro Dam Road just past the overpass over the railway tracks. Turn right onto Coniston Hydro Dam Road. Between Highway 17 and the parking area, the road crosses feldspathic sandstone beds of the Mississagi Formation.

18.4 km — Park in pullout area opposite the gates to the Coniston Hydro Dam. We will walk to the first series of outcrops which are on the north side of the railway tracks (UTM 513570E, 5146857N). This is Stop 2-5 of Davidson (1995), Stop 7 of Davidson (1997), Stops C-1 and C-2 of Easton et al. (1999), and Stop 1-2 of Davidson et al. (2002).

Stop 3a. Mississagi Formation sandstone, Southern Province side of the Grenville Front

Two major faults of the Murray fault system in the Southern Province, the Creighton fault and the Murray fault itself (Card 1978), converge eastward and meet just north of Alice Lake, 2 km west of here (*see* Figure 4, 5). East-northeast of this juncture, a narrow valley in line with the Murray fault marks the Grenville Front. North of this valley are well-preserved Huronian Supergroup sandstones (Mississagi Formation) and Nipissing gabbro (not observed at the stop) at low metamorphic grade, and south of it, high-grade migmatitic gneisses of the Grenville Province. The covered interval between the two conceals the Wanapitei fault, and is as little as 25 m wide in places between Alice Lake and the village of Wahnapitae, 6 km to the northeast. There is no field evidence to suggest that the Murray and Wanapitei faults are not one and the same, contrary to

published maps of this area (Lumbers 1975; Dressler 1984); the two interpretations are illustrated in Figure 5.

Mississagi sandstone at this stop displays obvious primary sedimentological features — crossbedding is well preserved and indicates that the beds face the same way as they dip, namely steeply to the northwest. Cleavage in the silty interbeds dips steeply southward and is axial planar to a major, southwest-plunging syncline whose surface trace lies about 2 km to the northwest (*see* Figure 4a). Metamorphic grade is low: cleavage in the silty interbeds is given by aligned sericite; thin sections show the presence of minor greenish biotite, indicating that the grade is probably no higher than middle greenschist facies. To the east, the trace of the Wanapitei fault lies in the valley along which a power line runs; to the west it passes just south of the slag heaps that can be seen in the distance.

Walk back to the outcrop by the road on the opposite side of the valley and railway line.

Stop 3b. Gneisses on the Grenville Province side of the Grenville Front

UTM co-ordinates 5135633E 5146768N

The outcrop on the south side of the railway crossing is composed of migmatitic quartzo-feldspathic, mafic and minor pelitic gneiss, and includes narrow mylonite zones that diverge southwestward from the Murray-Wanapitei fault line. The pelitic gneiss contains kyanite and sillimanite, and amphibolite contains garnet, attesting to middle to upper amphibolite facies.

This outcrop exposes a highly deformed mix of granitoid and hornblende gneiss, some with garnet, cut by mylonite zones whose rotated feldspar porphyroclasts indicate south-side-up displacement. These rocks clearly represent an entirely different crustal level to that exposed just 70 m to the north, which implies several kilometres of vertical displacement along the Wanapitei fault, provided that the metamorphism in these rocks is younger than that in the Mississagi Formation. In Street Township to the northwest, geochronology by Corfu and Easton (2000) indicates that the metamorphism is indeed Grenvillian and culminated at *circa* 995 Ma.

Return to vehicles. Retrace route back to Highway 17 and continue east on 17.

20.2 km — Highway 17 and Coniston Hydro Dam Road, turn right and continue east on Highway 17.

22.7 km — Bridge over the Wanapitei River in Wahnapitae village. The Wanapitei fault which, as at Stop 3, marks the Grenville Front in this area, lies in the river valley. Bare hills on the north (left) side of the river are underlain by well-bedded Mississagi sandstone that faces northwest, away from the front, similar to what we observed at Stop 3. The large roadcut to the right (south), just past the bridge and the variety store exposes kyanite-bearing metasedimentary gneiss and mafic gneiss (garnet-bearing amphibolite) derived from gabbro; both of which are cut by coarse-grained pegmatite, itself deformed.

27.2 km — Junction Highway 17 and 537, continue east on Highway 17.

31.5 — Junction Highway 17 and Sunset Road.

32.6 — Pull off onto gravel area on the south side of the Highway. This locality is Stop 1, Day 3 in Easton, James and Jobin-Bevans (2010).

Stop 4 (Optional). Shear-Zone Hosted Orthopyroxene Hornblendite Body

UTM co-ordinates 525451E 5152056N

Examine the outcrop and large blasted boulders present on the west side of the pullout. They belong to an orthopyroxene hornblendite body of the East Bull Lake intrusive suite that is present within a high-strain zone that extends subparallel to the highway. Examples of these highly strained felsic gneisses can be examined in outcrops at the base of the hill east of the pullout. The top of the ridge south of the road and above the stop consists of layered leucogabbronorite of the East Bull Lake intrusive suite. Thus, although proximal to rocks of the suite here, the orthopyroxenite body is not

directly in contact with other rocks of the suite, although that relationship has been observed elsewhere in Street and Awrey townships (Easton and Murphy 2002). Note the large, equant, orthopyroxene crystals, and the fine-grained amphibole matrix. Mineral chemistry indicates that the amphiboles are magnesium hornblende, tremolite and cummingtonite (Buckley et al. 1997; Easton and Murphy 2002). The amphiboles occur as individual grains and complexly exsolved and intergrown grains. Olivine grains from a body located closer to the Grenville Front are Fo_{73}. Orthopyroxene compositions lie in the bronzite field and are magnesium-rich (En_{75-80}). (Easton and Murphy 2002).

Turn around an retrace route westward on Highway 17

32.7 km — Junction Highway 17 and Sunset Road. Turn right (north) onto Sunset Road and proceed for 400 m. Pull over and park on the left shoulder of the road at the entrance to MTO gravel pit 402002 (phone 705-477-5478). Walk to the well exposed outcrops in the central part of the pit. This stop was used by Davidson (1997) but was not included in the guidebook descriptions.

Stop 5. Gneisses in the Grenville Province 1100 metres south of the Grenville Front

UTM co-ordinates of gate 524338E 5152230N

A variety of predominantly migmatitic gneisses are exposed in the well-exposed outcrops in the floor and northwestern wall of this pit. We are approximately 1 km south of the Grenville Front at this stop The grey, garnet-rich, gneisses may have been metasedimentary rocks. In contrast, a variety of mafic rocks, some garnet-bearing, some not, are interlayered with the grey gneisses. Some of these mafic rocks are large rafts in the gneisses (Photo 1), whereas others are thinner, boudinaged and aligned, and may represented dismembered dikes. Garnet is abundant and occurs in both the melanosome and leucosomes of the gneisses.

Proceed to the most northeastern exposed outcrop. While doing so, examine the gneisses closely and attempt to assess possible protoliths. Are these metasedimentary or metavolcanic rocks or are they highly deformed orthogneisses? Once at the northeast end of the outcrop, the protolith of some of these gneisses will become readily apparent, due to an area of lower-strain present in a macroscopic fold nose (Photo 2).

This locality illustrates the perils of trying to identify protolith in many gneissic terranes, especially in areas of poor, incomplete, and lichen-covered exposures. Remember, we are only a kilometre south of the Grenville Front at this locality, and only at upper amphibolite conditions. What cannot be ascertained at this stop is whether these gneisses reflect a rather simple metamorphic history, for example an Archean high-grade metamorphic event followed by reworking during the Grenville, or multiple metamorphic episodes throughout the Archean, the Paleoproterozoic and the Mesoproterozoic.

Return to vehicles and retrace route back to Highway 17.

34.4 km — Turn right and head west on Highway 17 toward Sudbury.

43.3 km — Junction Highway 17 and 537 in Wahnapitae just east of the bridge. Turn left onto 537 and head south.

48.8 km — Metamorphosed Sudbury swarm diabase dike is exposed on the west side of the roadcuts which consist mainly of Grenville gneisses.

60.5 km — Junction in Wanup, continue west (straight) toward Highway 400-69.



Photo 1. Grey migmatite with mafic pods at Stop 5.



Photo 2. Fold in grey migmatite at Stop 5 (524200E 5152280N). In the nose of the fold, even though the rock is still deformed and metamorphosed, it is clear here that the protolith of the rock was a matrix-supported conglomerate of unknown stratigraphic affinity.

61.8 km — Pull over and park in the pullout area on the right side of the road. This stop will examine the two large roadcuts to the west and east of the parking area.

Stop 6. Gneisses in the Grenville Province 6 kilometres southeast of the Grenville Front

Pullout 512055E 5137085N

The pullout splits a near continuous road cut, approximately 800 m long, and up to 15 m high, into eastern and western halves. The road cut was created during the process of four-laning Highway 69 and is located just to the south of the part of Dill Township that was mapped in detail by Kwak (1968) and Davidson (1997, 1998). This road cut is illustrative of what rocks look like in the Grenville Front tectonic zone only 7 km south of the Grenville Front. The key takeaway is that is near impossible to easily relate the rocks that we see here to any of the rock units that we see on the north side of the front in either the Southern or the Superior provinces.

In contrast to Stop 5, which appears to be dominated by an abundance of metasedimentary gneisses, the large roadcuts to the west and the east are more representative of much of the Grenville Front tectonic zone in the Sudbury area, where mafic gneisses predominant, albeit with slivers and layers of rocks that may have originally been metasedimentary. In examining the mafic gneisses in the two roadcuts, pay attention to features such as degree and style of leucosome formation; the presence or absence of garnet, and the abundance of garnet in some of the mafic gneisses, which far exceeds what would normally be generated in a mafic rock during a single-stage metamorphic event. We will examine the west roadcut first.

Stop 6A. West Roadcut. From west to east the roadcut consists of:

- Approximately 100 m of interlayered grey to locally rusty, thin layered paragneiss and weakly layered, variably migmatitic mafic gneiss and deformed granitoid layers
- 13 m wide outcrop gap
- Approximately 25 m of variably migmatitic, texturally varied mafic gneiss. This unit hosts a thin sub-horizontal pegmatite.
- Approximately 55 m of gneissic gabbro, with distinct east and west contacts. Along the western contact, relict large plagioclase crystals occur in the groundmass as dark equant to lath shaped crystals and as isolated crystals up to 20 mm long (Photo 4c). Energy dispersive X-ray analysis indicates that the large crystals are andesine, with a composition of An_{46}. Plagioclase xenocrysts are common occurrence in Sudbury swarm dikes in the Sudbury area (Stop 1; Davidson 1997, p.16). The presence of large plagioclase crystals only along the western contact

may simply reflect better preservation at that locality, as the gneissic gabbro at the contact has moderate magnetic susceptibility (1.1 to 2.3 x 10^{-3} SI units) compared with the rest of the body (<0.75 x 10^{-3} SI units). The fact that the gneissic gabbro is likey a metamorphosed Sudbury swarm gabbroic dike is confirmed by geochemistry, as these samples have the high TiO_2 (>2.5 wt.%), barium (>700 ppm) barium and Zr (>300 ppm) contents typical of the Sudbury swarm (analyses 1-4, Table 4).

The gneissic gabbro hosts a near-vertical feldspar-rich pegmatite dike (Photo 3). The eastern part of the pegmatite dike contains bluish apatite crystals (Photo 4a) and allanite (Photo 4b). Scintillometer readings from the eastern part of the dike range from 50 to 98 ppm U and 130 to 208 ppm Th (Easton, unpublished data). The age of the pegmatite is not known, but is younger than *circa* 1240 Ma, the age of the host Sudbury dike.

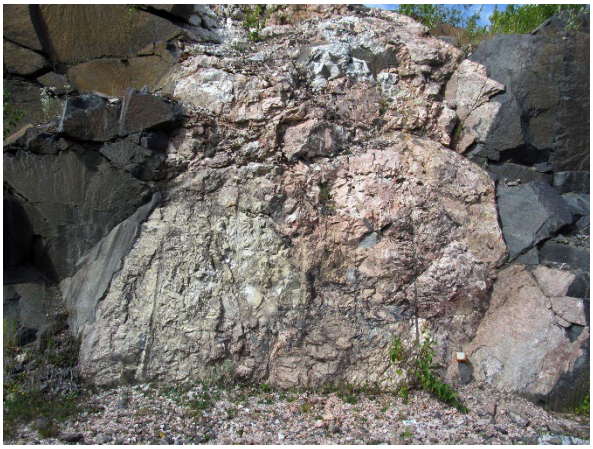


Photo 3. Near-vertical pegmatite dike cutting gneissic gabbro of the Sudbury dike swarm at Stop 6a. Dike is unevenly zoned, with the right (west) side dominated by potassium feldspar, and the left (east) side being more radiogenic and containing more quartz, allanite and apatite.

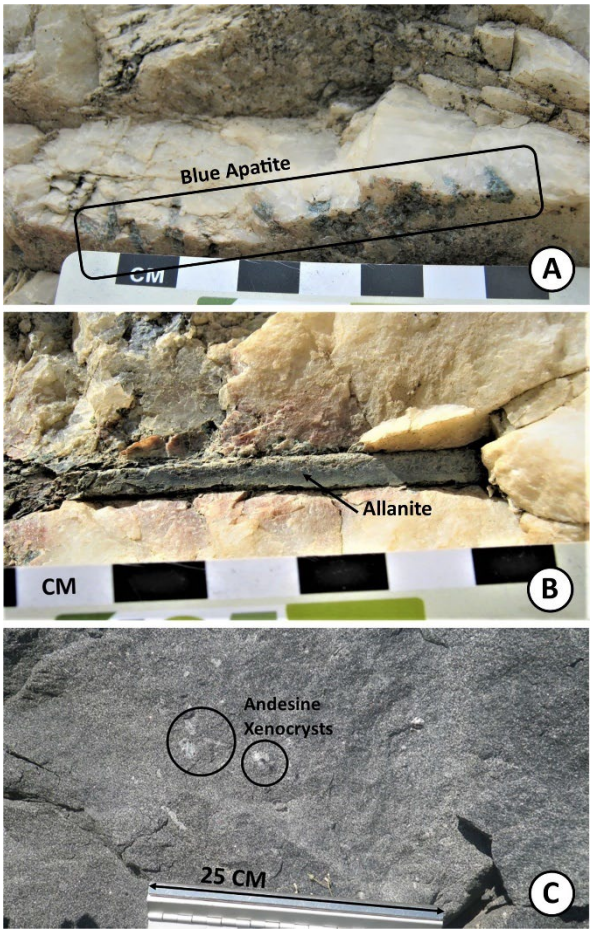


Photo 4. A) Blue apatite in "trains" along albite crystal boundaries in granitic pegmatite. **B)** Single allanite crystal in granitic pegmatite. **C)** Andesine xenocrysts in metamorphosed mafic dike adjacent to pegmatite. Photos from Péloquin et al. (2020).

Table 4. Summary of geochemical and mineral chemistry data from mafic rocks at Stop 6. Work was performed at the OGS Geoscience Laboratory. Co-ordinates in NAD83, Zone 17. Analysis 4 from Easton (2003), analysis 6 from Peck et al. (1995).

Analysis Number	1	2	3	4	5	6	7
Sample Number	**19RME-2001**	**19RME-2003**	**19RME-2005**	**99RME-0337**	**19RME-2003**	**90DCP-226**	**19RME-2004**
Easting (m)	511908	511949	511898	556111	511951	402140	512003
Northing (m)	5137095	5137093	5137080	5109966	5137095	5143082	5137095
Rock Name	gneissic gabbro, cgr	gneissic gabbro, east contact	gneissic gabbro, west contact	Sudbury dike	leucogabbroic gneiss	gabbro. East Bull Lake	migmatitic mafic gneiss
SiO_2 (wt %)	46.37	46.28	46.98	45.53	44.61	48.82	51.10
TiO_2	2.87	3.18	3.01	2.96	1.38	1.55	0.42
Al_2O_3	17.01	15.88	15.61	16.61	18.11	17.70	13.98
$Fe_2O_3^{total}$	15.51	16.41	16.08	16.92	13.26	15.13	8.61
MnO	0.193	0.208	0.201	0.200	0.125	0.20	0.160
MgO	5.56	5.23	5.35	5.95	7.63	2.54	10.29
CaO	7.26	7.45	7.63	7.76	11.03	8.74	12.38
Na_2O	3.19	3.39	3.22	3.49	2.36	2.76	1.44
K_2O	1.40	1.38	1.38	1.25	1.02	1.32	0.57
P_2O_5	0.639	0.710	0.666	0.59	0.038	0.11	0.034
CO_2	<0.023	0.038	0.042	<0.03	0.116	0.09	0.159
S	0.135	0.211	0.113	0.08	0.249	n/a	0.044
LOI	0.65	0.42	0.41	<0.05	0.57	1.50	1.00
Total	100.74	100.63	100.63	99.73	100.16	100.37	100.06
Ba (ppm)	769	805	780	700	204	399	64
Rb	30	31	27	1	31	59	42
Sr	389	373	390	363	360	263	96
V	206	257	240	233	439	261	212
Pb	5.9	7.3	5.6	6	3.5	nr	1.9
Th	3.7	4.8	4.0	2.5	<1.5	nr	<1.5
U	2.2	<1.6	<1.6	0.7	<1.6	nr	<1.6
Nb	15.1	16.1	15.4	16	2.2	nr	0.7
Y	36.4	41.4	38.6	39	18.8	17	10.7
Zr	229	259	241	243	42	78	29
Ni	90	67	80	93	12	20	164
Cr	83	97	90	45	35	nr	531
Cu	84	96	66	58	70	282	65
Au (ppb)	0.7	0.7	1.o	25	1.5	nr	10.2
Pd	0.4	0.6	0.6	<8	<0.14	nr	24.7
Pt	0.3	0.4	0.4	<5	0.1	nr	30.6
amphibole	n/a	H	H	n/a	MH	n/a	FH
pyroxene	n/a	none	none	n/a	none	n/a	diopside
biotite	n/a	10%	10%	n/a	trace	n/a	trace
plagioclase	n/a	An_{27-31}	An_{27-31}	n/a	An_{52-65}	n/a	An_{74-77}
oxide	n/a	ilmenite	ilmenite	n/a	ilmenite	n/a	none
sulphide	n/a	pyrite	pyrrhotite	n/a	pyrrhotite	n/a	pyrite

Notes: *Major element oxides are in weight %; trace element data are in parts per million, except for Au, Pd, Pt which are in parts per billion. Plagioclase composition of xenocrysts in sample 19RME-2005 is* An_{46-55}.
Abbreviations: *cgr, coarse-grained; FH, ferrohornblende; H, hastingsite; LOI = loss-on-ignition; MH, magnesiohastingsite; n/a = not applicable; nr, not reported. Amphibole classification of Leake et al. (1997).*

Table 4 — continued.

Analysis Number	8	9	10	11	12
Sample Number	**19RME-2011**	**19RME-2008**	**19RME-2014**	**19RME-2009**	**19RME-2010**
Easting (m)	511971	512277	512313	512290	512312
Northing (m)	5137100	5137074	5137069	5137073	5137069
Rock Name	gneissic granodiorite	garnet amphibolite	garnet amphibolite	ultramafic dike	paragneiss
SiO_2 (wt %)	73.31	41.91	41.72	46.39	90.32
TiO_2	0.27	0.99	1.00	0.56	0.03
Al_2O_3	13.98	19.86	18.98	10.52	5.78
$Fe_2O_3^{total}$	2.61	16.90	18.00	10.99	0.47
MnO	0.022	0.154	0.165	0.166	0.09
MgO	0.81	7.30	8.16	17.60	0.86
CaO	1.87	11.31	9.92	9.77	0.39
Na_2O	3.15	0.95	0.79	1.33	0.87
K_2O	3.70	0.52	0.59	0.51	0.82
P_2O_5	0.024	0.024	0.023	0.105	0.012
CO_2	0.035	0.312	0.133	0.239	<0.023
S	0.011	0.306	0.239	0.050	0.019
LOI	0.41	0.58	1.10	1.61	0.46
Ba (ppm)	1861	131	106	162	1683
Rb	126.6	14.6	13.6	8.1	13.7
Sr	267	551	454	217	96
V	15	774	797	230	<3
Pb	39.3	<1.7	<1.7	<1.7	<1.7
Th	34.7	0.28	0.19	1.11	0.63
U	4.1	0.16	0.07	0.38	0.34
Nb	7.4	<0.7	<0.7	1.6	<0.7
Y	5.2	3.9	4.3	11.7	4.8
Zr	136	11	8	45	104
Ni	6	40	31	593	5
Cr	25	58	37	1891	23
Cu	<9	167	263	30	21
Au (ppb)	0.6	2.3	2.6	1.1	<0.6
Pd	0.16	0.23	0.15	4.94	<0.14
Pt	0.19	0.28	0.22	4.36	<0.06
TREE	160.91	18.73	17.21	47.78	76.26

Notes: *Major element oxides are in weight %; trace element data are in parts per million, except for Au, Pd, Pt which are in parts per billion.*
Abbreviations: *LOI = loss-on-ignition; n/a = not applicable; nr, not reported; TREE = total rare earth elements.*

Table 5. Summary of geochemical data from felsic and mafic intrusive rocks at Stops 7 and 8. Work was performed at the OGS Geoscience Laboratory. Co-ordinates in NAD83, Zone 17.

Analysis Number	**1**	**2**	**3**	**4**	**5**
Sample Number	**21RME-021 Stop 7**	**21RME-022 Stop 7**	**21RME-024**	**21RME-020 Stop 7**	**21RME-023 Stop 8**
Easting (m)	514042	5129235	514123	514046	514151
Northing (m)	5129295	5129037	5129524	5129293	5129037
Rock Name	gneissic megacrystic granodiorite	gneissic megacrystic granodiorite	gneissic megacrystic granodiorite	fine-grained gneissic diorite	fine-grained gneissic gabbro
SiO_2 (wt %)					
TiO_2	>0.75	>0.75	0.71	>0.75	>0.75
Al_2O_3					
$Fe_2O_3^{total}$					
MnO	0.15	0.14	0.08	0.16	0.14
MgO					
CaO					
Na_2O					
K_2O					
P_2O_5					
CO_2	0.20	0.167	0.190	0.430	0.196
S	0.059	0.054	0.017	0.103	0.137
LOI	0.14	0.33	0.39	0.66	0.82
Ba (ppm)	2265	1996	<2700	666	324
Rb	60.0	56.8	95.5	44.4	22.5
Sr	370	384	363	434	327
V	28	46	33	218	185
Pb	17	15	18	12	8
Th	4.1	4.3	4.7	3.4	<1.9
U	<1.3	<1.3	<1.3	<1.3	<1.3
Nb	15.7	24.4	16.3	15.1	5.3
Y	32.0	31.7	20.3	47.2	31.7
Zr	743	541	544	46	85
Ni	5	6	3	37	85
Cr	12	11	<7	22	129
Cu	25	17	13	46	46
Au (ppb)					
Pd					
Pt					
TREE	>129	>129	>99	>120	>53

Notes: *Major element oxides are in weight %; trace element data are in parts per million, except for Au, Pd, Pt which are in parts per billion.*
Abbreviations: *LOI = loss-on-ignition; n/a = not applicable; nr, not reported; TREE = total rare earth elements.*



Photo 5. Non-migmatitic, leucogabbroic gneiss at Stop 6a. This rock is cut by the gneissic gabbro of the Sudbury dike swam, and thus is older than 1240 Ma. Scale card is 10 cm long.



Photo 6. Gneissic granodiorite to monzogranite at Stop 6a, possibly related to the Creighton granite. Scale card is 10 cm long.

- Approximately 16 m of non-migmatitic, gneissic leucogabbro (Photo 5). In thin section, the gneissic leucogabbro has a texture characterized by 120° grain boundaries and consists predominantly of brown amphibole and andesine. Brown amphibole is common under granulite facies conditions (Froese 1973). This is consistent with metamorphic conditions in this part of Dill Township identified by Kwak (1968). Geochemical data suggests the leucogabbro may be part of the East Bull Lake intrusive suite (emplaced *circa* 2475 Ma; compare analyses 5 and 6, Table 4).
- Approximately 10 m of migmatitic mafic gneiss.
- 4 m of migmatitic granodiorite gneiss (Photo 6), similar to the host rock to samples C96-2 and C96-1 in Corfu and Easton (2001). This granodiorite gneiss has elevated U and Th contents (analysis 8, Table 4) typical of early Paleoproterozoic felsic rocks in the Sudbury area such as the Creighton granite and the Copper Cliff rhyolite (both *circa* 2460 Ma).
- 10 m wide outcrop gap
- Approximately 35 m of migmatitic mafic gneiss, locally with deformed granitoid veins and layers present.

Stop 6B. Roadcut. From west to east the roadcut consists of:

- Approximately 125 m of interlayered grey to locally rusty, thin layered paragneiss, complexly folded, and weakly layered, variably migmatitic mafic gneiss and deformed granitoid layers.
- Approximately 45 m of migmatitic mafic gneiss, locally with deformed granitoid veins and layers present. This gneiss hosts a 30 m long, discordant zoned non-radiogenic pegmatite vein.
- Approximately 200 m of garnet amphibolite (25-50% garnet) (analysis 9, 10, Table 4) (Photo 7). At one point, it is cut by a near-vertical ultramafic dike (analysis 11, Table 4). A thin (1 m thick), near vertical quartzose gneiss band is likely of metasedimentary origin given its high silica content (analysis 12, Table 4).
- Approximately 100 m of interlayered grey to locally rusty, thin layered paragneiss and weakly layered, variably migmatitic mafic gneiss and deformed granitoid layers.

All of the units are steeply-dipping (70-85°) to the southeast. The age of the paragneiss and mafic gneiss is not well constrained, although Easton and Murphy (2002) and Easton and James (1997) observed that migmatitic mafic gneiss in the Street Township area to the northeast were Archean, whereas non-migmatitic mafic gneisses and amphibolite units were Proterozoic. If the migmatitic granodiorite gneiss in the west roadcut is indeed correlative with similar granitoid rocks in the Street Township and the Sudbury area that are *circa* 2460 Ma, this would suggest that the migmatitic mafic gneisses in the roadcut are Archean.

The garnet-rich mafic gneisses (25-50% garnet) present in the eastern third of the roadcut (Photo 6) are of potential economic interest. Garnet-rich mafic gneisses occur as discontinuous lenses in the Grenville Front tectonic zone from Sudbury to River Valley and have been mined locally in Street Township as a source of garnet (Easton and Murphy 2002). Kwak (1968) shows the presence of other outcrops of garnet-rich gneiss to the northeast of the roadcut, suggesting that they may be more abundant in Dill Township than previously suspected. Easton and Murphy (2002) and Easton (1996, 2003) suggested that these garnet-rich rocks may represent metamorphosed hydrothermal altered rocks similar to those present in volcanogenic massive sulphide (VMS) systems.

Return to vehicles and continue west on Highway 537.

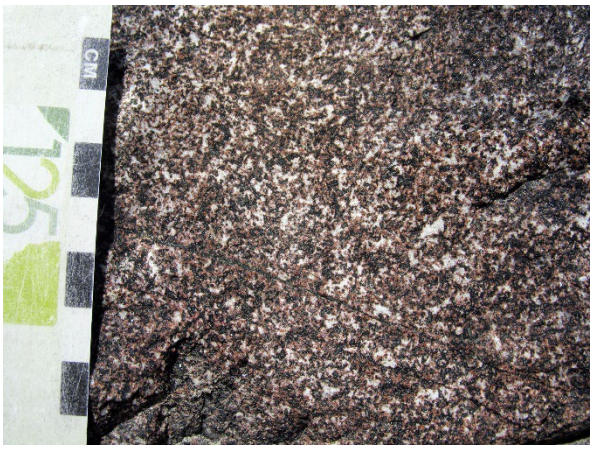


Photo 7. Close-up of garnetite that constitutes much of the roadcut at Stop 6b. Scale card is 1 cm long. Analyses 9 and 10, Table 4, are from this unit.

63.1 km — Optional Stop – Large roadcuts dominated by grey orthogneiss occur on both sides of the highway (UTM co-ordinates north side 511060E 5136470N, south side 511010E 5136460N), with an approximately 15 m wide Sudbury swarm dike similar to that observed at Stop 6A present in the north roadcut (UTM 511032E 5136448N).

63.3 km — Junction 537 and Estaire Road, turn south onto Estaire Road.

63.6 km — Cross from Grenville Front tectonic zone into Nepewassi domain.

72.3 km — Junction Estaire Road and Nelson Road, turn right onto Nelson Road.

73.7 km — Pull over and stop by roadcut on north side of road.

Stop 7 – Estaire pluton, granulite, Nepewassi domain

UTM co-ordinates514045E 5129295N

At this location, we are approximately 15 km south of the Grenville Front, and approximately 6.5 km south of the northern boundary of Nepewassi domain. This roadcut consists predominantly of green (Photo 8a) and pink (Photo 8b), garnet-bearing, potassium feldspar megacrystic granodioritic gneiss. The gneiss incorporates irregular pods of fine- to medium-grained gneissic diorite (Photo 8c), which could be a co-magmatic phase (analysis 4, Table 5). The green coloration is a typical effect of granulite-facies metamorphism.

In thin section, orthopyroxene occurs in the gneissic diorite, along with clinopyroxene- garnet [$Alm_{58}Adr_2Grs_{18}Prp_{19}Sps_4$]-hastingsite-biotite (4.5 wt. % TiO_2) and andesine. Opaque minerals are pyrrhotite and pyrite. Orthopyroxene is locally altered to calcite and is the only mafic phase that shows alteration. The presence of orthopyroxene indicates that this rock reached temperatures of at least 800°C (Pattison et al. 2003).

The overall mineralogy of the green and pink granodioritic gneiss is similar, but there are a few key differences. In the green rock, garnet [$Alm_{66}Adr_3Grs_{21}Prp_7Sps_4$] is abundant, and occurs primarily along the margins of potassium feldspar megacrysts and large plagioclase laths. Amphibole is hastingsite, plagioclase is oligoclase, potassium feldspar contains 1.3-1.9 wt. % BaO, and biotite contains (3-5 wt. % TiO_2). Both magnetite and ilmenite are present, with minor amounts of Al in magnetite and Mn in ilmenite; no sulphide minerals were observed. Fluorapatite and 50-200 micron-size zircon are also present. In contrast, in the pink rock, garnet [$Alm_{67}Adr_2Grs_{20}Prp_{10}Sps_2$] is sparser, and plagioclase phenocrysts have andesine cores and oligoclase rims. Opaque minerals are ilmenite (stoichiometric) and minor pyrrhotite and pyrite, and allanite are also present. Neither sulphide mineral nor allanite are in the green rock.



Photo 8. A) Fresh surface of garnet-bearing potassium megacrystic gneissic granodiorite at Stop 7 showing green coloration suggestive of granulite facies metamorphism. **B)** Fresh surface of garnet-bearing potassium megacrystic gneissic granodiorite at Stop 7 showing pink coloration. **C)** Fine-grained mafic gneissic diorite raft hosted in potassium megacrystic gneissic granodiorite. Scale card in all images is 10 cm long.

The presence of potassium megacrystic granodiorite and fine-grained dioritic rocks is typical of rocks of the Killarney intrusive suite (*circa* 1740 Ma), which are common in the eastern part of Nepewassi domain (cf. Easton 2014). Preliminary geochemical data for granodiorite samples of the Estaire pluton are characterized by high Ba (>2000 ppm) and high Zr (>500 ppm) contents (analyses 1-3, Table 5.) These geochemical characteristics resemble those of the West Bay pluton in Nepewassi domain, located south of Verner (Van de Kerckhove and Easton 2016). A sample of this unit, from the roadcut at the road junction 100 m from the stop was collected for geochronology, but results were not available at the time of the trip.

Return to vehicles and continue to road junction.

73.8 km — Junction, Nelson, McVittie, Secord roads, turn left onto McVittie Road.

74.1 km — Pull over and stop, roadcuts are present on both sides of the road. We are interested in the north end of the roadcut on the west side of the road.

Stop 8 – Estaire pluton, intrusion breccia, amphibolite, Nepewassi domain

UTM co-ordinates 514150E 5129080N.

From north to south, the roadcut exposes a spectacular intrusion breccia with rafts of fine-grained mafic material (gabbro to diorite) hosted by medium-grained granodiorite. The centre part of the outcrop is medium-grained gneissic granodiorite to monzogranite. The south end of the outcrop is a fine- to medium-grained gneissic gabbro to diorite (analysis 5, Table 5). As at the last stop, well-developed intrusion breccias and intercalation of mafic and felsic magmatic phases are typical of the Killarney intrusive suite.

74.4 km — Turn around and return to Nelson Road, right onto Nelson Road.

75.8 km — Nelson Road and Estaire Road, turn left and head north.

81.6 km — Pull over and stop, examine roadcut on east (right) side of road.

Stop 9 – Quartzite, Nepewassi domain

UTM co-ordinates 513333E 5134250N

This roadcut consists of medium layered quartzite, compositionally a quartz arenite, with thin, boudinaged mafic layers (dikes?) (Photo 9). The stop is near the southern end of a 2.2 km long, up to 300 m wide, belt of quartzite, part of which was quarried for silica flux between 1910 and 1924, first by the Canadian Copper Company and later by the International Nickel Company of Canada (MDI 41I07SW00002).

Quartzite units occur throughout the Nepewassi domain, most notably in the French River area but also as thin slivers in southeastern Nepewassi domain (Easton 2014; Van de Kerckhove 2016). A sample from this roadcut was collected for geochronology, but results were not available at the time of the trip.



Photo 9. Quartzite at stop 9. Note thin amphibolite layer above the end of the hammer handle (which is 40 cm long).

Krogh (1989) reported a predominantly Archean population (≥2650 Ma) with a metamorphic age of *circa* 1060 Ma. Quartzites in southeastern Nepewassi domain studied by Van de Kerckhove (2016) also had predominantly Archean populations (≥2650 Ma, but generally <2700 Ma). Van de Kerckhove (2016) suggested that the populations in the quartzites were suggestive that they may be correlative with the Lorrain or Bar River Formations of the Huronian Supergroup, however, there is nothing unique in the zircon

populations of any of the Nepewassi domain quartzites that is diagnostic that they are part of the Huronian Supergroup versus representing another stratigraphic package. Davidson (1997, p.29-32) has an extensive discussion regarding whether or not Huronian Supergroup rocks are present in the Grenville Front tectonic zone.

Biotite from near this Stop show evidence for excess argon, yielding ages between 950 to 1020 Ma (Fairbairn, Hurley and Pinson 1960; Hanson and Gast 1967), which are similar to zircon ages from the GFTZ in the Sudbury area (Krogh 1994; Corfu and Easton 2000).

Return to vehicles and continue north on Estaire Road.

82.5 km — Enter Wanup pluton

83.3 km — Pull over on right shoulder, outcrop opposite on west side of road.

Stop 10 – Wanup pluton, flattened megacrystic granodiorite, Nepewassi domain

UTM co-ordinates 512357E, 5135635N

This roadcut exposes rocks of another megacrystic granite pluton in Nepewassi domain, the Wanup pluton. We are at the north end of the pluton at this stop, and overall, the Wanup pluton exhibits a more intense gneissic fabric than does the Estaire pluton that we saw at Stops 7 and 8. There are 2 rock types present in the roadcut, a matrix-rich, gneissic megacrystic diorite to granodiorite (Photo 10a, 10b) and a matrix-poor, megacrystic granodiorite to monzogranite (Photo 10c). As was the case for the Estaire pluton, it is likely that the Wanup pluton may be a Killarney intrusive suite body.

Return to vehicles, continue north on Estaire Road.

84.7 km — Pull over on right shoulder, outcrop on east side.





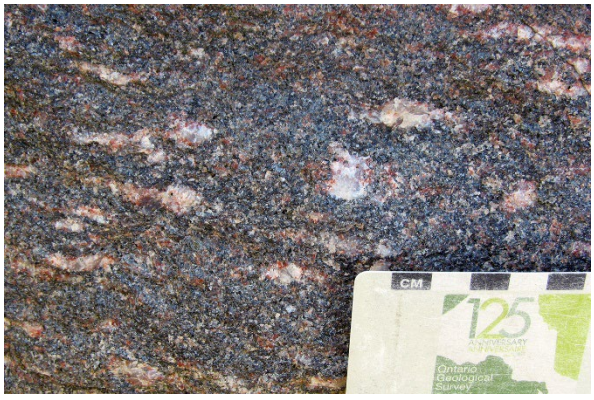


Photo 10. A) weathered surface of potassium feldspar megacrystic gneiss granodiorite of the Wanup pluton from an outcrop 200m to the southeast of Stop 10. **B)** fresh surface of matrix-poor potassium feldspar megacrystic gneiss granodiorite of the Wanup pluton from Stop 10. **C)** fresh surface of matrix-rich potassium feldspar megacrystic gneiss granodiorite of the Wanup pluton from Stop 10. Scale card in all images is 10 cm long.

Stop 11 – Nepewassi terrane boundary?

UTM co-ordinates 511070E 5136160N

This 100 m long, up to 3 m high, roadcut exposes a variety of tectonites related to the boundary between the Nepewassi domain to the south and the Grenville Front tectonic zone to the north.

The south end of the roadcut consists of irregularly layered migmatitic paragneiss with boudinaged mafic layers (Photo 11). Continuing northward, is a transition into a zone of flattened grey granodioritic gneisses and amphibolite (Photo 12). Present within this zone are extremely deformed, porphyroclastic granitic pegmatite dikes that are parallel to the near-vertical gneissic fabric present throughout the roadcut (Photo 13).

Near the north end of the roadcut, large potassium feldspar porphyroclasts are present locally in the tectonites (Photo 14). Isoclinal folds are also present in the tectonites (Photo 15).

Return to vehicles, continue north on Estaire Road.



Photo 11. Irregularly layered migmatitic paragneiss with boudinaged mafic layers at the south end of Stop 11. Hammer handle is 33cm long.



Photo 12. Flattened grey granodioritic gneiss (left) and flattened amphibolite (right) at Stop 11. Hammer handle is 33cm long.



Photo 13. Flattened, migmatitic amphibolite (left) and porphyroclastic granitic pegmatite dike (right) at Stop 11. Hammer handle is 33cm long.

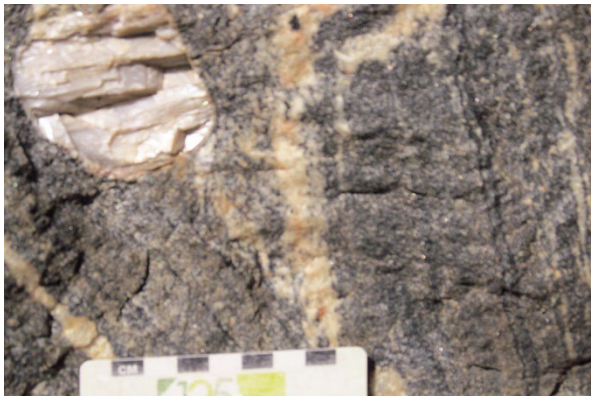


Photo 14. large potassium feldspar porphyroclast at Stop 11. Scale card is 10 cm long.



Photo 15. Folding of leucosome layer in grey, dioritic gneiss, north end of Stop 11. Scale card is 10 cm long.

85.0 km — Junction with 537, continue north on Estaire Road.

86.6 km — Pull over on right, outcrop on right (east) side.

Stop 12 – Migmatitic amphibolite

UTM co-ordinates 510020E 5137675N

This roadcut is representative of much of the fine- to medium-grained amphibolite units with 1% to 5% thin, stringy feldspathic leucosome present in the GFTZ west of the Wanapitei River. This rock was described as "common amphibolite" by Kwak (1969). The affinity of these amphibolite units has not been firmly established. Both Lumbers (1975) and Dressler (1984) considered them as possible equivalents of the Nipissing intrusive suite, however, the presence of leucosome suggests that they might be older. Geochemical data (unpublished) obtained by the author indicates that the amphibolite bodies are more primitive than typical Nipissing intrusive suite rocks.

Return to vehicles and continue north on Estaire Road.

87.7 km — Pull over on right shoulder (UTM 510020E, 5137675N), outcrop on right (east) side. **Optional** – mafic pod (migmatitic amphibolite) with straight gneiss on north side.

88.5 km — Pull over on right shoulder, outcrop on the right (east) side of Estaire Road. Stop 4.5 of Davidson, Carmichael and Pattison (1990), Stop 2-10 of Davidson (1995).

Stop 13 – Garnet amphibolite and schists

UTM co-ordinates 509410E 5139350N

This approximately 80 m long roadcut consists of several different phases. The southern end of the roadcut consists of weakly migmatitic, garnet amphibolite (analysis 8, Table 6), which is in sharp contact with the pelitic and semi-pelitic schists that constitute the bulk of the outcrop. There are at least 3 schist units in the roadcut, from south to north these are.

- Biotite-garnet schist with boudinaged quartz layers (quartz veins?)
- Muscovite-garnet schist with andesine porphyroblasts and intercalated with thin layers of semi-pelite to psammite (analysis 5, Table 6). Garnet [$Alm_{75}Adr_{0}Grs_{7}Prp_{14}Sps_{5}$].
- Muscovite-staurolite-garnet schist (analysis 6, Table 6) with andesine porphyroblasts. Staurolite is black on the weathered surface and is best seen on slabbed surfaces. Garnet [$Alm_{76}Adr_{1}Grs_{6}Prp_{9}Sps_{5}$].

The schist unit can be traced almost continuously over approximately 3.5 km and is near its widest extent at this stop. Only 900 m to the north, on Highway 69 (Photo 16), the unit is only about 15 m thick.

Davidson, Carmichael and Pattison (1990) report that at this stop, in thin section, kyanite forms single crystals in contact with all the other minerals (garnet-biotite-muscovite-plagioclase-quartz) and that it is commonly grown across large muscovite flakes. Sillimanite occurs as bundles of needles, generally associated with, or replacing, biotite, particularly at the edges of garnet

porphyroblasts. Kwak (1971) reported kyanite and potassium feldspar together as well. Figure 7 presents thermobarometric curves which indicate equilibration at ~8 kbar and 710°C. This is ~1 kbar and 40°C higher than the result obtained from a sample collected only 1.2 km to the north at the golf course, suggesting a thermal gradient of ~50°C/km (Davidson et al. 1990).



Photo 16. Muscovite-potassium feldspar schist on the east side of Highway 69, 900 m north of Stop 13 (analysis 7, Table 6). Schist is bounded to the south by garnet amphibolite (not in the photo) and to the north by irregularly layered grey, migmatitic, tectonite of intermediate composition (lower left of the photo).

It is tempting to think that these schists might be metamorphosed equivalents of the Huronian Supergroup, mostly likely the McKim Formation which has the bulk-rock major element composition capable of forming the observed mineral assemblages, most notable a high-aluminum content (compare analyses 1-4, Table 6). Although the major element geochemistry supports this possibility, trace element data (analyses 5-7, Table 6) are less conclusive, as none of the samples analyzed have Ni/Co or Cr/Zn ratios typical of post-Archean fine-grained sedimentary rocks, such as the McKim Formation (between 1 and 2.5 and 1 and 1.6, respectively, Tang, Chen and Rudnick 2016). In fact, the schist sample from the site shown in Photo 16 (analysis 7, Table 6), is similar to that of the Nepewassi domain quartzite sample from Stop 9 rather than the McKim Formation.

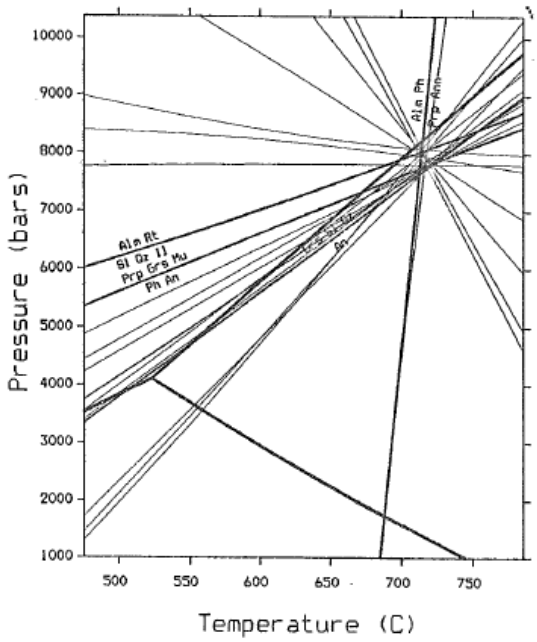


Figure 7. TWEEQ plot showing 22 thermobarometric curves (5 independent) for pelitic schist with the main-stage assemblage garnet-kyanite-sillimanite-muscovite-biotite-plagioclase-quartz-ilmenite-rutile from a sample collected on the hill above Stop 13, as reported by Davidson, Carmichael and Pattison (1990).

Rare earth element (REE) patterns for these rocks are inconclusive. The schist samples from this Stop have patterns (Figure 8) consistent with the post-Archean Australian shale composite (Taylor and McLennan 1985), but also with the pattern found in FIII rhyolites (Lesher et al. 1978) associated with volcanogenic massive sulphide (VMS) systems.

If these schists are part of the McKim Formation, this correlation is only possible if the McKim Formation has undergone extreme tectonic thinning (from approximately 1,000 m thick immediately north of the Grenville Front to approximately 100 m or less here. In addition, the adjacent Huronian Supergroup units, such as the Mississagi Formation, which are sandstone-dominated, are nowhere to be seen. It could be that some of the neighbouring mafic host rocks, such as the garnet amphibolite at this stop, be metamorphosed Huronian Supergroup volcanic

rocks. Alternatively, as discussed at Stop 6b, the schists could represent metamorphosed hydrothermally altered rocks, similar to those present in VMS systems.

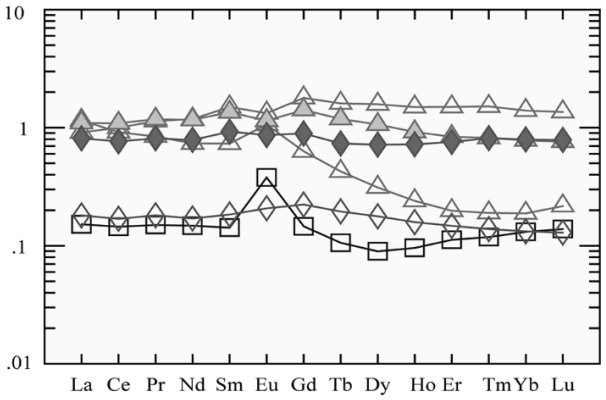


Figure 8. Rare earth elements for selected samples normalized to the post-Archean Australian shale composite (Taylor and McLennan 1985). Schist samples from Stop 13 (open and filled triangles, analyses 5, 6, Table 6) and a McKim Formation sample from the Southern Province (filled diamonds) are straight lines centred around 1, suggesting that they could be metasedimentary rocks of the McKim Formation. Similarly, a sample of quartzite from Stop 9 (open diamonds, analysis 4, Table 6) parallels the composite, but at 1/10 the REE content, likely due to the abundance of quartz. Another schist sample (photo 16, open squares, analysis 7, Table 6) and the felsic rock from Stop 6B (open triangles, analysis 8, Table 4) have unusual concave heavy rare earth patterns, suggesting that they may have been affected by alteration.

Return to vehicles and continue north on Estaire Road.

88.5 km — Junction Estaire Road, Gladu Road, and Bentley Avenue. Park vehicles and walk to older roadcuts on the south side of Gladu Road.

Stop 14 (Optional). Mylonitic rocks near the Grenville Front

UTM co-ordinates 508520E 5141335N

We are approximately 800 m south of the Grenville Front at Stop 14. The two old roadcuts south of Gladu Road consist of thin-layered, compositionally varied, straight gneisses, with some near vertical, gneissic fabric parallel, porphyroclastic granite pegmatite dikes. As we saw at Stop 5, protolith of these gneisses is not easily determined.

Return to vehicles and continue north on Estaire Road.

91.5 km — **Optional.** follow Bentley Avenue to turnaround at end for a view across the Grenville Front.

Retrace route to Estaire Road, take Highway 69 back into town to Science North.

End of road log.

Table 6. Summary of geochemical from rocks at Stop 13 as well as average samples of the McKim Formation. Analysis 1 from Card, Innes and Debicki (1977, Table 8, p.40); analyses 2, 3, from Kwak (1968, 1971), with samples 3 from near Stop 13; analyses 4-8 from Easton (unpublished data, 2022).

Analysis Number	**1**	**2**	**3**	**4**	**5**	**6**	**7**	**8**
Sample Number	**Average of 8 Samples, McKim Fm**	**Average of 14 samples**	**232C**	**21RME-102 Stop 9**	**21RME-029 Stop 13**	**21RME-030 Stop 13**	**21RME-063**	**21RME-027 Stop 13**
Easting (m)				513333	509443	509401	509866	509443
Northing (m)				5134250	5139333	5139388	5140422	5139333
Rock Name	semi-pelite	schist	schist	quartzite	schist	schist	schist	garnet amphibolite
SiO_2 (wt %)	57.90	59.20	60.95					
TiO_2	0.83	1.11	1.23	0.0	>0.75	>0.75	0.12	>0.75
Al_2O_3	21.94	20.75	19.90					
$Fe_2O_3^{total}$	8.10	8.54	5.48					
MnO	0.06	0.08	0.05	0.00	0.09	0.05	0.00	0.21
MgO	2.93	2.90	2.60					
CaO	1.01	1.34	1.49					
Na_2O	1.29	1.49	2.27					
K_2O	2.77	3.25	3.74					
P_2O_5	0.12	0.15	0.10					
CO_2	0.11	nr	nr	0.074	0.09	0.10	0.11	0.06
S	nr	nr	nr	0.035	0.174	0.105	<0.03	0.131
LOI/H_2O^+	3.20	1.56	1.94	0.90	2.98	2.93	1.93	0.79
Ba (ppm)				95	1802	1968	1288	123
Rb				47.7	197.0	196.8	18.8	17.4
Sr				39	57.7	68.7	50.0	116
Ga				5.6	31	30	14	16
V				11	283	261	24	268
Pb				3	16	15	5	9
Th				<1.9	13.8	11.2	<1.9	<1.9
U				<1.3	4.4	3.2	<1.3	<1.3
Nb				0.8	15.8	13.8	1.9	3.5
Y				4.8	45.9	34.5	3.0	21.9
Zr				82	258	200	54	60
Ni				9	92	81	9	127
Cr				37	184	150	17	221
Cu				3	257	166	2	106
Ni/Co				2.97*	0.42	0.49	4.55*	n/a
Cr/Zn				5.97*	0.26	0.22	1.21	n/a
Au (ppb)								
Pd								
Pt								
TREE				32	202	206	27	<27

Notes: *Major element oxides are in weight %; trace element data are in parts per million, except for Au, Pd, Pt which are in parts per billion.* ******* *indicates Ni/Co or Cr/Zn with Archean ratios.*
Abbreviations: *LOI = loss-on-ignition; n/a = not applicable; nr, not reported; TREE, total rare earth element.*

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Field Trip 3 – Magmatism and Brecciation in the Footwall Rocks of the Southwestern Sudbury Structure

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Introduction

This one-day field trip presents geological highlights from the Ontario Geological Survey (OGS) Southwest Sudbury Structure bedrock mapping project. This project is part of a collaborative program with the OGS, the Mineral Exploration Research Centre at the Harquail School of Earth Sciences, Laurentian University, and the private sector.

Regional Geology

The Sudbury Igneous Complex (SIC) is interpreted to represent a melt sheet produced by the impact of a meteorite at 1850 Ma (Dietz 1964; Krogh et al. 1984; Davis 2008). The SIC is part of the Southern Province of the Canadian Shield and is located north of the Grenville Front astride the southern contact of the Archean Superior Province (Figure 1). The term Sudbury Structure, which is used throughout this field guide, refers to the SIC, the Sudbury Basin that contains rocks of the Whitewater Group, and the outer zone of brecciated footwall rocks. The Sudbury Structure is geographically subdivided into north, south and east ranges (Figure 2).

In the Sudbury area, Archean rocks of the Superior Province are part of the Abitibi Subprovince and consist of the Levack Gneiss Complex (2711 to 2647 Ma: Krogh et al. 1984; Wodicka and Card 1995), the Benny greenstone belt (2680-2700 Ma: Ontario Geological Survey 2019) and the Ramsey–Algoma granitoid complex (Card 1979), which includes the Cartier (2642 Ma: Meldrum et al. 1997) and Birch Lake (2651 Ma: Kamo 2006; Easton and Heaman 2008; Gordon, et al. 2018a) batholiths (Figure 2). The southern boundary of the Superior Province is overlain unconformably by the supracrustal rocks of the Huronian Supergroup. The Huronian Supergroup was deposited in a continental rift and on a continental platform between 2450 and 2219 million-years ago (Krogh et al. 1984; Bennett et al. 1991) and has been interpreted to represent a partial Wilson cycle (Young 1983; Hoffman 1989; Bennett al. 1991; Young et al. 2001).

The Sudbury area has been intruded by numerous dikes, sills and plutons of various ages (Figure 2). Known intrusions and intrusive suites include, in chronological order, the 1) Joe Lake gabbro (2660 Ma: Bleeker et al. 2015); 2) Matachewan dike swarm (2480-2460 Ma: Heaman 1997; Bleeker et al. 2012); 3) Drury Township, Falconbridge and Frood intrusions of the East Bull Lake Intrusive Suite (2480 Ma: Krogh et al. 1984; James et al. 2002; Keays and Lightfoot 2020); 4) Creighton and Murray plutons (2460 Ma: Bleeker et al. 2015); 5) Nipissing Intrusive Suite (2219 to 2210 Ma: Davey et al. 2019; Noble and Lightfoot 1992; Corfu and Andrews 1986: Bleeker et al. 2015); 6) Trap dike swarm (1750 Ma: Bleeker et al. 2015); 7) Sudbury dike swarm (1238 Ma: Krogh et al. 1987), and; 8) Grenville dike swarm (590 Ma: Kamo et al. 1995).

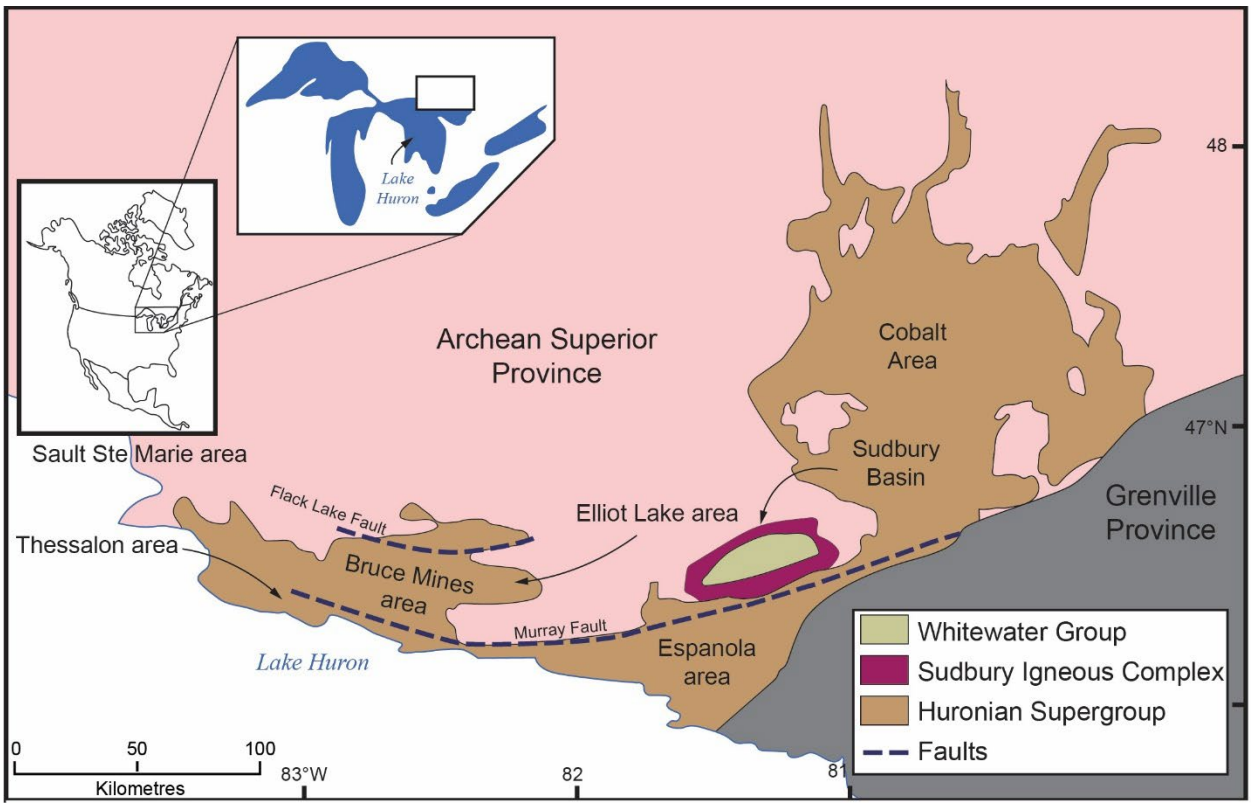


Figure 1. Sketch map showing the regional setting of the Sudbury Igneous Complex and the Huronian Supergroup (*modified from* Young et al. 2001).

The Sudbury area has been affected by several episodes of deformation and metamorphism. Regional metamorphism is thought to have reached mid-greenschist to lower-amphibolite facies and generally increases to the south (Card 1978; Card et al. 1984; Fox 1971). Country rocks adjacent to the SIC were thermally metamorphosed but have since been overprinted by regional metamorphism (Dressler et al. 1991; Jørgensen et al. 2019; Généreux et al. 2021). Ductile deformation of the Southern Province has historical been interpreted to have started prior to or concurrent with emplacement of the Nipissing Intrusive Suite during the Blezardian Orogeny (2415-2219 Ma; Raharimahefa et al. 2014; Stockwell 1982), and continued during the Penokean Orogeny (1890-1830 Ma; Dressler 1984a; Bennett et al. 1991), the Yavapai and Mazatzal orogenies (1770-1600 Ma; Bailey et al. 2004; Raharimahefa et al. 2014; Papapavlou et al. 2017), and the Grenville Orogeny (1120-980 Ma; Carr et al. 2000). Recent studies have attributed most of the deformation in the Sudbury area to the Yavapai–Mazatzal orogenies (Bailey et al. 2004; Raharimahefa et al. 2014; Papapavlou et al. 2017). The lower age limit of ductile deformation is constrained by the age of the undeformed Sudbury dike swarm (1238±4 Ma; Krogh et al. 1987).

Three types of ore environments are associated with the SIC: 1) *contact deposits*, which are hosted in depressions at the base of the SIC; 2) *offset deposits*, which occur in quartz diorite dikes that extend for several kilometres from the SIC into the footwall rocks, and; 3) *footwall deposits*, which are found in the brecciated footwall rocks directly underlying the SIC (cf. Lightfoot 2017). It is generally accepted that nickel-copper-platinum group element (Ni-Cu-PGE) deposits in Sudbury are primarily magmatic and formed by differentiation of a sulphide melt during crystallization of the SIC (Keays and Crocket 1970

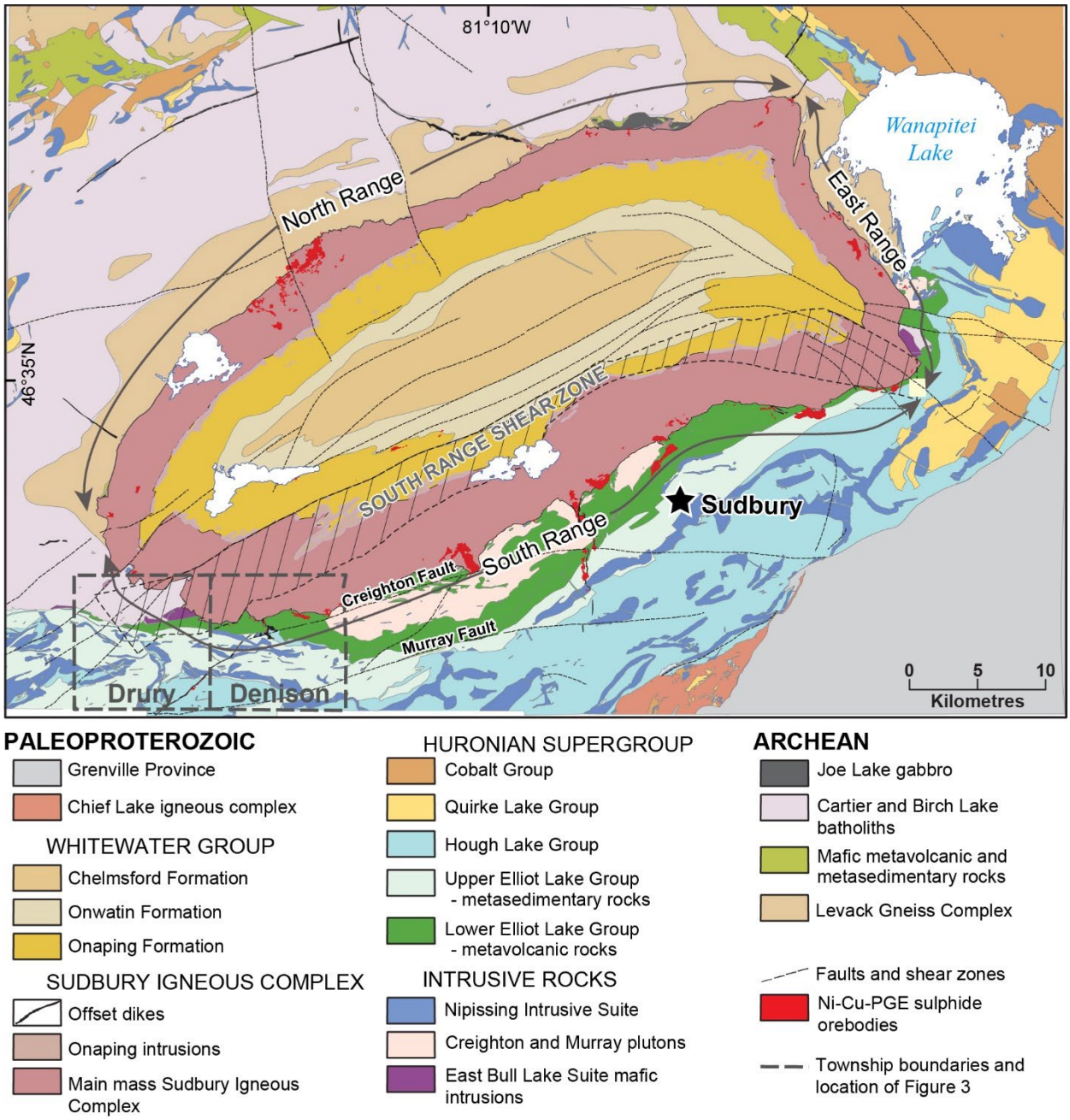


Figure 2: Geological map of the Sudbury area (*modified from* Ames and Farrow 2007). The location of Drury and Denison townships are shown in the lower left of the figure.

Naldrett et al. 1994; Naldrett 1999; Ames and Farrow 2007), with some involvement of hydrothermal fluids (Farrow et al. 1994; Jago et al. 1994; Morrison et al. 1994; Molnár and Watkinson 2001; Péntek et al. 2008). Many deposits in the South Range of the SIC were modified during post 1850-Ma tectonic events (Lightfoot 2017).

Mining and Exploration History

The southwestern portion of the Sudbury Structure has been an important mining and exploration area since the discovery of Ni-Cu-PGE mineralization related to the SIC. The first deposit discovered in the area was the Worthington deposit in 1884 (cf. Lightfoot 2017). Since then, the southwestern Sudbury Structure has supported numerous past-producing mines and currently hosts one active mine (Totten Mine), advanced prospects (Victoria Project) and undeveloped mineral occurrences (Figure 3; Table 1). The majority of mineral deposits are associated with SIC-related rocks and occur as contact deposits or as offset deposits within the Worthington and Vermilion offset dikes. Commodities include Ni-Cu-PGE, gold (Au), cobalt (Co), silver (Ag), iron (Fe), tin (Sn), selenium (Se), tellurium (Te) and arsenic (As) (Table 1). Several contact and offset deposits were modified by post-SIC deformation and ore is now hosted within shear zones (*i.e.* Vermilion and Chicago mines).

In addition to SIC-related mineralization, known prospects, occurrences and discretionary occurrences include: 1) Ni-Cu±PGE mineralization hosted by the Nipissing gabbro, volcanic and sedimentary rocks of the Huronian Supergroup, and within shear zones; 2) Au-Cu bearing quartz veins; and, 3) uranium-thorium (U-Th) mineralization in pyritic quartz-pebble-rich conglomeratic arenites in the lower Matinenda Formation of the Huronian Supergroup (Figure 3). Quartz veins and quartzites in the area have also been quarried for silica.

Table 1. Mineral deposits in Drury and Denison townships. Table includes names for mines and prospects. Occurrences and discretionary occurrences are not included here. Reference numbers in table correspond with mineral occurrence symbol on Figure 3.

Ref No.	Name	Commodity* (Primary/Secondary)
1	Totten Mine	Ni, Cu / PGE, Co, Au, Ag
2	Totten #1	Ni, Cu / PGE, Co, Au, Ag
3	Worthington Mine	Ni, Cu, PGE / Co, Au, Ag
4	Worthington #2	Ni, Cu, PGE / Co, Au, Ag
5	Howland Pit	Ni, Cu / PGE, Co
6	Robinson Mine	Ni, Cu / PGE, Co
7	Aer Mine (Rosen and Gersdorffite mines)	Ni, Cu / PGE, Co
8	Victoria Mine	Ni, Cu / PGE, Au
9	Vermilion Mine	Ni, Cu / PGE, Au, Ag, Co, Fe, Sn, Se, Te, As
10	Crean Hill Mine	Ni, Cu, PGE / Au, Co, Fe, Ag, Se, Te
11	Lockerby Mine	Ni, Cu, Co / PGE, Au, Ag
12	Ellen Pit	Ni, Cu / Au, Ag, PGE, Fe, Co, Se, Te
13	Chicago Mine	Ni, Cu / PGE, Au, Ag, Co, Fe, Se, Te
14	Sultana Nickel Mine	Ni, Cu
15	Delta Occurrence	Ni, Cu / PGE, Au
16	McIntyre Mine	Ni, Cu
17	Victoria Project	Ni, Cu, PGE
18	Alanaen and Maki West (Kerr Addison Prospect)	U, Th, Cu

*Primary and secondary commodities as listed in the Ontario Mineral Inventory Database (Ontario Geological Survey 2022).

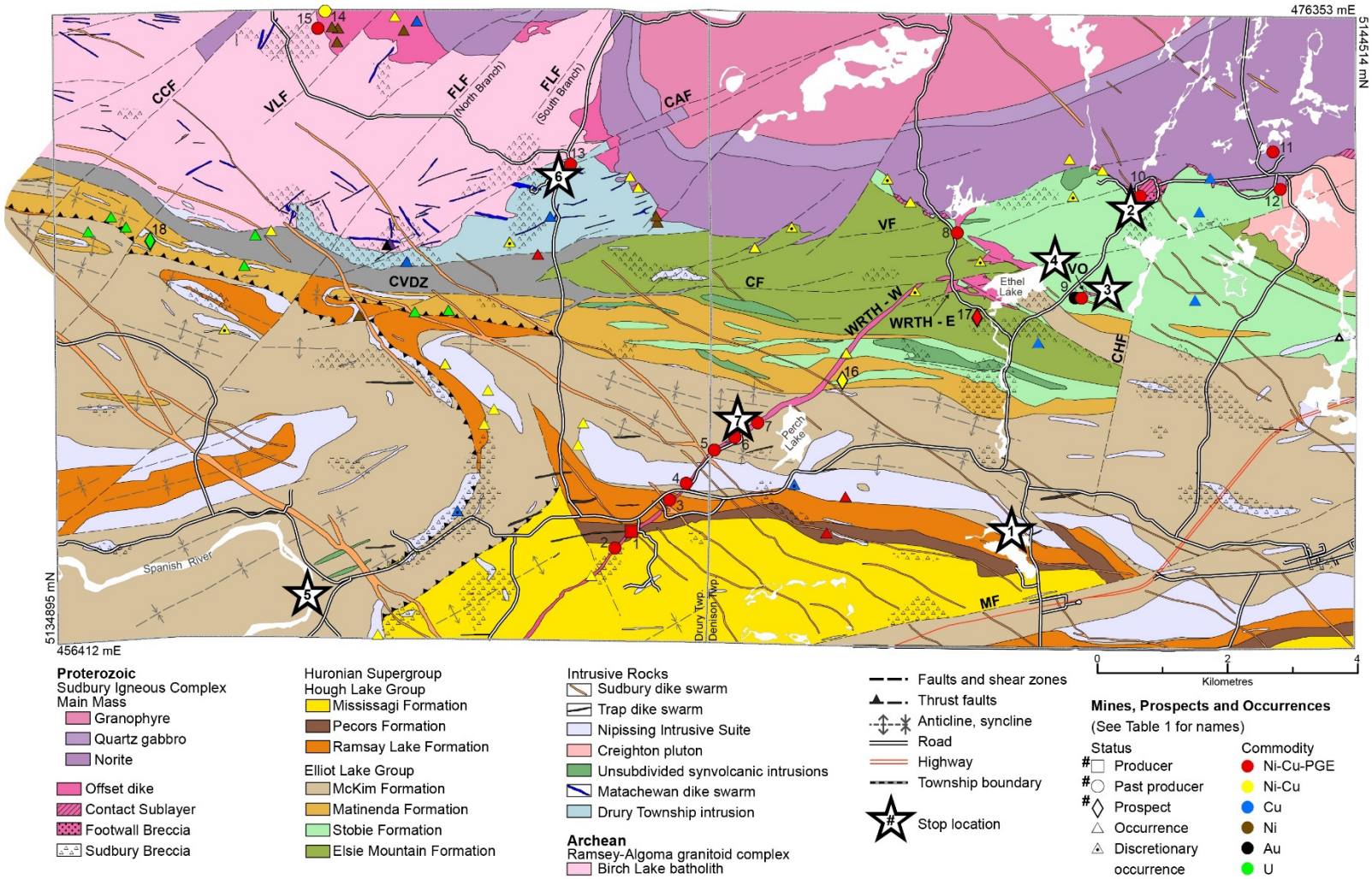


Figure 3. Geological map of Drury and Denison townships (*after* Gordon et al. 2018a; Gordon and Généreux 2017; Gordon 2018). Mineral occurrences are from Ontario Geological Survey (2022) and Gordon et al. (2018a). Universal Transverse Mercator coordinates are in North American Datum 1983, Zone 17. CCF = Cameron Creek Fault, VLF = Vermilion Lake Fault, FLF = Flack Lake Fault, CAF = Chicago Fault, CF = Creighton Fault, VF = Victoria Fault, CHF = Crean Hill Fault, MF = Murray Fault, WRTH-W = Worthington Offset dike, western limb, WRTH–E = Worthington Offset dike, eastern limb, VO = Vermilion Offset dike, CVDZ = Creighton-Victoria deformation zone.

Geological Overview of the Southwestern Sudbury Structure

Geological mapping in the southwest Sudbury Structure was completed between 2015 and 2018 and focussed on Drury and Denison townships (Figure 3) located approximately 50 km west of the City of Greater Sudbury (Figure 2).

For sake of clarity, the authors have omitted the prefix "meta" for the rock names in this field guide (*e.g.* gabbro *versus* metagabbro).

Sudbury Igneous Complex (SIC)

The SIC is elliptical in shape and approximately 30 km x 60 km in size (Figure 2). It is made-up of three main components: 1) Main Mass, which is a differentiated igneous body; 2) Contact Sublayer, a basal magmatic breccia; and, 3) Offset dikes, quartz diorite dikes emplaced in the footwall of the SIC (Giblin 1984; Dressler et al. 1991; Ames et al. 1997, 1998, 2002). All three components are exposed in the southwest Sudbury Structure.

Main Mass

The SIC, as exposed in the southwestern Sudbury Structure, exhibits the complete Main Mass stratigraphic sequence (Figure 3). From top to bottom, it consists of:

1. Granophyre: leucocratic monzogranite and upper plagioclase-rich phase of granodiorite.
2. Transition zone quartz gabbro: melanocratic to mesocratic quartz gabbro and monzogabbro with cumulus magnetite and apatite.
3. South Range norite: leucocratic to mesocratic quartz monzogabbro and norite.

Contact Sublayer

The main mass of the SIC overlies an extensive zone of magmatic breccia known as the Contact Sublayer. The Sublayer is heterolithic in matrix composition and clast type, variably gossanous and laterally discontinuous. It is generally classified into two groups based on the composition of its matrix: Sublayer Norite consists of breccia with a noritic matrix, whereas Sublayer Granite Breccia contains a granitic matrix (cf. Lightfoot 2017).

In the southwestern Sudbury Structure, the magmatic breccia along the base of the SIC (Figure 3) has been subdivided into 4 types based on matrix composition and clast abundance (cf. Gordon et al. 2018a).

1. Clast-rich Sublayer Norite (>30% clasts): heterolithic breccia with abundant gabbroic and lesser amounts of granitoid clasts in a noritic to leuconoritic matrix.
2. Clast-poor Sublayer Norite (<30% clasts): heterolithic breccia with a noritic matrix.
3. Sublayer Granite Breccia: Heterolithic breccia (>35% clasts) with abundant granitoid clasts and few mafic clasts in a pink-weathering matrix.
4. Heterolithic breccia with gabbroic and anorthositic gabbro clasts in a variably textured, leucogabbroic matrix.

Clast-rich and clast-poor Sublayer Norite occurs at the SIC–Archean granitoid and SIC–Huronian Supergroup contacts. The Sublayer Granite Breccia unit, which likely represents a variation of the classic Sublayer Granite Breccia, occurs exclusively at the SIC–Archean granitoid contact. Breccia type 4 occurs exclusively at the SIC–Drury Township intrusion contact and, although it has been tentatively classified as sublayer, it may represent Footwall Breccia.

Offset Dikes

Offset dikes of the SIC, also known as Offset Sublayer, are radial, concentric, and discontinuous segmented bodies that were emplaced into the footwall of the SIC (cf. Lightfoot 2017). Offset dikes consist of two phases: quartz diorite (QD) and a sulphide-enriched inclusion-bearing quartz diorite (IQD) (Grant and Bite 1984).

The Worthington Offset is a branching radial offset dike that trends southwest from the SIC contact (Figure 3). The proximal part, occupying a possible embayment in the Victoria Mine area north of Ethel Lake, occurs as numerous faulted segments (Grant and Bite 1984). Southward from the SIC, the offset dike narrows and then broadens before bifurcating into eastern and western limbs.

The eastern limb tapers gradually over a distance of 1500 m at the surface, trends southeast, and broadens at depth (Grant and Bite 1984). The western limb extends to the southwest for at least 15 km, with a thickness between 30 and 100 m. The dike truncates volcanic and sedimentary rocks of the Huronian Supergroup as well as Nipissing gabbro and is crosscut by the Creighton Fault (Figure 3). The Worthington Offset contains both the inclusion-poor QD and mineralized IQD phases. The contact between phases is sharp and inclusions of QD are present within IQD.

The Vermilion Offset dike crops out east of Ethel Lake, south of Crean Hill (Figure 3). The Vermilion Offset occurs within a 200 m long, northwest-striking zone of discontinuous QD and IQD pods at the contact of Sudbury Breccia and sedimentary and volcanic rocks of the Stobie Formation (Grant and Bite 1984; Szentpeteri et al. 2003).

Breccias in the Footwall Rocks

Two types of breccia occur within the footwall of the SIC: Footwall Breccia and Sudbury Breccia.

Footwall Breccia

Footwall Breccia is a parautochthonous breccia that occurs in discontinuous lenses and sheets between the Contact Sublayer and underlying footwall rocks. It consists mainly of brecciated and partially melted footwall rocks, and has an igneous to granoblastic matrix (Lakomy 1990; McCormick et al. 2002). Within the southwestern Sudbury Structure, Footwall Breccia has been identified at the Crean Hill Mine area (Figure 3) (Généreux et al. 2021), where it consists of breccia dikes and pods of various compositions hosted in volcanic rocks of the Stobie Formation.

Sudbury Breccia

All rock types in the Sudbury area that are older than *circa* 1850 Ma contain various amounts of impact breccia, locally called Sudbury Breccia. It is a parautochthonous breccia with an aphanitic to microcrystalline matrix, and occurs as discontinuous veins or tabular bodies within the footwall rocks. Sudbury Breccia is thought to have formed by cataclasis (Lafrance and Kamber 2010; O'Callaghan et al. 2016; Rousell et al. 2003) and/or frictional melting of the target rocks during cratering (Dressler 1984a; Lafrance and Kamber 2010; O'Callaghan et al. 2016; Rousell et al. 2003; Thompson and Spray 1994).

Within the southwestern Sudbury Structure, Sudbury Breccia occurs as fine-grained, dark green to greenish-yellow veins, dikes and irregular shaped bodies of various thicknesses and orientations. Thicker breccia veins, typically several metres in width, are usually clast-rich and form corridors along major lithologic contacts and structures. Two major heterolithic Sudbury Breccia belts were identified in Drury Township: 1) at the contact between the Drury Township intrusion and the Archean granitoid, and 2) following the folded contacts between Huronian sedimentary rocks and Nipissing gabbro in south-central Drury Township (Figure 3). These breccia belts range from a few metres to several hundred metres wide. In the belts, breccia matrix typically makes-up more than 40% of the outcrops.

Footwall rocks of the SIC

Archean Ramsey-Algoma granitoid complex

Monzogranite, granite and granodiorite belonging to the Archean Ramsey–Algoma granitoid complex are the oldest rocks in the southwestern Sudbury Structure (Figure 3). The contact between the Archean basement and rocks of the Huronian Supergroup is known to be unconformable (Stockwell 1964; Card 1990) but is highly sheared in the Sudbury area. The transition between the Cartier and Birch Lake batholiths of the Ramsey-Algoma granitoid complex is reportedly within Drury Township, but the exact location of the contact has not been determined (Tolman 1929; Meldrum et al. 1997). U/Pb zircon geochronology on a granitoid sample collected in western Drury Township yielded a minimum age of 2645±1 Ma (Gordon et al. 2018a), which suggests that the granitoid rocks in Drury Township are older than the Cartier batholith and more similar in age to the Birch Lake batholith.

Paleoproterozoic Huronian Supergroup

The Huronian Supergroup extends from the east shore of Lake Superior across the north shore of Lake Huron, and northeastward across the Cobalt Embayment to the Noranda area in northwestern Quebec (Bennett et al. 1991) (Figure 1). It consists of a southward thickening, up to 12 km thick, package of sedimentary and volcanic rocks that are subdivided, from oldest to youngest, into the Elliot Lake, Hough Lake, Quirke Lake and Cobalt groups (Bennett et al. 1991; Robertson et al. 1969). The minimum age of the Huronian Supergroup is constrained by the age of the intruded Nipissing Intrusive Suite (2210-2219 Ma: Davey et al. 2019; Corfu and Andrews 1986; Noble and Lightfoot 1992; Bleeker et al. 2015). Its maximum depositional age is constrained by the age of the felsic volcanic rocks of the Copper Cliff Formation (2452-2460 Ma: Krogh et al. 1984; Ketchum et al. 2013; Bleeker et al. 2015).

The oldest and lowermost Elliot Lake Group consists of an intercalated sequence of sandstone, conglomerate, siltstone, mudstone, and local volcanic rocks (Card 1978). With the exception of the carbonate-bearing Serpent Formation (Quirke Lake Group), the overlying Hough Lake, Quirke Lake and Cobalt groups each contain cyclical repeating sequences of lower conglomeratic units, middle siltstone and mudstone units, and upper sandstone units (Roscoe 1969). Rocks of the lower Huronian Supergroup, specifically the Elliot Lake and Hough Lake groups, are represented in the southwestern Sudbury Structure.

In the Sudbury area, most of the Huronian Supergroup strata are subvertical and are approximately west-northwest- to east-trending. Reversals of facing direction define the synclines, anticlines and thrust faults in the area. Despite folding and faulting within formations, the overall younging direction of the stratigraphy is southward. Unit thicknesses stated herein are apparent thicknesses.

Elliot Lake Group

The Elliot Lake Group is the lowermost unit in the Huronian Supergroup and the only unit to contain volcanic rocks. The basal volcanic rocks are interpreted as fissure eruptions related to deep-penetrating crustal faults during rifting (Card 1978). In the Sudbury area, the volcanic rocks are up to 3 km thick and are subdivided into the Elsie Mountain, Stobie and Copper Cliff Formations (Card 1978) (Figure 4). The felsic volcanic rocks of the Copper Cliff Formation are not exposed in Drury and Denison townships and will not be discussed further in this guide.

The Huronian Supergroup volcanic rocks in Drury and Denison townships exhibit significant lateral variation from east to west and will be discussed as 3 separate segments: western, central and eastern. In the western segment, the majority of rocks previously identified as volcanic have been reclassified as mylonites of the Creighton-Victoria deformation zone (Figure 3) (Gordon et al. 2015; Simard et al. 2016; Généreux et al. 2016; Généreux et al. 2017; Gordon et al. 2018a). In the central segment, which is east of the Creighton-Victoria deformation zone, volcanic rocks of the Elsie Mountain Formation are dominated by massive and pillowed basaltic flows that are locally amygdaloidal and porphyritic, with minor amounts of intercalated arenite and siltstone. South of the Elsie Mountain Formation, bimodal volcanic rocks of the Stobie Formation are intercalated with arenite of the Matinenda Formation (Figure 3). In this area, the Stobie Formation consists predominantly of massive and pillowed basaltic to rhyolitic flows with interbedded arenite, siltstone, and minor amounts of pyroclastic rocks. The central and eastern segments are separated by an unnamed northwest-trending fault in Denison Township (Figure 3). The eastern segment is dominated by the bimodal Stobie Formation, which is bound to the north and east by the SIC and Creighton pluton, respectively. The Elsie Mountain Formation is largely absent in the eastern segment, except for a thin sliver of basalt exposed adjacent to the Creighton pluton.

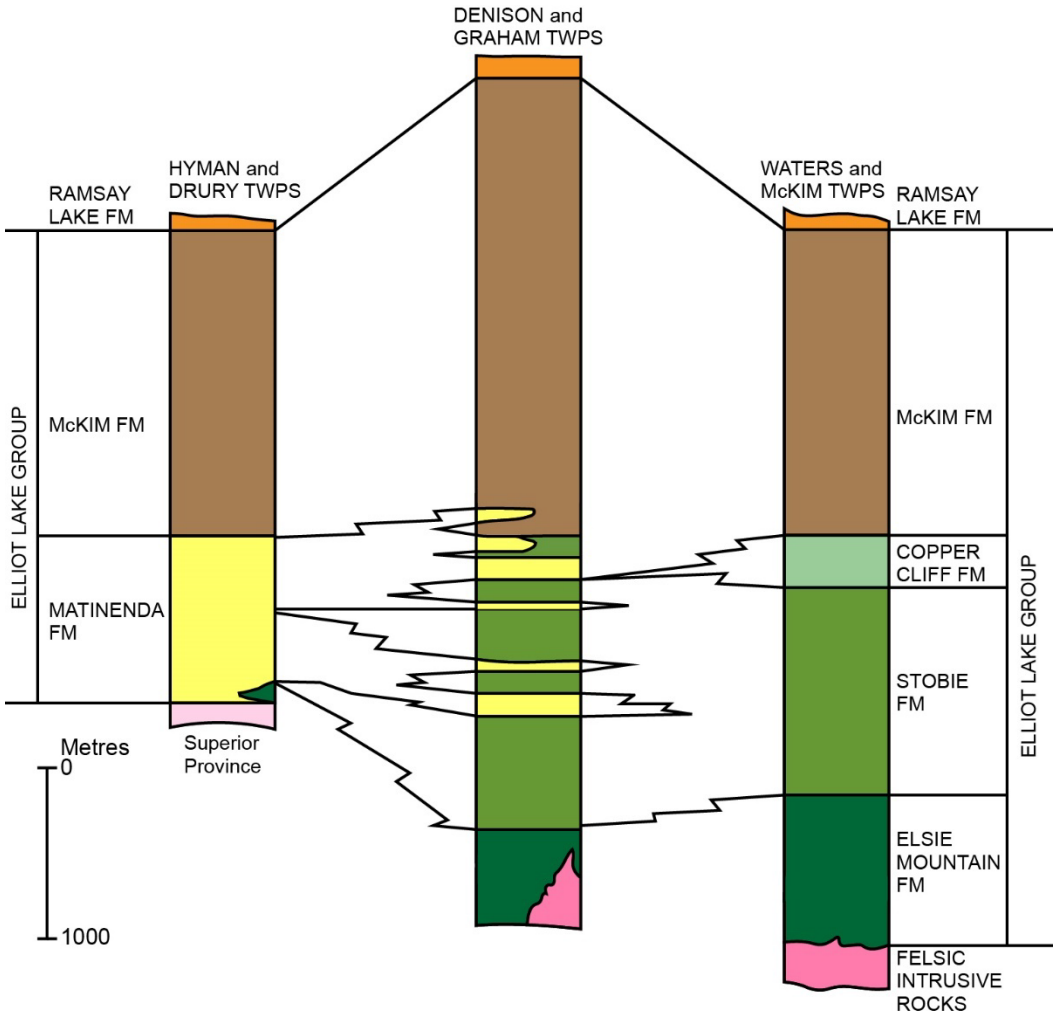


Figure 4. Generalized stratigraphy and facies relationships of the Elliot Lake Group in the Southwest Sudbury Structure (*modified from* Card 1978).

The Matinenda Formation, which hosts the uranium-rich pyritic quartz pebble conglomerates in the Elliot Lake area (Figure 1), is interpreted to have been deposited in a braided fluvial environment (Fralick and Miall 1989). In the southwestern Sudbury Structure, the Matinenda Formation is up to 1-km thick and thins eastward where it eventually disappears from the Stobie–McKim Formations contact but crops out as discontinuous layers within the Stobie Formation (Figure 3). The Matinenda Formation consists of subfeldspathic arenite and quartz arenite with quartz-pebble conglomeratic beds. The quartz-pebble conglomeratic beds locally contain pyrite and elevated concentrations of U and Th. The arenites, which constitute the bulk of the formation, are massive to crudely bedded, locally displaying graded beds and cross-bedding.

The McKim Formation, which is interpreted to represent a marine transgression that gradually drowned the Matinenda fluvial plain (Fralick and Maill 1989), is one of the most aerially extensive Huronian Supergroup units in the southwestern Sudbury Structure. The preserved sequence was significantly thickened (up to 1.5 km) by folding (Figure 3). The McKim Formation consists of interbedded mudstone, siltstone and minor sandstone. The amount of interbedded sandstone increases significantly eastward through Denison Township. The sandy and silty turbidites are

typically thickly laminated to thinly bedded and commonly display cross-bedding, graded beds, ripples and scour marks. Staurolite, chloritoid and rutile porphyroblasts are present, predominantly in the muddier beds.

Hough Lake Group

The Hough Lake Group contains sedimentary rocks of the Ramsay Lake, Pecors and Mississagi Formations. It is the lowest of the 3 groups that display the cyclical repetition of conglomerate—siltstone-mudstone—–sandstone. Each cycle is interpreted to represent a sequence of glaciogenic —marine—fluvial and/or shallow marine deposition (Roscoe 1969; Robertson 1976; Fralick and Miall 1989).

The Ramsay Lake Formation is up to 300 m thick and consists of 2 conglomerate units: the "beige" and "grey" members, and, an upper sandstone unit, the "sandy" member (Gordon et al. 2018a). The "beige" member is a bilithic conglomeratic subfeldspathic arenite with a beige, quartz-rich matrix and clasts of granite and quartz. The "grey" member is consistent with the classic description of the Ramsay Lake Formation (cf. Young 1991; Bennett et al. 1991). It is a heterolithic conglomeratic wacke or sandstone with a grey, quartz-rich matrix. The "sandy" member is composed of crudely bedded, subfeldspathic arenite, wacke and quartz arenite.

The Pecors Formation is 100 to 300 m thick and exposed between the Ramsay Lake and Mississagi Formations (Figure 3). The formation consists of siltstone and mudstone. The siltstone and mudstone are thickly laminated and locally exhibit cross-beds and load structures.

The Mississagi Formation is at least 2 km thick, including thickening by folding (Figure 3). Sandstones of the Mississagi Formation consist of well-sorted, fine- to medium-grained subfield-spathic arenite and quartz arenite. They are thinly to thickly bedded, locally contain beds of siltstone, and commonly display cross-beds.

Proterozoic Intrusive Rocks

Drury Township Intrusion, Matachewan Dike Swarm and Huronian Synvolcanic Intrusions

The Drury Township intrusion, Matachewan dike swarm, and volcanic rocks of the Huronian Supergroup were all emplaced during continental rifting associated with the Matachewan Igneous Event (Heaman 1997). The Drury Township intrusion and Matachewan dike swarm are also tentatively genetically linked to eruption of the basal Huronian Supergroup mafic volcanic rocks (Fahrig 1987; Vogel et al. 1998).

The Drury Township intrusion is interpreted as one of several leucogabbro-anorthosite sills of the East Bull Lake Intrusive Suite (Prevec 1993; Prevec and Baadsgaard 2005). Intrusions of the East Bull Lake Suite were emplaced between ~2491 and 2475 Ma and occur in a discontinuous east-northeast-trending belt along the Archean–Proterozoic contact between Elliot Lake and the Ottawa River (Krogh et al. 1984; James et al. 2002; Bleeker et al. 2012; Bleeker et al. 2015). The Drury Township intrusion is up to 2 km thick and is exposed at the Ramsey–Algoma granitoid complex – Huronian Supergroup – SIC contact in Drury Township (Figure 3). The intrusion is composed of medium- to coarse-grained, locally pegmatitic, vari-textured anorthositic gabbro with a marginal gabbroic phase.

The Matachewan dike swarm is an extensive radial dike swarm consisting of north- and northwest-trending mafic dikes, which intruded granitoids of the Superior Province and crop out over 300,000 km^2 in Ontario and southwestern Quebec (Halls and Bates 1990). The Matachewan dike swarm was emplaced in 2 main pulses. The first, earlier pulse at *circa* 2480 Ma is believed to have been coincident with emplacement of the East Bull Lake Intrusive Suite (Krogh et al. 1984; James et al. 2002; Bleeker et al. 2012; Bleeker et al. 2015). The second, and "main pulse" of the Matachewan dike swarm. occurred at *circa* 2460 Ma (Heaman 1997; Bleeker et al. 2012; Bleeker et al. 2015). In the southwestern Sudbury Structure, northwest- and northeast-trending Matachewan

dikes crosscut the Drury Township intrusion and the Archean Ramsey-Algoma granitoid complex (Figure 3). These mafic dikes are fine- to medium-grained and locally plagioclase-phyric. The northeast trend for the Matachewan dike swarm in the Sudbury area is atypical; it is possible that their trend represents a small-scale, concentric dike swarm (Gordon et al. 2018a).

Mafic sills that are geochemically similar to the Elsie Mountain and Stobie formations intrude the volcanic rocks of the Elsie Mountain and Stobie Formations as well as sedimentary rocks of the Matinenda and McKim Formations (Figure 3) (Gordon et al. 2018a; Gordon 2021, 2022). These mafic sills likely represent a combination of synvolcanic intrusions and intercalated mafic flows. Locally, peperite textures are preserved where the mafic sills intruded the McKim Formation, suggesting that Huronian Supergroup volcanism continued (or resumed) during deposition of the McKim Formation (Gordon et al. 2018a; Gordon 2021). Most of these mafic sills are 50-100 m wide, but a few larger intrusions are up to 200 m wide and 2 km long (Figure 3). All are roughly east-trending and fine- to medium-grained.

Creighton Pluton

The Creighton pluton is a 2455-2460 Ma subvolcanic sill that was emplaced into the Huronian Supergroup mafic volcanic rocks during rifting (Bleeker et al. 2015). It is interpreted as the high-level magma chamber to rhyolites of the Copper Cliff Formation (Bleeker et al. 2015). The western extent of the Creighton pluton crops out in the northeastern corner of Denison Township (Figure 3). It intruded the mafic volcanic rocks of the Stobie and Elsie Mountain Formations and is truncated to the north by the SIC. The pluton consists of leucocratic granite and porphyritic quartz monzonite. Inclusions of porphyritic quartz monzonite are locally found in the granite phase.

Nipissing Intrusive Suite

The voluminous Nipissing Intrusive Suite is exposed for over 400 km from the Ontario–Quebec border to Sault Ste. Marie. The intrusive suite was emplaced between 2219 to 2210 Ma (Davey et al. 2019; Corfu and Andrews 1986; Noble and Lightfoot 1992; Bleeker et al. 2015) as part of the Ungava large igneous province during continental rifting (Ernst and Bleeker 2010; Davey et al. 2019).

In the southwestern Sudbury Structure, numerous Nipissing sills intruded the Huronian Supergroup and adjacent Archean basement rocks (Figure 3). Individual sills can reach up to 400 m thick and extend over 2 km in length. The larger Nipissing sills typically occur near or at the contact between the Ramsay Lake and McKim Formations. These larger sills are crudely differentiated, with gabbro and melagabbro phases. Pegmatoidal and/or anorthositic pods occur within the thicker portions of the sills. Smaller, narrower sills, on the other hand, are undifferentiated.

Trap Dike Swarm

In the Sudbury area, east- to east-northeast-trending mafic dikes, interpreted to belong to the Trap dike swarm (*circ*a 1750 Ma; cf. Bleeker et al. 2015), crosscut the Worthington Offset dike, Sudbury Breccia, folded Nipissing sills and Huronian Supergroup stratigraphy (Figure 3). These dikes are typically less than 50 in width and consist of undeformed quartz diabase. Field relationships suggest that these mafic dikes postdate the SIC event and they are tentatively linked to a post-Penokean rifting event (cf. Bleeker et al. 2015).

Dikes of Unknown Affinity

East-northeast-trending mafic dikes of similar appearance to the Trap dikes, but which are geochemically distinct, crosscut the Archean basement, folded Nipissing sills and Huronian Supergroup strata (Gordon et al. 2018a). The mafic dikes are fine- to medium-grained, locally plagioclase-phyric and are variably foliated. In the southern central portion of Drury Township, northwest-trending felsic dikes crosscut folded Nipissing sills. The felsic dikes are up to 5 m wide, quartz-rich, massive and contain inclusions of Nipissing gabbro (Gordon et al. 2018a).

These mafic and felsic dikes have not been assigned to any known intrusive suite and likely represent one or more magmatic events that occurred after regional folding. They are not included on Figure 3; readers are referred to Gordon et al. (2018a) for more information.

Sudbury Dike Swarm

The Mesoproterozoic Sudbury dike swarm (*circa* 1238 Ma: Krogh et al. 1987) consists of northwest-trending olivine diabase dikes that crosscut the Superior and Southern provinces, and their deformed and metamorphosed equivalents occur within the northwestern Grenville Province (Ketchum and Davison 2000). The Sudbury dike swarm extends ~300 km west and northwest from the Sudbury area. Different tectonic settings for their emplacement have been proposed, which include continental rifting (Shellnut and MacRae 2012), a back arc setting (Ernst and Bleeker 2010), or mantle-plume upwelling (Easton et al. 2021).

In the southwestern Sudbury Structure, undeformed, northwest-trending dikes of the Sudbury dike swarm crosscut the Ramsey-Algoma granitoid complex, the Drury Township intrusion, rocks of the Huronian Supergroup and the SIC (Figure 3). The Sudbury dikes consist of undeformed olivine diabase, are fine- to medium-grained and locally plagioclase-phyric.

Structure

The southwestern Sudbury Structure displays multiple generations of structural fabrics and major structures, which are described below in chronological order.

Murray Fault and Northwest-Trending Faults

The Murray Fault is a major east- to east-northeast-trending structural feature in the Southern Province (Figure 1). It is thought to have originated as an extensional fault during Huronian sedimentation and has been periodically reactivated during subsequent tectonic events (Card and Hutchinson 1972; Card 1978; Zolnai et al. 1984). The Murray Fault truncates the Huronian Supergroup strata in the southeastern area of Denison Township (Figure 3), where rocks of the McKim Formation are juxtaposed against rocks of the Mississagi, Pecors and Ramsay Lake Formations.

Northwest-trending faults produced significant offsets of the Huronian Supergroup strata. In Denison Township, rocks of the Stobie and McKim Formations are juxtaposed against rocks of the Elsie Mountain Formation along a northwest-trending fault (Figure 3). This configuration of Huronian Supergroup strata is truncated to the north by the SIC and crosscut by the Creighton-Victoria deformation zone. Reactivation of the northwest-trending faults is indicated by sheared sublayer along the fault. These structures may represent southward extension of the similarly oriented Onaping Fault system which, like the Murray Fault, is thought to have originated as extensional faults during deposition of the Huronian Supergroup and reactivated during subsequent orogenic events (cf. Card 1978; Zolnai et al. 1984).

Pre-impact Foliation

An early, southeast-trending foliation is locally preserved within basaltic rocks of the Elsie Mountain Formation adjacent to the Creighton Pluton contact. This foliation is locally crosscut by Sudbury Breccia, which also contains randomly oriented clasts of the foliated basalt and granitoid indicating that deformation started before the Sudbury impact event (Gordon 2018).

Bedding Parallel Thrust Faults

Bedding-parallel thrust faults are recognized on the basis of stratigraphic repetition and/or absence of specific Huronian Supergroup formations. The thrust faults are folded along with Huronian Supergroup strata (Gordon et al. 2018a; Généreux et al. 2018) and, thus, predate regional folding. In Drury Township, a bedding-parallel foliation is associated with these thrust faults and is locally preserved in the matrix of Sudbury Breccia, indicating that these faults formed after the impact event (Généreux et al. 2018).

Regional Folds, Foliation and Lineation

The present structural configuration of the Huronian Supergroup strata in Drury Township is strongly controlled by kilometre-scale, northeast-trending isoclinal folds. Adjacent to the Superior–Southern provinces contact, smaller scale folds are west-northwest to east-trending, following the orientation of the Superior–Southern provinces contact. Throughout Drury Township, Nipissing sills are folded along with the Huronian Supergroup strata or follow their axial trace. Folding decreases in intensity eastward, into Denison Township, and appears to be restricted to the McKim Formation (Figure 3).

A moderate to strong east- to east-northeast-trending regional foliation is ubiquitous within the Huronian Supergroup stratigraphy and Sudbury Breccia. The regional foliation overprints Sudbury Breccia, and locally crenulates the bedding-parallel foliation associated with the early thrust faults (Généreux et al. 2018). It is axial planar to folds on outcrop and contains a southeast- to southwest-plunging stretching lineation that parallels fold axes. Both fabrics also rotate along with fold orientations adjacent to the Southern-Superior provinces contact (Généreux et al. 2018).

Creighton-Victoria Deformation Zone

The Creighton-Victoria deformation zone (CVDZ) consists of an east-southeast-trending, 200 to 400-m wide mylonite zone that occurs along the contact between the Archean Superior Province and Paleoproterozoic Southern Province across Drury Township (Figure 3) (Gordon et al. 2015; Simard et al. 2016; Généreux et al. 2016; Généreux et al. 2017; Gordon et al. 2018a). Eastward into Denison Township, the CVDZ extends into a 1.5 km wide deformation corridor bound by the Creighton and Victoria faults (Généreux et al. 2017). The corridor consists of discrete, 5 to 20-m wide, shear zones that follow internal contacts within weakly to moderately foliated Huronian volcanic and sedimentary rocks.

The CVDZ is characterized by a strong west-northwest- to east-trending subvertical foliation, and a steeply plunging eastward-trending stretching lineation. Shear sense indicators correspond to dextral (horizontal) shearing in the west, and south-over-north dextral (oblique) shearing to the east (Généreux et al. 2016; Généreux, et al. 2017). The development of the CVDZ is interpreted as coeval with regional folding (Généreux et al. 2018).

Northeast-trending Faults

This fault group includes the Cameron Creek, Fairbank Lake, Chicago and Vermilion Lake faults, as well as unnamed faults of similar orientation (Figure 3). These northeast-trending brittle-ductile faults produce significant offset of SIC-related rocks, folded Huronian Supergroup strata and Nipissing sills.

Minor Structures

Shattercones, which are impact-related conical fractures with distinctive cone or fan-shaped features, are found within arenites and conglomerates of the Mississagi and Ramsey Lake Formations, respectively.

The regional foliation and mylonitic fabric are locally overprinted by north-northwest-trending crenulation cleavage and kink bands. Most kink bands observed are S-shaped and locally form conjugate sets of centimetre-scale box-folds (cf. Généreux et al. 2016; Gordon et al. 2018a). Local, brittle, northwest-trending faults also crosscut all rock types, including Sudbury dikes.

FIELD TRIP DETAILS

Geological Maps

Geological compilation maps covering all or parts of the area of the field trip include Ames et al. (2005), Dressler (1984b) and Card and Lumbers (1973). A detailed geology map is available for Drury Township (Gordon et al. 2018b).

ROAD LOG

Note: Caution should be taken when parking vehicles on the shoulder of the roads and when examining outcrops along any road along the field trip route. All UTM co-ordinates are given in NAD 83 datum, zone 17.

Figure 3 shows the location of the field trip stops. The mileages in the road log represent the distance from one stop to another.

38.5 km (30 minutes) – Starting at the Willet Green Miller Centre in Sudbury, head west toward Ramsey Lake Road. Turn left onto Ramsey Lake Road and continue west for 1.9 km. Use the left 2 lanes to turn left (south) on Paris Street and continue for 4 km. Use the right lane to take the Highway 17W ramp to Sault Ste. Marie. After 800 m, continue straight to merge onto Hwy 17. Drive west on Hwy 17 for 30 km. Turn right (north) onto Fairbank Lake/Totten Mine Road and continue for 1.1 km. Stop 1 will be on the right (east) side of the road.

Stop 1 Nipissing sill, Trap and Sudbury dikes

UTM coordinates 0471368E 5136502N

Exposed on the east side of Fairbank Lake/Totten Mine Road is a large outcrop of Nipissing gabbro intruded by mafic dikes of the Trap and Sudbury dike swarms (Figure 3). The Nipissing gabbro is part of a northwest-trending sill that is up to 300-m wide and 7-km long (Figure 3). The Nipissing gabbro is green-grey, massive and medium-grained. Along its northern margin, it is intruded by a Trap dike (Figure 5). The contact between dikes is sharp, strikes east-northeast and dips steeply southward.

The Trap dike is massive, grey in colour, homogeneous and fine-grained. It is at least 20-m wide. Along the contact with the Trap dike, the Nipissing gabbro is strongly amphibolitized. Both the Nipissing sill and the Trap dike were metamorphosed to greenschist facies. Plagioclase is partially sausseritized and pyroxene is completely replaced by amphibole and, locally, biotite.

Approximately 50 m south of the Trap dike, the Nipissing sill is intruded by olivine diabase of the Sudbury dike swarm (Figure 6). This Sudbury dike can be traced for over 8 km and truncates the folded Huronian Supergroup stratigraphy and the Worthington Offset dike (Figure 3). The contact between the Sudbury dike and Nipissing sill is sharp, strikes northwest and dips steeply northward. The Sudbury dike is at least 25 m wide and has a distinctive brown weathered surface. It is magnetic, massive, fine- to medium-grained and plagioclase-phyric. It is also relatively unaltered, its primary mineralogy consisting of plagioclase, clinopyroxene, orthopyroxene and olivine. The Sudbury dike is truncated by late, northwest-trending brittle faults adjacent to the southern contact with the Nipissing gabbro and within the central area of the exposure (Figure 6).

6.4 km (~10 minutes) – Head north on Fairbank Lake Road toward Bay Street and drive for 1.4 km. Turn right (north) onto Crean Hill Road and continue for 2.1 km. Turn right at the fork to stay on Crean Hill Road and continue for 2.9 km. Stop 2 is past the gates and will be on the right (east) side of the road.

Stop 2 Crean Hill - Footwall Breccia

UTM coordinates 0473051E 5141707N

Just beyond the gates and on both sides of the access road are stripped outcrops that expose variably mineralized Crean Hill Footwall Breccia. The exposure is within 200 m of the vertically-dipping SIC contact. The open pit of the past-producing Crean Hill Mine is located just beyond the fence at the northern edge of the outcrops (Figure 7).



Figure 5. Nipissing gabbro crosscut by a northeast-trending Trap dike (Stop 1).

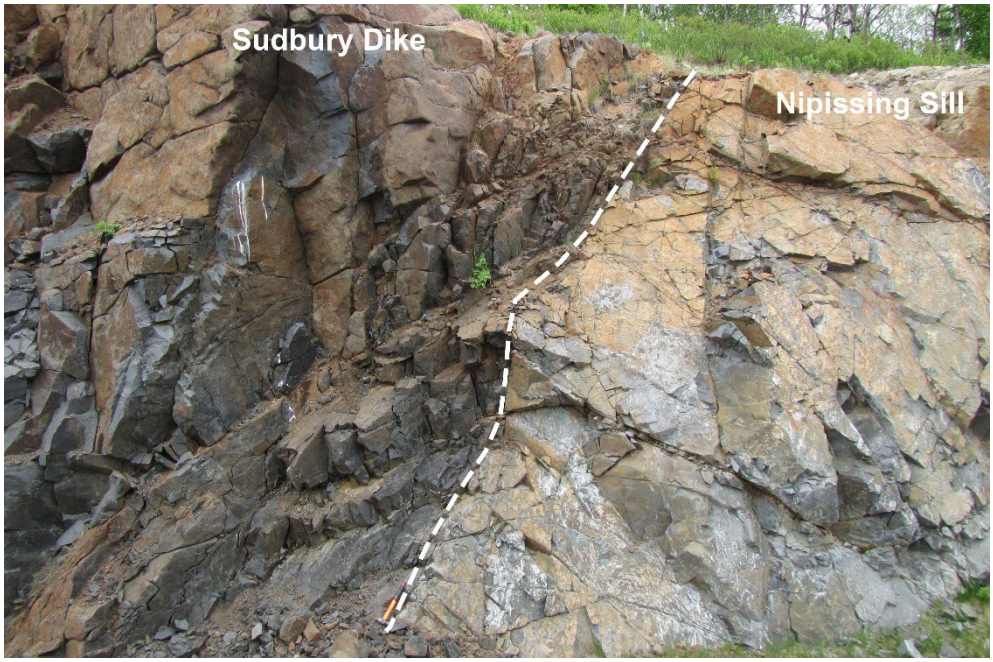


Figure 6. Sudbury dike in contact with Nipissing gabbro and crosscut by northwest-trending brittle faults (Stop 1).



Figure 7. Partial view of the Crean Hill outcrops with the Crean Hill open pit in the background (Stop 2).

Crean Hill mine, currently owned by Vale Canada Limited, operated sporadically from 1909 to 2000, producing 744,747 tons grading 2.14% Ni and 2.9% Cu to the end of 1916 (Card 1968). The main orebody consisted of sulphide breccia hosted in brecciated Huronian Supergroup volcanic rocks along the SIC contact (Coleman 1913; Knight 1917; Card 1968), with disseminated low-sulphide PGE mineralization occurring in the underlying brecciated footwall rocks.

The Crean Hill stripped outcrops expose brecciated basalt with minor dacite and quartzofeldspathic arenite of the Huronian Supergroup, which are crosscut by dikes and pods of Footwall Breccia (Figure 8). Four compositions of breccia are found in the outcrop: monolithic felsic, mafic, and quartzofeldspathic breccias, and heterolithic dioritic breccias (Généreux et al. 2021).

Monolithic breccias occur as irregular veins in dacite (felsic breccia) and basalt (mafic and quartzofeldspathic breccias). Partial melt patches make up 5–20 vol.% of the host rocks, and consist of quartz and plagioclase, with hornblende porphyroblasts. They terminate *en biseau* and cut across mafic breccia veins.

Heterolithic dioritic breccias occur as pods (bilithic breccia), dikes (breccia dikes), and anastomosing veins (mixed breccia) within basalt (Figure 9). They formed as melts with contact-parallel flow textures, which are defined by elongate wispy clasts that wrap around basalt clasts. Partial melt textures are not observed within the dioritic breccias, but narrow (<1 cm) quartz-plagioclase leucosomes occur along their contact with the host basalt.

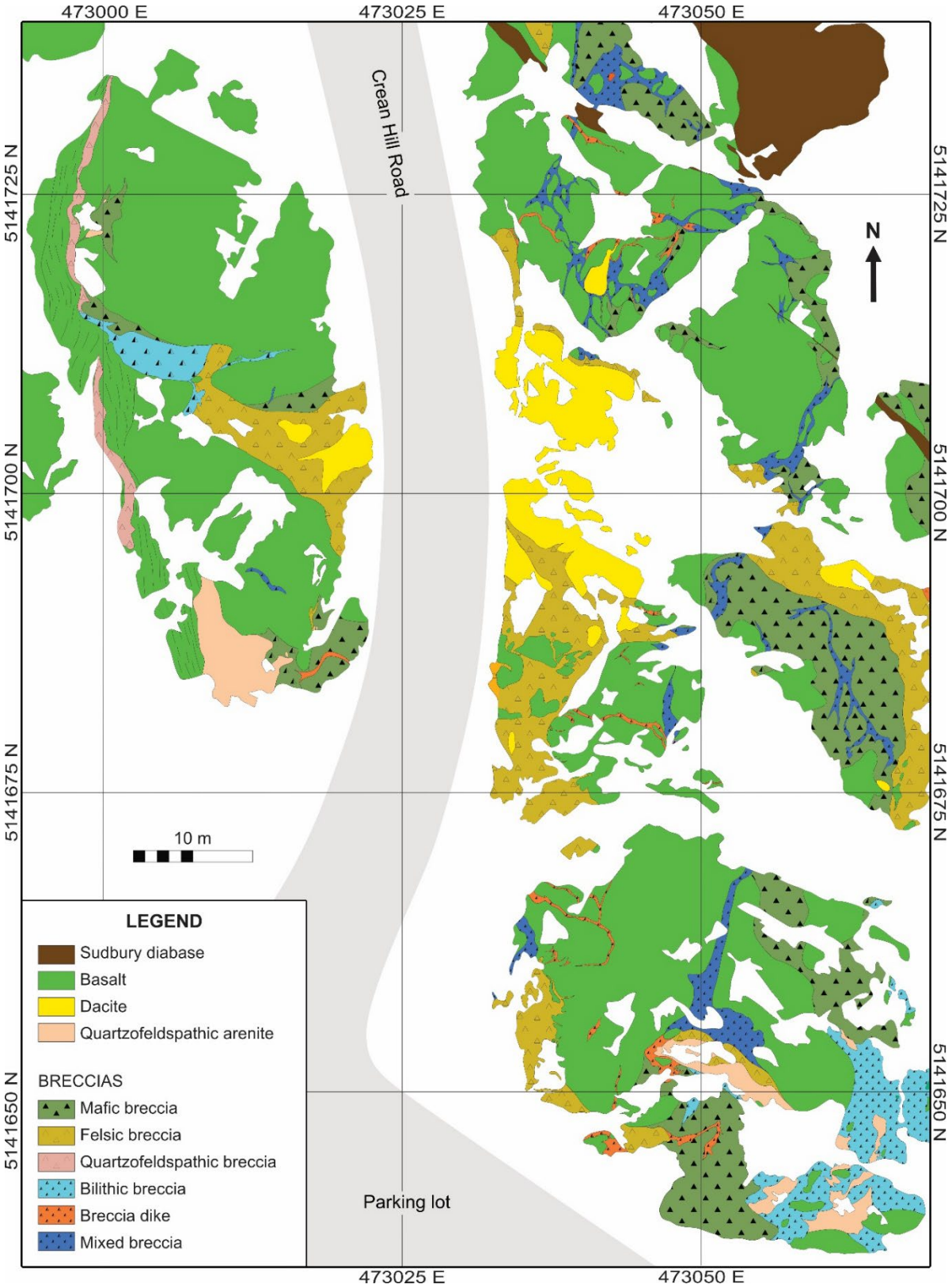


Figure 8. Geological map of the Crean Hill outcrops (*from* Généreux et al. 2021). UTM coordinates are in NAD 83, Zone 17 (Stop 2).

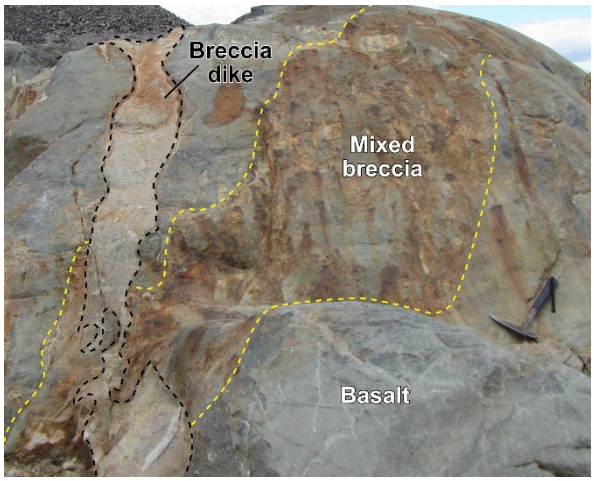


Figure 9. Mixed breccia unit crosscut by a heterolithic dioritic breccia dike (*from* Généreux et al. 2021) (Stop 2).

A detailed study of the Crean Hill breccias by Généreux et al. (2021) showed that they have significantly lower SiO_2 and higher TiO_2 than the SIC, suggesting that the breccias likely did not form by injection of SIC melt into the fractured target rocks. Modeling of partial melt compositions during contact metamorphism showed that the dioritic breccia matrices are too mafic to have formed by anatexis during contact metamorphism. Instead, their composition mirrors that of their host rocks, thus they are best interpreted as locally-derived shock melts that formed during shock compression and which were trapped in the basement rocks during cooling of the SIC (Généreux et al. 2021). The breccias were subsequently modified by contact metamorphism ($T \geq 750°C$) during cooling of the melt sheet, and by later syn-tectonic regional metamorphism at upper greenschist to amphibolite conditions.

Low-sulphide PGE mineralization is hosted in anastomosing veins of 'mixed' dioritic breccia, which contain a medium-grained dioritic matrix intermingled with a fine-grained basaltic matrix. This mixed breccia is locally crosscut by irregularly shaped dioritic breccia dikes (Figure 9) and displays a strong mottled texture that is further complicated by the presence of leucosomes.

Mineralized mixed breccia appears gossanous on outcrop, containing <5 vol.% of disseminated pyrrhotite, pentlandite, chalcopyrite, gersdorffite and precious metal minerals (PMM), which are commonly associated with epidote-quartz(-calcite) alteration patches. Shear zones appear to be the main control on the distribution of PGE (Gibson et al. 2010), with strongly foliated mixed breccia generally enriched in Au, Pt and Pd.

1.6 km (~3 minutes) – Head southwest on Crean Hill Road toward Fairbank Lake Road and drive for 1.4 km. Turn left (southeast) on the gravel road. Continue east on the gravel road for 200 m, keep left at the fork. Stop 3 is on the north side of the road.

Stop 3 Vermilion Mine – Vermilion Offset Dike, Sudbury Breccia and Stobie Formation

UTM coordinates 0472251 E 5140280 N

Francis L. Sperry discovered sperrylite ($PtAs_2$) at the Vermilion mine (Wells 1889), which is also the type locality for arsenohauchecornite ($Ni_{18}Bi_3AsS_{16}$), michenerite (PdBiTe) and violarite ($FeNi_2S_4$). The mine operated from 1887 to 1916 and produced over 4000 tonnes of ore with exceptionally high grade of 6.64% Ni and 6.89% Cu, including 180 tonnes at 20-25% Cu-Ni, 125 g/t Ag, 125 g/t Pd, 46.9 g/t Pt and 10.3 g/t Au (Holloway et al. 1917). The stripped outcrop exposes the Vermilion Offset quartz diorite (Figure 10 and 11). This offset dike is not connected to the Main Mass of the SIC (Grant and Bite 1984), but instead forms a lens at the contact between Sudbury Breccia and mafic volcanic rocks of the Stobie Formation (Figure 10).

At the northernmost part of the outcrop are slightly deformed basaltic volcanic rocks of the Stobie Formation, where bedded lapilli tuff is interlayered with vesicular basalt, pillow breccia and possible flow-top breccia (Figure 12). Volcanic textures such as hyaloclastite, amygdules, and lapilli are generally well-preserved but are locally overprinted by coarse acicular amphiboles.

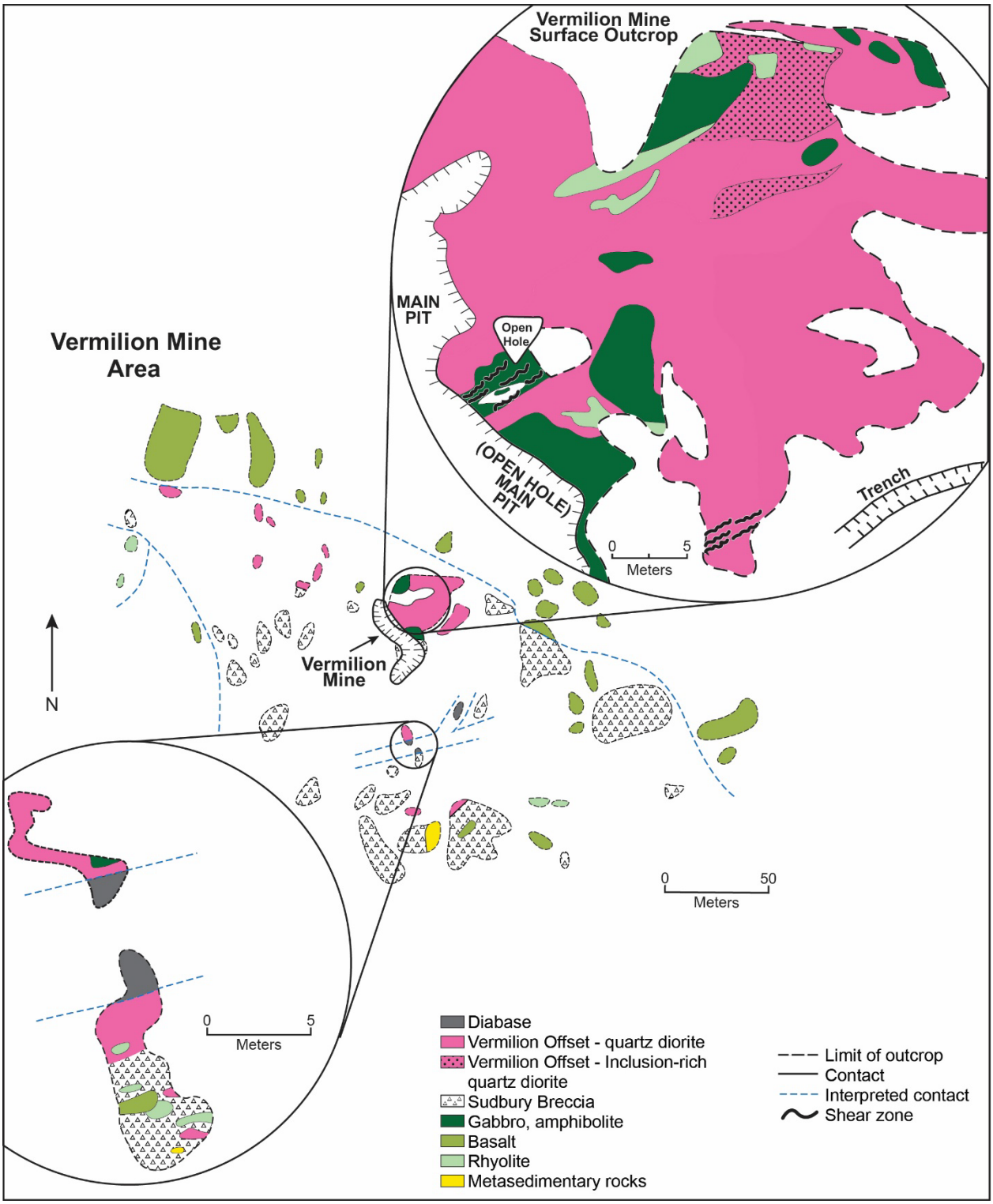


Figure 10. Geology of the Vermilion Offset (*modified from* Grant and Bite 1984). Detailed geology of the Vermilion Mine surface outcrop *modified from* Lightfoot et al. (1997) (Stop 3).



Figure 11. Overview of the stripped outcrop at the Vermilion mine (Stop 3).

The central portion of the main outcrop exposes the Vermilion quartz diorite, which is generally medium-grained and contains up to 20% of small (<5 cm), partially digested, felsic and mafic clasts. At the top of the main outcrop is a 10-m wide section of finer-grained and foliated inclusion-bearing quartz diorite (IQD), which contains 20-40% of dacite, basalt and amphibolite clasts that are up to 10 m in size. IQD is in sheared contact with basaltic rocks to the north and Sudbury Breccia to the South.



Figure 12. Basaltic pillow breccia with well-preserved hyaloclastite (Stop 3).



Figure 13. Matrix-supported Sudbury Breccia with elongated clasts (Stop 3).

Sudbury Breccia is best exposed south and southwest of the main quartz diorite outcrop, closest to the parking area. It occurs as variably foliated, matrix-supported, heterolithic breccia (Figure 13) within brecciated mafic volcanic rocks.

The Vermilion ore occurred in shear-hosted sulphide veins and irregular lenses ranging from a few centimetres to 40 cm in diameter. The shear zones are found mainly in Sudbury Breccia and strike parallel to the quartz diorite–Sudbury Breccia contact (Szentpéteri et al. 2003). Disseminated sulphides and platinum-group minerals (PGM) also occur in quartz diorite and foliated Sudbury Breccia, and are associated with irregularly distributed epidote-albite-chlorite alteration patches that range from a few centimetres to several metres in size. The close spatial association of PGM with alteration minerals, their finely disseminated nature, the presence of sulphides-PGM in secondary hydrothermal veins, and the occurrence of sulphide veins within shear zones all suggest a complex multistage magmatic-hydrothermal and metamorphic-hydrothermal origin of the sulphide-PGM assemblages (Szentpéteri et al. 2003).

4 km (15 minutes) – Return to Crean Hill Road (~200 m). Turn left (southwest) toward Fairbank Lake Road for 1.3 km. Turn right (northwest) on Fairbank East Road and continue north for approximately 1 km. Turn right (east) on the gravel road. Drive east for 1.5 km. Park at marked trail. Walk north on trail for 150 m. Trail leads to outcrops exposed along the powerline.

Stop 4 Creighton Fault and Stobie Formation

UTM coordinates 474656E 5140835N

Along the powerline is an excellent exposure of ductile deformation observed within the deformation corridor bound by the Creighton and Victoria Faults. The Creighton Fault can be traced across most of Denison Township as a prominent, relatively continuous, topographic low with periodic outcrops. Along the powerline, a 20-m wide flow-banded dacite is in sheared contact with a relatively massive, locally amygdaloidal, basaltic flow. Both the intermediate and mafic volcanic rocks are part of the Stobie Formation.

Ductile deformation is expressed in the intermediate unit as a strong east-northeast-trending foliation that is consistently oriented counter-clockwise to flow banding, suggesting apparent dextral shearing (Figure 14). An intersection lineation between the flow banding and the foliation steeply plunges to the southeast. The same kinematic indicators are observed in other shear zones in the area, including the Creighton-Victoria mylonite zone farther west (Généreux et al. 2017).

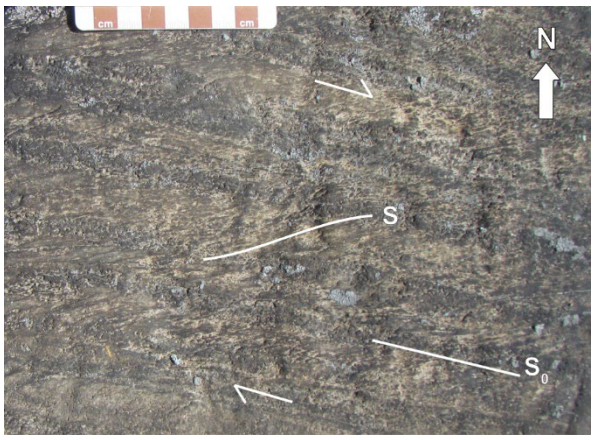


Figure 14. Strong pervasive foliation (S) in felsic volcanic rock consistently oriented counter-clockwise to flow banding (S_0), suggesting apparent dextral shearing (*from* Généreux et al. 2017) (Stop 4).

13 km (25 minutes) – Walk south 150 m along trail to return to the gravel road. Travel west for 1.5 km. Turn left (south) on Fairbank East Road and continue for 1.1 km. Continue south onto Crean Hill Road for 2.1 km. Turn right (west) onto Fairbank Lake Road. Fairbank Lake Road turns into Spanish River Road after 7 km, continue along Spanish River Road for another 4 km. Where Spanish River Road turns southward, continue straight (west) onto High Falls Road for 200 m. Stop 5 is at the bend along High Falls Road. The outcrops are exposed north of the road. Use the small gravel road adjacent to the outcrop for parking.

Stop 5 Synvolcanic mafic sill and McKim Formation

UTM coordinates 0460806E 5135978N

On the north side of Spanish River Road, a large outcrop contains turbidites of the McKim Formation and a synvolcanic mafic sill (Figure 3).

The McKim Formation consists of thickly laminated mudstone with siltstone layers, and contains porphyroblasts of chloritoid and staurolite (Figure 15A). Bedding is subvertical and trends northeast. There is a strongly developed bedding-parallel foliation overprinted by regional east-northeast-trending foliation (Figure 15B).

North of the McKim Formation a 50-60 m wide mafic sill is exposed. The sill trends parallel to bedding and exhibits distinctly rounded and lobate, mafic enclaves that are enclosed in a felsic, mica-rich and garnet-bearing matrix (Figure 16). This texture has been interpreted as a peperite, formed as a result of a mafic sill intruding what were originally unconsolidated wet sediments of the McKim Formation. The mafic enclaves are geochemically similar to that of the mafic volcanic rocks of the Elsie Mountain Formation (Gordon et al. 2018; Gordon 2021, 2022).

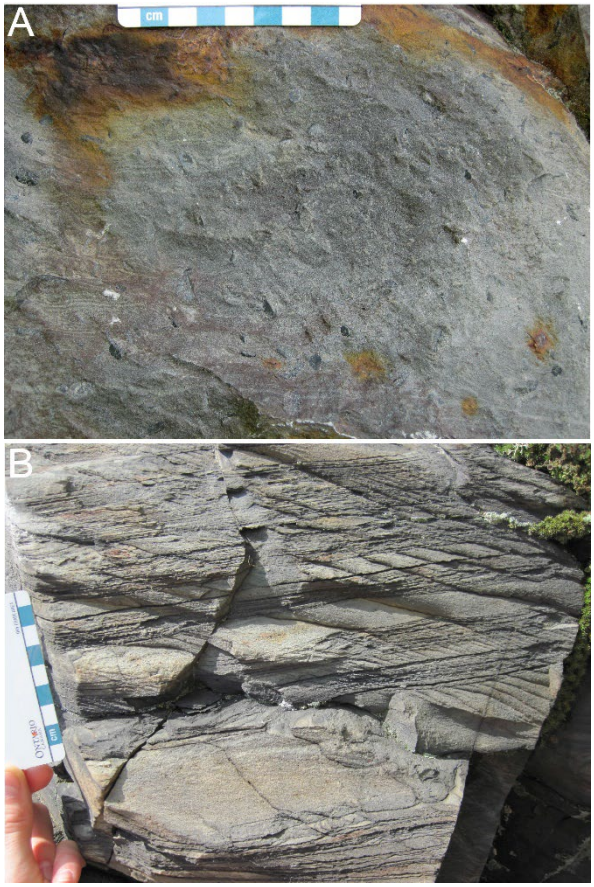


Figure 15. A) Laminated mudstones of the McKim Formation with chloritoid and staurolite porphyroblasts. **B)** Bedding parallel foliation in thickly laminated mudstones of the McKim Formation overprinted by regional east-northeast-trending foliation, vertical face (Stop 5).



Figure 16. Peperite with fine-grained, amoeboid mafic lobes encompassed by a felsic matrix (Stop 5).

10 km (15 minutes) – Drive east on High Falls Road and continue straight onto Spanish River Road for 4 km, turn left (north) on Fairbank Lake Road, drive north for 5.5 km. Stop 6 is on the left (west) side of Fairbank Lake Road.

Stop 6 Drury Township Intrusion and Matachewan dike swarm

UTM coordinates 0464241E 5142157N

On the west side of Fairbank Lake Road, outcrops of the Drury Township intrusion are exposed. The Drury Township intrusion consists of anorthositic gabbro that is medium- to coarse-grained, locally pegmatitic and vari-textured (Figure 17A). Mineralogy is characterized by greenschist grade assemblages. Plagioclase is almost entirely saussuritized and pyroxenes are replaced by amphibole and chlorite. On the east side of the road, the anorthositic gabbro varies from undeformed to mylonitized (Figure 17B). Where deformed, shear zones and foliation trend northeast and are parallel to the adjacent Chicago Fault (Figure 3). The anorthositic gabbro is also crosscut by a narrow, plagioclase-phyric mafic dike of the Matachewan dike swarm. The Matachewan dike is fine-grained, massive and metamorphosed to greenschist facies.

8 km (10 minutes) – Drive south on Fairbank Lake Road for 5.5 km. Turn left (east) to stay on Fairbank Lake Road and continue for 1.8 km. Turn left (north) on Kidd Copper Mine Road and drive north for 700 m. Stop 7a is on the left (west) side of Kidd Copper Mine Road.

Stop 7 Aer Kidd Property - Worthington Offset Dike

The beginning of Kidd Copper Mine Road is Vale Canada Limited property and is gated. The property changes ownership to SPC Nickel Corp. at the crest of the hill just before the Howland pit. The Worthington Offset dike crops out on the west side of the road and is intermittently exposed within the Aer-Kidd Property from the Howland pit to Perch Lake (Figure 18).



Figure 17. A) Coarse-grained to pegmatitic anorthositic gabbro of the Drury Township intrusion. **B)** Mylonitized anorthositic gabbro of the Drury Township intrusion (Stop 6).

The Worthington Offset dike has been mined periodically since 1885 and is host to Sudbury's most recently developed deposit, Totten Mine, with grades of 1.42% Ni, 1.9% Cu and 4.8 g/t PGM (Lightfoot 2017; Lightfoot and Farrow 2002). Three past producing mines can be found on the Aer-Kidd property (Figure 18). The first is the Howland Pit, which has been filled in and partially reclaimed. The second is the Robinson Mine (Stop 7b), which includes a shaft cap, a large, fenced hole, and small adit (Figure 19). The third past-producing site is the Rosen and Gersdorffite mines (also known as Aer Mine), which are located northeast of the old mill site and contains a shaft cap.

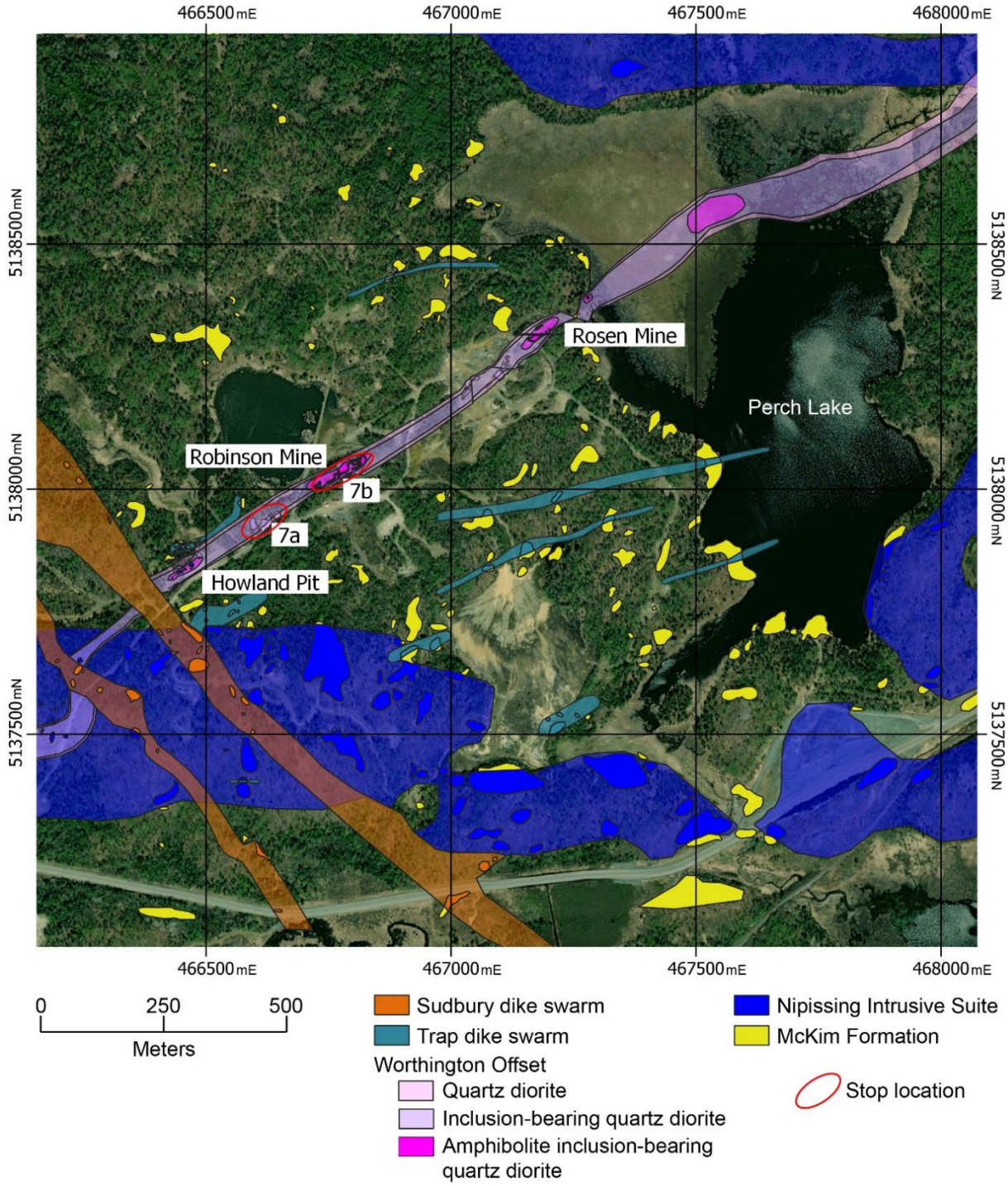


Figure 18. Geological map of the Aer-Kidd Property along the Worthington Offset dike. Past-producing mines are located within the dike where amphibole- and inclusion-bearing quartz diorite (AIQD) pods are observed. Stop 7a and 7b are located between the Howland Pit and Robinson Mine. UTM coordinates are in NAD 83, Zone 17 (Stop 7).

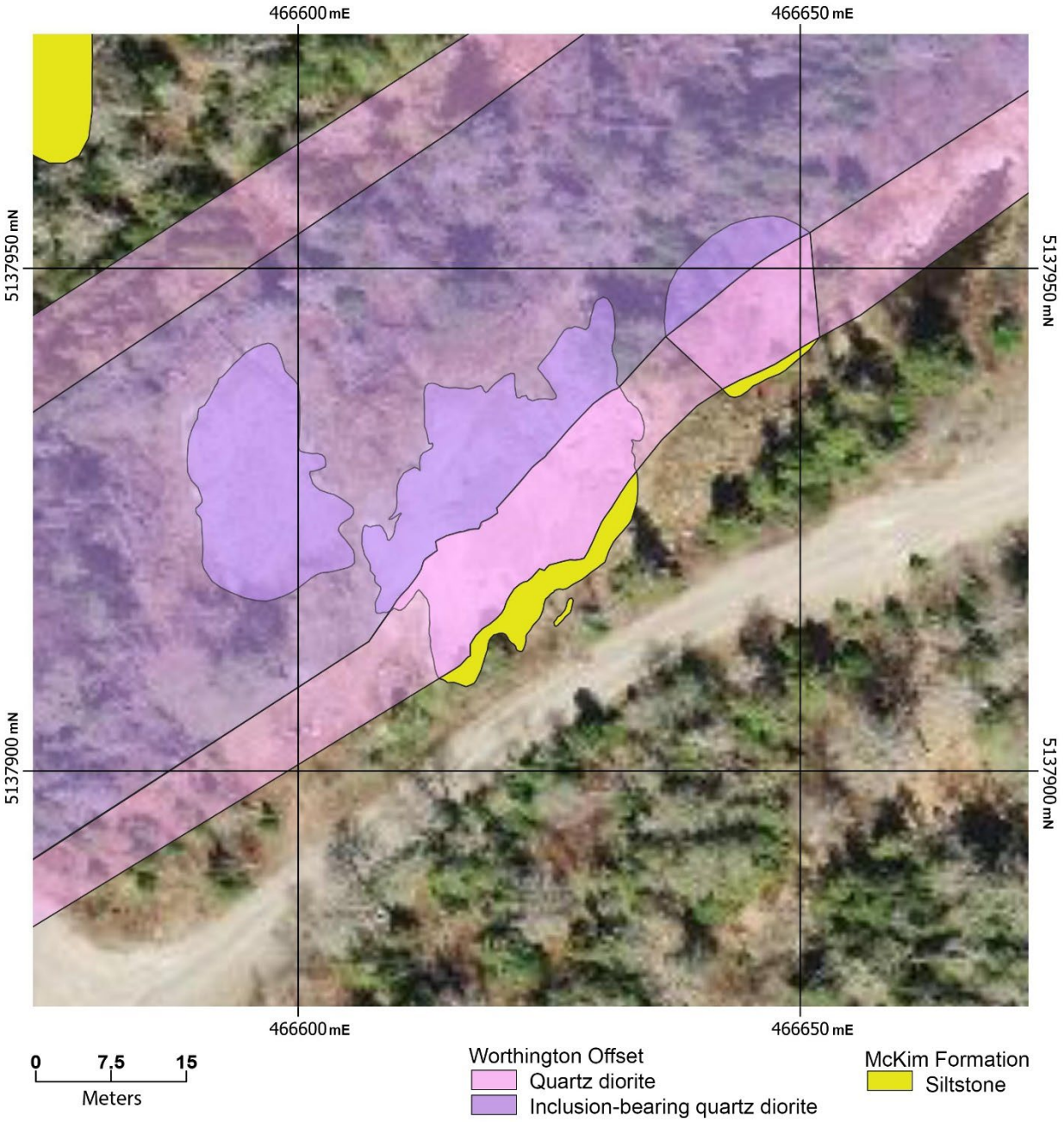


Figure 19. Geological map of the stripped outcrop between the Howland Pit and the Robinson Mine sites (Stop 7a). The Stop includes siltstones of the McKim Formation, as well as the quartz-diorite (QD) and inclusion-bearing quartz-diorite (IQD) phases of the Worthington Offset dike. UTM coordinates are in NAD 83, Zone 17.

There are many stripped and forested outcrops of the Worthington Offset dike over the length of the property. The two best surface exposures are: 1) the large, stripped outcrop on the left side of the road between the Robinson Mine and Howland Pit (Stop 7a) (Figure18), and 2) the Robinson Mine site (Stop 7b) (Figure 18, 19).

Stop 7a: Worthington Offset Dike and McKim Formation

UTM coordinates 0466620E 5137925N

On the west side of Kid Copper Mine Road, past the Howland Pit and core farm, is a large, stripped outcrop sloping north. This outcrop shows the phase separation of the dike well and is representative of the weakly mineralized to unmineralized portions of the Worthington Offset.

The outcrop contains a small section of McKim Formation siltstone (near the road) in sharp contact with Worthington quartz diorite (QD) (Figure 19). The siltstone is thickly laminated and exhibits a moderate east-northeast-trending foliation. QD is medium-grained, massive, with local veins and jointing, and contains several rounded sedimentary enclaves adjacent to the contact with the siltstone (Figure 20A). Farther north, the QD phase is crosscut by inclusion-bearing quartz diorite (IQD). The IQD phase is heterolithic and contains inclusions of QD, amphibolite, basalt and siltstone that are enclosed in a massive, medium-grained quartz diorite matrix (Figure 20B). Most inclusions are subrounded and range from a centimeter to sub-meter in diameter. Sulphide blebs are visible throughout the matrix of the IQD.

Stop 7b: Mineralization in the Worthington Offset Dike

UTM coordinates 0466765E 5138015N

Northward up the hill, toward the fenced hole and behind the shaft cap, is the Robinson Mine outcrop. The hill exposes a large portion of the Worthington Offset dike, but the best exposure is the stripped outcrop located south of the fence adjacent to the Robinson pit (Figure 21). The southern edge of the outcrop consists of weakly foliated siltstone of the McKim Formation. QD is found farther up the hill, but it is much thinner than that at Stop 7a. Further north is the sharp contact between the QD and IQD. The mineralized matrix of the IQD at this site has a slightly higher grade than at Stop 7a. Massive sulphides are found further north, just below the fence.



Figure 20. A) Sharp and linear contact between the Worthington Offset dike and the McKim Formation. Note the rounded sedimentary enclaves in QD near the contact. **B)** Heterolithic, massive IQD with sub-rounded inclusions derived from local host rocks. Note the pervasive sulphide burns throughout the quartz diorite matrix (Stop 7a).

200 m (1 minute) – Continue northeast along Kidd Copper Mine Road for 200 m. Stop 7b is on the left (north) side of the road.

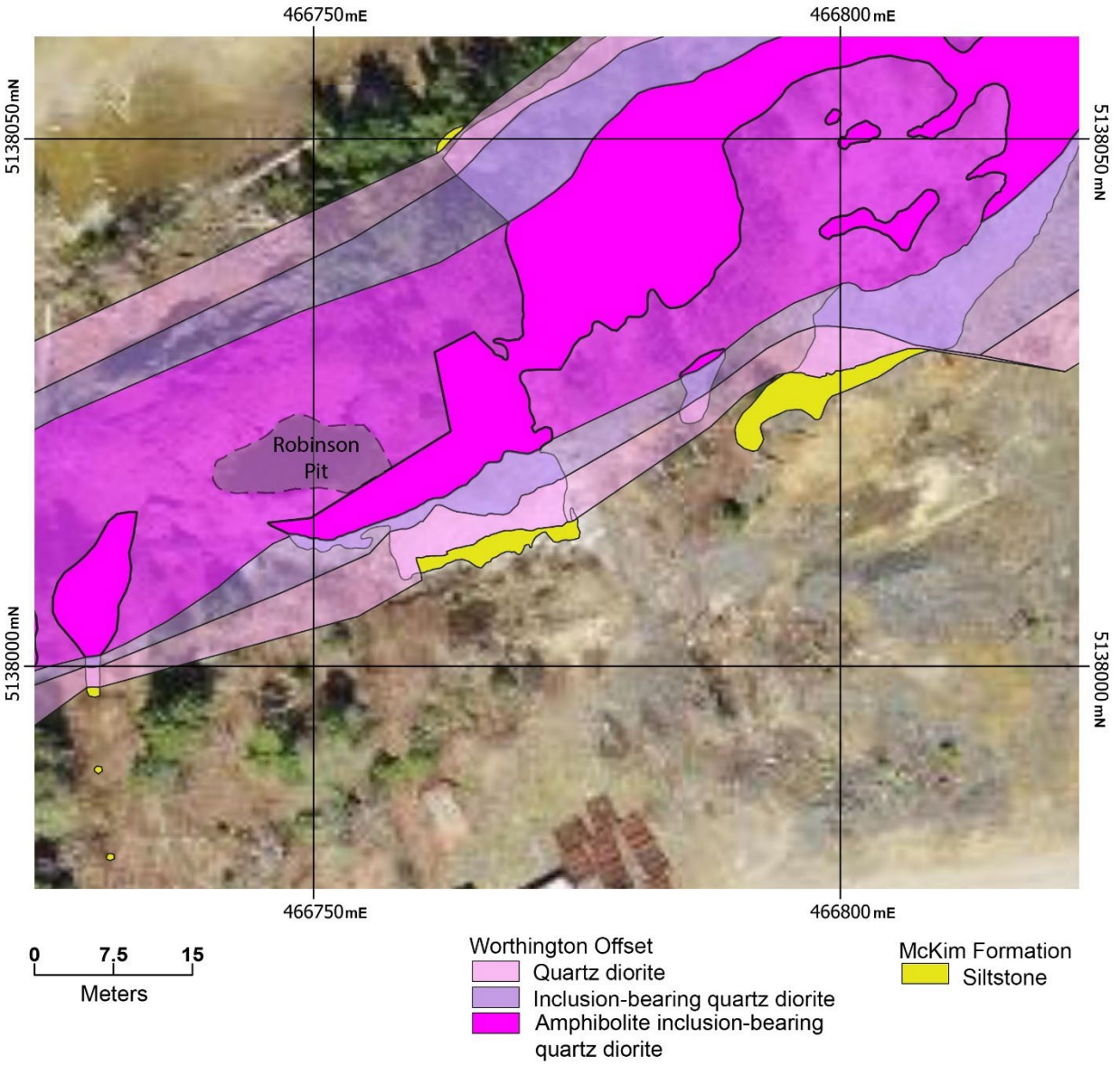


Figure 21. Geological map of the stripped outcrop found on the Robinson Mine site at Stop 7b. The Stop includes siltstones of the McKim Formation, and quartz diorite (QD), inclusion-bearing quartz diorite (IQD) and amphibole- and inclusion-bearing quart diorite (AIQD) phases of the Worthington Offset dike. Massive sulphide mineralization is hosted within the AIQD phase present along the northern edge of the exposure.

A sub-unit of IQD is found at this location, and locally is called amphibolite-bearing IQD (AIQD). The contact between IQD and AIQD is transitional, which is why the latter is considered a sub-unit of IQD rather than a distinct phase of the dike. AIQD contains almost exclusively large, rounded, amphibolite inclusions, with massive sulphides wrapping around them (Figure 22A). The amphibolite inclusions are dark green, massive, coarse-grained, and can range from a few centimetres to several metres in diameter. These inclusions are thought to have been derived from the nearby Nipissing sills. Similar inclusions have been reported at the Totten Mine, where they are referred to as "Sudbury Gabbros" (Lightfoot 2017).

Massive to semi-massive sulphide mineralization is associated with AIQD and occurs solely within the quartz diorite matrix. The amphibolite inclusions themselves are not mineralized, thus they are quite dilutive to the mineralization of this sub-unit. This style of mineralization is representative of the mineralization observed in the Worthington Offset dike. Chalcopyrite, pyrrhotite, and pentlandite are the dominant sulphide minerals (Figure 22B), and gersdorffite and niccolite have been identified locally in drill core. Sulphide mineralization is not a necessary feature of AIQD, but the latter is always present where mineralization occurs. The distribution of mineralized AIQD is complex and varied. On this property, 4 modeled vertical shoots are known to contain significant sulphide mineralization (Howland, Robinson, Rosen, and Perch Lake). Elsewhere along the Worthington Offset, mineralization is structurally controlled, occurring in bends, folds, and boudins along the dike.

Shear zones are observed throughout the Aer Kidd property, including at this outcrop, where an east-northeast-trending shear zone displaced the offset dike by less than a metre. Such localized displacement is observed along several other shear zones observed on surface and in drill core, and has been reported on many historical mine maps. The orientation of the shear zone is similar to other east-trending shear zones in Denison and Drury townships, including the CVDZ (Figure 3), and likely formed during the same deformation event.

A narrow diabase dike is also found at this stop. The dike crosscuts the Worthington Offset dike and has been tentatively assigned to the Trap dike swarm. It is massive, fine-grained, weakly magnetic and can be followed for several metres. Diabase dikes are commonly observed in drill core. Two large olivine diabase dikes of the Sudbury dike swarm also cut through the property.

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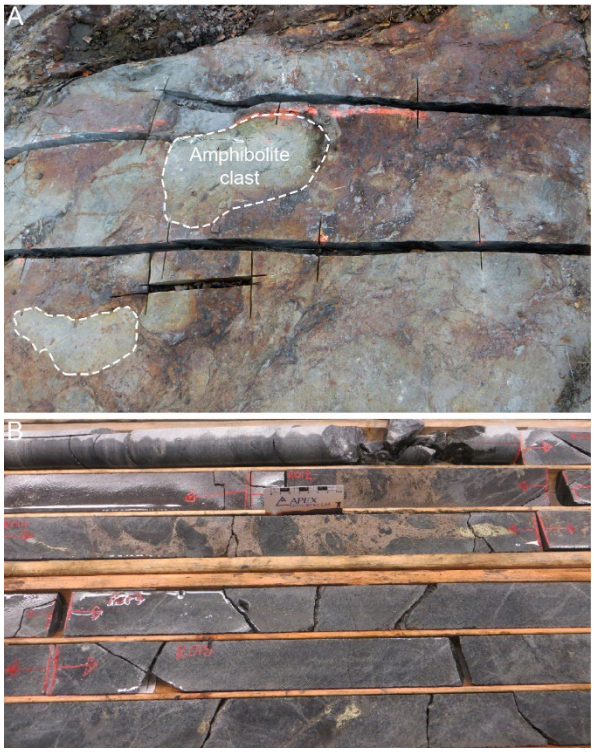


Figure 22. A) Mineralized outcrop on the surface exposure of the Robinson Mine, showing AIQD with amphibolite inclusions. The QD matrix between the inclusions hosts massive to semi-massive sulphide mineralization. **B)** Mineralized AIQD in drill core. Two massive sulphide stringers are hosted within quartz diorite matrix and wrap around rounded amphibolite inclusions. Note the smaller amphibolite inclusions the sulphide stringer vein. Dominant sulphide minerals are pyrrhotite and pentlandite, with lesser chalcopyrite (Stop 7b).

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Field Trip 4 – Overview of the Sudbury Structure

Sudbury District RGP Office

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Introduction

This field trip is based on the "Sudbury Structure Field Trips" that have been given for over 20 years by the Sudbury District Office of the Ontario Geological Survey, Resident Geologist Program. There are, therefore, numerous contributors, not all of which have been recorded over the years. The Rousell and Brown (2009) Field Guide is also heavily leaned upon.

Greater detail on the Sudbury Structure is presented in the introductory section of Field Trip 1 at this meeting "A Traverse Across the Sudbury Impact Structure" (Bleeker et al. 2022). Furthermore, Lightfoot (2017) gives a very in-depth treatment of the Sudbury Igneous Complex and its associated nickel deposits.

This one-day field trip will provide an overview of the Sudbury Structure, one of the most prolific nickel camps in the world and the remnant of one of the world's largest impact craters. The sites visited provide a cross-section of the Sudbury Structure, including the footwall rocks, the Sudbury Igneous Complex and the crater-fill sedimentary rocks of the Whitewater Group.

In the footwall rocks, the evidence for the impact origin of the structure include the Sudbury Breccia (pseudotachylite) and shatter cones. The nickel deposits formed at the base of the Sudbury Igneous Complex, and in the associated quartz-diorite "Offset" dikes. The first of the crater-fill units, the Sandcherry Member of the Onaping Formation, is a fallback breccia from the impact. The subsequent crater-fill units, the Onwatin and Chelmsford Formations, will also be visited.

The Stops are not in stratigraphic order, but in the order that provides the best safety for the participants (e.g., avoidance of crossing major highways whenever possible).

Geology of the Sudbury Area – Brief Overview

Formation of the Sudbury Structure

It is generally accepted that the Sudbury Structure and its associated abundant metal endowment were the result of a collision of a large meteorite (or comet; e.g. Petrus et al. 2015) with the Earth approximately 1850 million-years-ago (Krogh et al. 1984; Davis 2008). The impact happened near the contact of the Archean Superior and the Paleoproterozoic Southern provinces.

The entire meteorite, as well as a tremendous mass of earth's crust, was volatilized on impact. A complex crater (Figure 1) developed and was filled with a 'melt sheet' of molten crustal material that segregated and crystallized to become the Sudbury Igneous Complex (SIC); which was then overlain by fragmental "fall-back" material and younger, more typical, sedimentary rocks that constitute the Whitewater Group.

At the time of the impact the Penokean Orogeny (1870-1835 Ma) was underway. It is posited that the impact occurred in the foreland marine basin at the leading edge of the orogen. The Penokean Orogeny continued after the collision and resulted in modification to the impact crater. The subsequent nearly 2 billion years of tectonic and erosional history also changed the shape and size of the Sudbury Structure, resulting in its current configuration.

Geology of the Sudbury Structure

After the impact, the rocks and structure generated by the impact were subjected to varying degrees and orientations of deformation. The Sudbury Structure, therefore, has been divided into three distinct "ranges" in order to distinguish sections with different footwall rock affinities and

different degrees of structural over-print. Figure 2 outlines the approximate location of the "North Range", "South Range" and "East Range". The East Range is distinguished primarily on its distinct and complicated structural overprint.

A stratigraphic section (Figure 3A) summarizes the rocks and features that developed as a result of the impact. These features are found in the country "Footwall" rocks as well as in the SIC and the Whitewater Group. A second cross-section (Figure 3B) summarizes the various ore forming environments within the entire Sudbury Structure.

In the vicinity of the Sudbury Structure, the Superior Province rocks comprise the Levack Gneiss Complex; a suite of 2711 to 2642 million-year old metamorphosed and intimately intercallated supracrustal and felsic and mafic intrusive rocks (Krogh et al. 1984; Wodicka and Card 1995). The gneisses are cut by the meta-granitoid intrusive rocks of the Cartier batholith (2642 Ma; Meldrum et al. 1997). The Southern Province rocks are characterized by shallow marine and terrestrial metavolcanic and metasedimentary rocks of the Huronian Supergroup (2460 to 2300 Ma; *see* Trip 5, this volume; Easton and Bennett 2022) and associated mafic-ultramafic intrusions associated with the East Bull Lake intrusive suite (*circa* 2480 Ma; James et al. 2002) and the Nipissing gabbro suite (*circa* 2217 Ma; Davey et al. 2019). The character of the distinct Superior versus Southern province rocks is reflected in the Sudbury Structure rocks in proximity with either subprovince (i.e. the nature of Sudbury Breccia in the North Range (Superior Province) is distinct from that of the South Range (Southern Province)).

Outside the Sudbury Structure proper (in what is referred to as the "Footwall" environment), rock types that resulted from the Sudbury Event include the Sudbury Breccia, Footwall Breccia and Offset Dikes (or Quartz Diorite - QD). Rocks within the Sudbury Structure include the Sudbury Igneous Complex (SIC) and the overlying Whitewater Group supracrustal rocks.

SIC Footwall Rock Types

Sudbury Breccia (Stops 3 and 8)

Sudbury Breccia resulted from the shattering of the rocks. that were at or near the surface of the Earth. due to the shock from the impact of the meteorite. This network of in-situ breccias is relatively abundant in zones that occur as far as 15 km from the Sudbury Structure contact, whereas more isolated occurrences have been identified up to 80 km from the exposed footwall contact. Breccia contacts are typically sharp and have a range of configurations from straight and regular through complex, anastomosing, riverine patterns. Sudbury Breccia dikes preferentially conform to pre-Sudbury Event structures and lithological contacts. Breccia Zones range from millimetres to hundreds of metres wide and may be traced for tens of kilometres (e.g., the South Range Breccia Belt).

Sudbury Breccia is typically a clast-supported breccia with a pseudotachylitic (glassy) to finely comminuted matrix that has been variably recrystallized. In proximity with the SIC, the matrix displays partial melt in pods and encased lithic fragments. Clasts are generally locally derived, with some exotic material, and are commonly equant with sub-rounded to sub-angular shapes.

Economic mineralization can occur in proximity to, and associated, with Sudbury Breccia. The Frood–Stobie and Broken Hammer mines are examples of deposits hosted in, or associated, with Sudbury Breccia.

Footwall Breccia (Stop 5: Access requires permission from City of Greater Sudbury)

Footwall Breccia is an economically significant unit that hosts deposits such as at the Levack Mine (Figure 4). This style of mineralization is most prevalent in the North Range. Footwall Breccia is not uniformly distributed around the Sudbury Structure contact environment and is absent at the contact in many places. Unmineralized Footwall Breccia is present at Stop 5.

This enigmatic rock type occurs as "Megabreccia" and discontinuous sheets at the SIC contact, as well as part of some Offset Dikes (i.e. Whistle and Foy), and as intrusions into Felsic Norite. It is thickest in "embayments" (interpreted slump features along the over-steepened crater wall) where they provided suitable sites for sulphide accumulation (Figure 4A).

Footwall Breccia is a heterolithic breccia with angular to sub-rounded fragments of varied sizes. Fragments are mostly locally derived and may include gabbro, diabase, mafic gneiss, and Huronian Supergroup sandstones – dependent upon the rock types of the footwall environment. Fragments of Sudbury Breccia have been identified in this unit, constraining the relative age of these two breccia types. The matrix is what makes this such an enigmatic lithology; it is crystalline rather than fragmental in nature. The matrix is sub-igneous at the SIC contact and has a metamorphic character with increased distance from the SIC. High temperatures at the SIC contact melted the breccia matrix resulting in the sub-igneous texture (Lakomy 1990). With increased distance from the contact, the temperature gradient resulted in a transition from an igneous-textured breccia matrix to a metamorphic-textured breccia matrix (Fedorowich et al. 2009).

Sudbury Igneous Complex (SIC)

The Sudbury Igneous Complex (SIC) represents the crystalline rocks generated from the impact melt sheet. The current theory suggests that the melt sheet was created on the floor of the evolving impact crater within moments of the impact of the meteorite. This dynamic fluid evolved with the crater. As the crater stabilized the melt sheet differentiated and eventually solidified to form the SIC. The fall-back material (the Onaping Formation of the Whitewater Group) was originally ejected from the crater as it was excavating itself, but then was deposited on top of the melt sheet and was likely partially consumed and incorporated into it.

The SIC, from bottom to top (or from Footwall contact to the base of the Whitewater Group), includes the Contact Sublayer/Offset Sublayer and the Main Mass (Norite, Quartz Gabbro and Granophyre) units. Contacts between the units are gradational and subject to interpretation. The character of the units in different regions within the Sudbury Structure can be distinct.

Contact Sublayer/Offset Sublayer (Stops 1, 5 and 9)

Sublayer rock types are inclusion-bearing, igneous-textured, gabbronoritic or quartz dioritic bodies of varied compositions. The two main Sublayer environments are contact and offset dike. Both environments have potential to host economic Ni-Cu±PGE (Platinum Group Element) mineralization. The Discovery Outcrop (Stop 1) is an example of mineralized Contact Sublayer, and the recently re-opened Totten Mine is an example of economic mineralization in the Offset Sublayer environment.

In the contact environment, rocks of the Sublayer are discontinuous lenses or sheets at the contact between the "Main Mass" of the SIC and the Footwall rocks. The presence and thickness of the Contact Sublayer appears to be controlled by the three-dimensional topography of the Footwall contact, with the Sublayer preferentially occurring in troughs (embayments and terraces) at the contact. The Contact Sublayer is a varied mixture of igneous silicate matrix, inclusions of silicate rock material, and magmatic Cu-Ni-Fe sulphides. Fragments include

- footwall country rocks that can be identified and correlated with those directly observed in the Footwall,
- xenoliths related to the SIC (similar to the basal Norite unit) and
- exotic inclusions (anorthosite to dunite) with no known affiliation or source.

In the Offset environment (Stop 9), the Sublayer is commonly referred to as Quartz Diorite or QD. This type of Sublayer occurs in thin, dike-like intrusions into the Footwall that occur in three morphologies

- radial offsets that extend out nearly orthogonally from the SIC,
- concentric offsets that are oriented sub-parallel to but outboard of the SIC, and
- discontinuous or breccia-hosted offsets.

Most of the offset dikes are composite intrusions consisting of a central core of inclusion-bearing (commonly sulphide-rich) quartz diorite flanked by relatively inclusion- and sulphide-poor quartz diorite.

Main Mass (Stop 4)

The Main Mass of the SIC is a gradational series of igneous rocks that can be interpreted as having crystallized from a single differentiating magma (Pattison 2009; Lightfoot et al. 1997). At the base of the Main Mass, the Norite unit is a massive, medium- to coarse-grained, cumulate textured, two-pyroxene gabbronorite with varied amounts of quartz (up to 15%) and numerous accessory minerals. In the South Range, the unit is black as a result of ilmenite growth in plagioclase grains. In the North Range, the plagioclase grains do not typically contain ilmenite in their crystal structure and the rock colour is generally grey. "Mafic Norite" can locally be distinguished at the base of the Main Mass as modally more orthopyroxene-rich and quartz-poor than the rest of the Norite, and geochemically by a sharp increase of magnesium.

The intermediate unit of the SIC Main Mass is a greenish-grey, two-pyroxene Quartz Gabbro. Contacts between the Quartz Gabbro and the flanking Norite and Granophyre are gradational. Quartz contents range from 15-60%. Passing stratigraphically up through the Quartz Gabbro there is an increase in the abundance of granophyric-textured quartzofeldspathic material. At Stop 4 this can be observed as an increase in the pink component of the rock relative to the green component.

The upper and most felsic of the SIC Main Mass units is the Granophyre (formerly called micropegmatite). Granophyric intergrowth of quartz-plagioclase-potassium feldspar modally predominate in this unit and intact crystalline plagioclase laths and prisms are present.

Whitewater Group (Stops 2, 6 and 7)

The Whitewater Group encompasses an accumulation of nearly 3 km of supracrustal clastic and chemical "sedimentary" basin-fill rocks. The lowermost Onaping Formation is an atypical clastic rock of contentious origin; however, it is overlain by more recognisably sedimentary units that include the Vermilion, Onwatin and Chelmsford Formations. The Whitewater Group has been recognized only within the confines of the Sudbury Structure. However, it is likely that related clastic sediments (most particularly of the Onaping Formation) were deposited well outside the structure, but were either

- not lithified and were dispersed and/or redeposited in other forms or
- were eroded away or
- have not yet been recognized or
- a combination of the above scenarios.

Onaping Formation (Stop 6)

Originally described as a tuff, this contentious formation is a thick (1.4 km) accumulation of heterolithic breccias with igneous textures. The base of the formation is intruded by Granophyre of the underlying SIC and the top grades up into the carbonate and mudstone rocks of the overlying Vermilion and Onwatin Formations.

Recent detailed work by Ames et al. (2009) has led to the interpretation that the Onaping Formation formed in a dynamic and changing environment with an internal stratigraphy that captures the early evolution of the impact crater. "The formation represents a succession of glass-rich breccias and coeval hypabyssal intrusions that have been hydrothermally altered to a variable degree." (Ames et al. 2009) The most compelling evidence that distinguishes the rocks of the Onaping Formation from the volcaniclastic rocks that they texturally resemble is the presence of

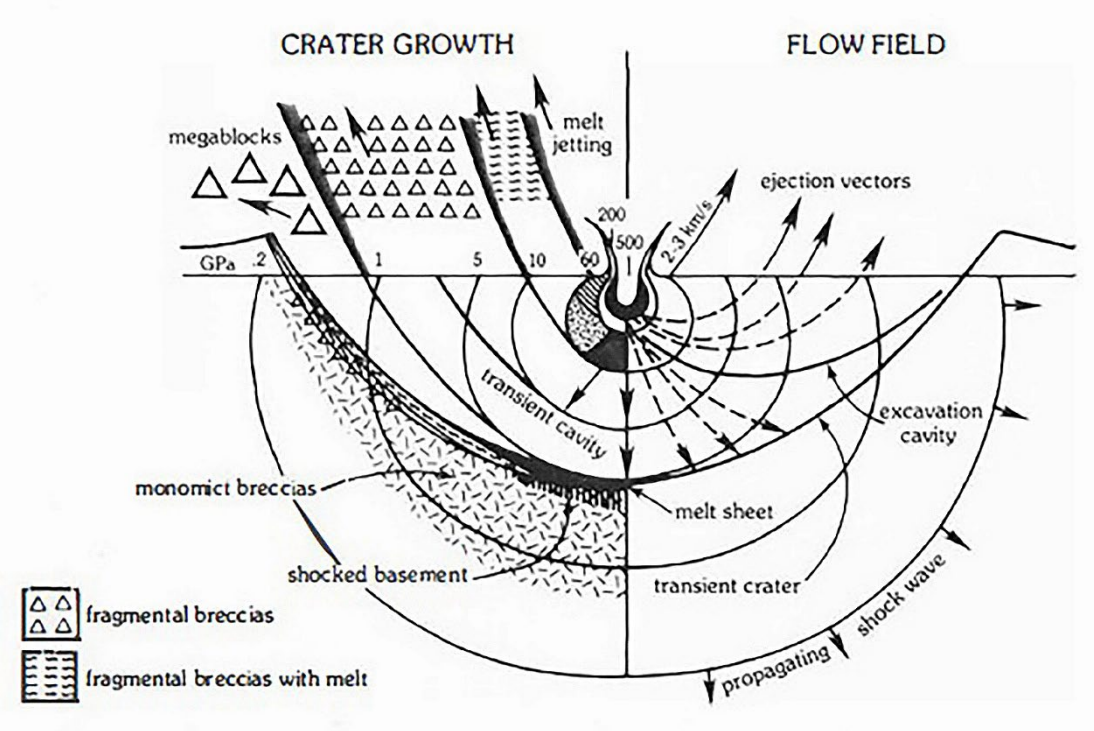


Figure 1. Complex crater development after a meteorite impact (*from* Stoffler et al. 1980 *in* Taylor 1982).

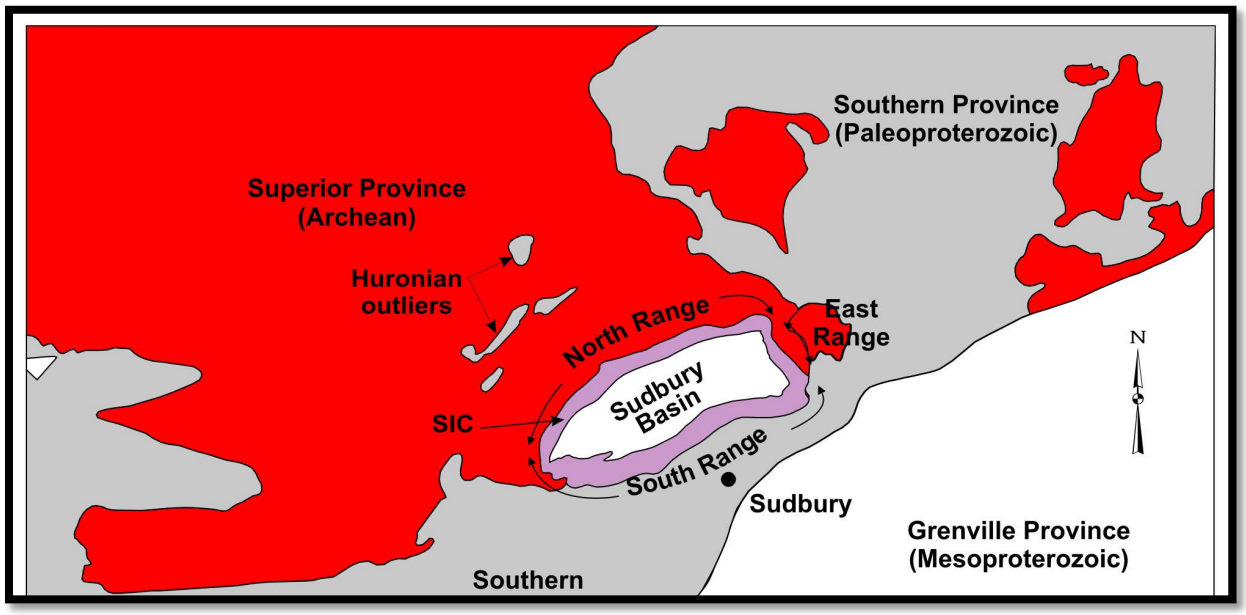


Figure 2. Sudbury Structure showing the surrounding geologic provinces and the 3 "ranges" of the SIC. Note the Grenville Province only formed in the area after the impact event (*from* Rousell and Card 2009).

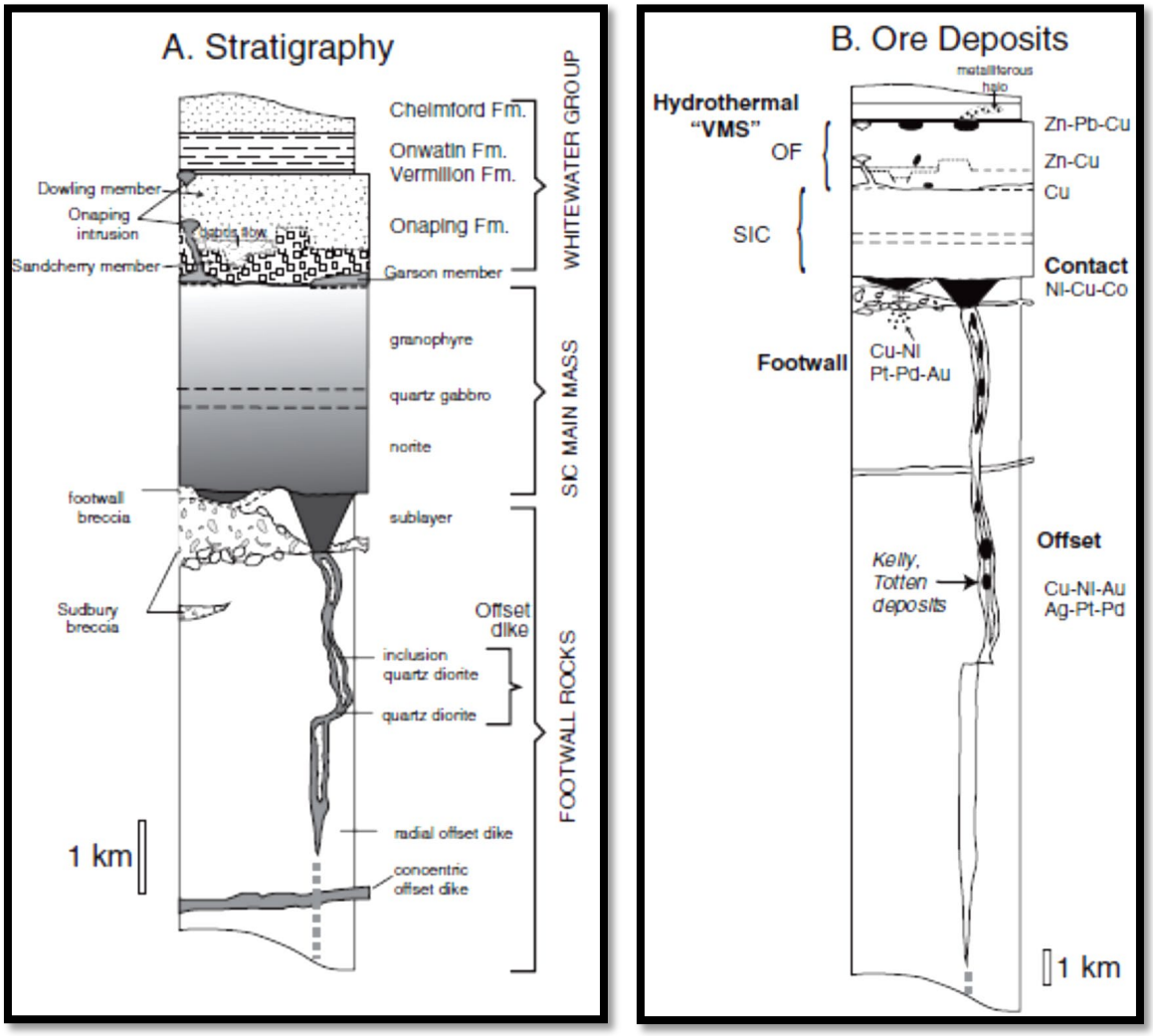


Figure 3. Schematic sections illustrating the **A)** stratigraphic and **B)** ore deposit environments associated with the Sudbury Structure. Note the difference in scale and the absence of the uppermost stratigraphy in B. SIC = Sudbury Igneous Complex, OF = Onaping Formation. Figure *from* Ames et al. (2008).

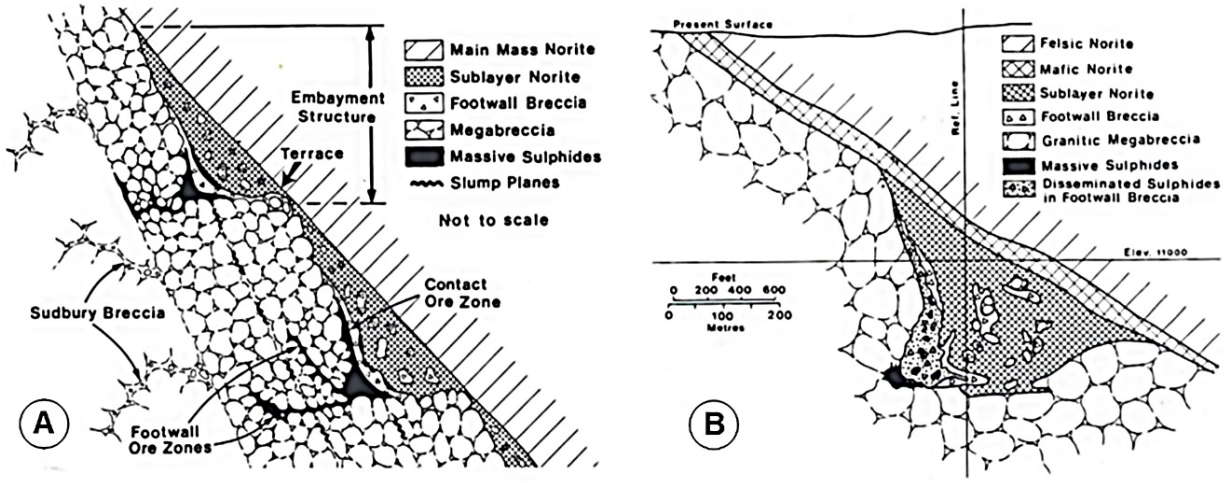


Figure 4. A) An idealized representation of the embayment structures and their relationship to slump terraces of the Sudbury Structure. **B)** A section through the Levack No. 4 ore body (*after* Morrison 1984).

shock metamorphosed quartz in the lithic fragments (French 1967; Peredery 1972). This is further supported by the presence of shock induced diamonds, and fullerenes (Becker et al. 1994; Masaitis et al. 1999), shatter cones in lithic fragments (Peredery 1972) and an iridium anomaly (Mungall et al. 2004).

At Stop 6, we encounter the chaotic fragmental rocks of the Sandcherry Member of the Onaping Formation. "Bombs" with lithic cores and well-developed, banded glass rims are distinctive features of these outcrops.

Vermilion Formation (no stop on this trip)

The Vermilion Formation is a carbonaceous argillite unit that was subjected to significant syn-depositional hydrothermal alteration. This alteration resulted in economic accumulations of zinc, lead, copper and silver mineralization (Errington and Vermilion mines). This unit, traceable around the Sudbury Structure in drill-core, is only exposed at surface in the southwest part of the basin in the vicinity of the past-producing Errington Mine. The average thickness of the unit is 13.5 m (Stoness 1994).

Onwatin Formation (Stop 7)

The Onwatin Formation consists of massive to laminated, carbonaceous and sulphidic argillite and siltstone with minor greywacke (Ames et al. 2009). This formation conformably overlies the Vermilion Formation. Its thickness has been interpreted as between 600 m and 1410 m. The formation is interpreted to have developed in a restricted, anoxic basin.

Chelmsford Formation (Stop 2)

The extensively exposed Chelmsford Formation is the uppermost preserved unit of the Whitewater Group and occupies the elliptical core of the Sudbury Structure. This unit comprises turbiditic greywackes with a minor argillic component (Ames et al. 2009). The contact between the Chelmsford Formation and the underlying Onwatin Formation is gradational. Well-defined sedimentary structures including channels, flute casts, ripples, convolute laminations and iron-carbonate concretions can be observed in outcrops of the Chelmsford Formation.

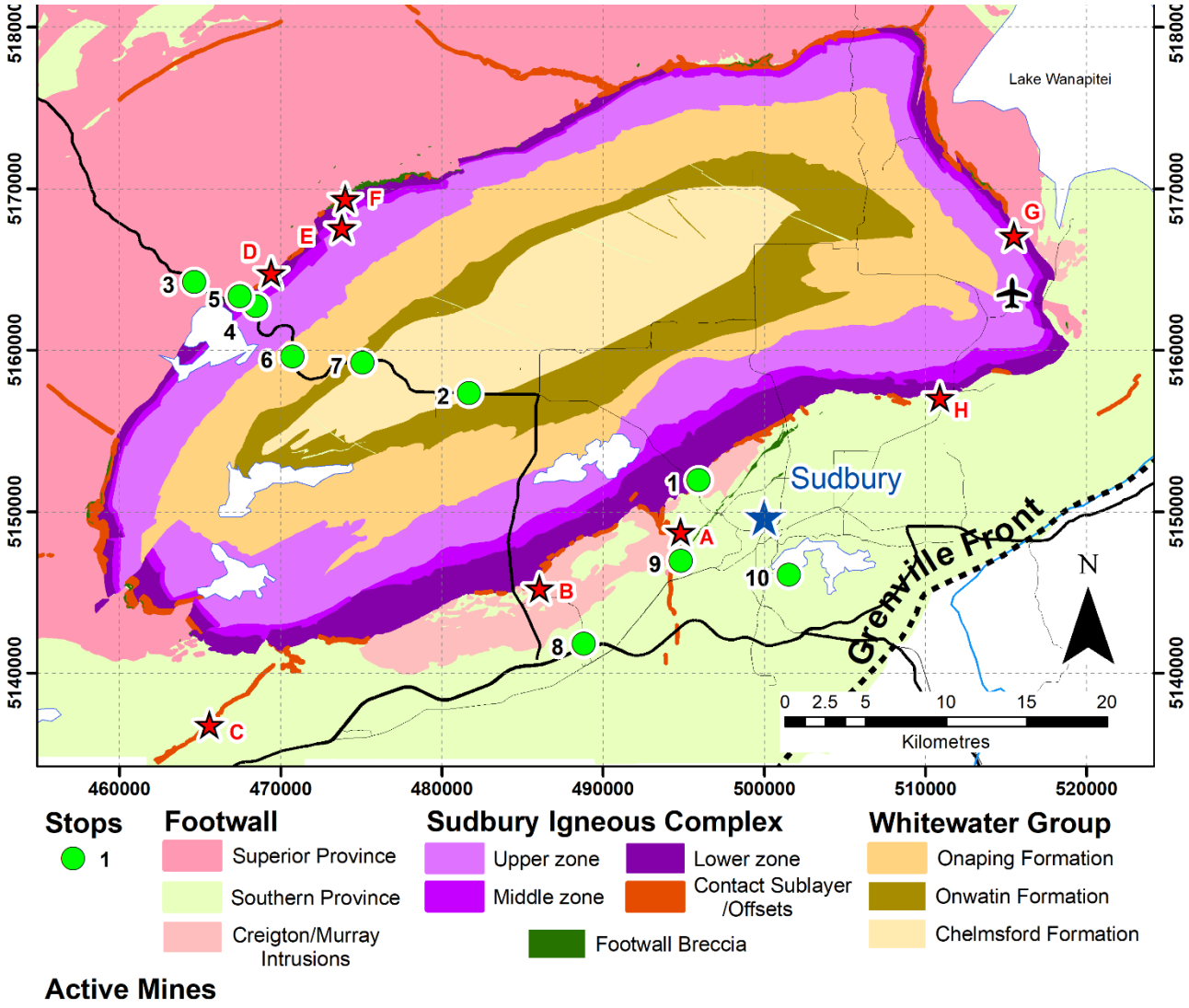


Figure 5. MAP OF FIELD TRIP STOPS (geology *from* Ames et al. 2005)

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Red stars indicate active mines

A Copper Cliff North Mine
B Creighton Mine
C Totten Mine
D McReedy West Mine
E Fraser Mine
F Coleman (Lower Coleman) Mine
G Nickel Rim South Mine
N Garson Mine

FIELD TRIP DETAILS

Geological Maps

Geological compilation maps covering all or parts of the area of the field trip include Ames et al. (2005), Dressler (1984b), and Card and Lumbers (1977).

ROAD LOG

Note: Caution should be taken when parking vehicles on the shoulder of roads and highways and when examining outcrops located along major roads along the field trip route. UTM co-ordinates are given in NAD 83 datum, zone 17; latitude and longitude are also provided in decimal degrees).

10.1 km — There are several routes to get to the Discovery Site from Science North. The "easiest" is: Turn left (west) out of Science North onto Ramsey Lake Road. Turn right (north) onto Paris Street (650m). Turn left onto Elm Street (2.75km) and continue on Elm Street/Regional Road 35 (6.7km) to turn off for Discovery Site on right (NE). Trail to site.

Stop 1 – Discovery Site

UTM co-ordinates 495917E, 5151952N latitude-longitude 46.5211243N, 81.0532319W

Protected site: NO HAMMERS

Caution: Outcrop is adjacent to the active transcontinental line of the Canadian Pacific Railway (CPR).

Nickel-copper mineralization was first identified in the Sudbury area by Alexander Murray of the Geological Survey of Canada in 1856. However, it wasn't until 1883 that the economic significance of the area was appreciated by the public. At this time construction of the Canadian Pacific Railway exposed rich copper-nickel mineralization at a site close to the spot commemorated here. In the 1970s the actual discovery outcrop was mined and is now occupied by the water-filled Murray Pit, a few hundred metres from here.



Stop 1. Sudbury Discovery Site: **A)** Gossaneous outcrop of Sublayer along strike from original Discovery Outcrop. **B)** Close-up of stringer sulphides. Ruler marked in inches on the top, centimeters on the bottom.

This discovery led to one of Canada's most active staking-rushes. In just over 100 years the Sudbury Structure has generated tremendous wealth and produced huge masses of critically important raw materials for Ontario's and the world's manufacturing.

At this site, in line with the original discovery outcrop, gabbro-peridotite inclusion-bearing Contact Sublayer is exposed. Mineralization, though perhaps not as rich, is similar to that found in 1883. The Clarabelle No. 2 open-pit and the head frame of the inactive Murray Mine are visible from the Stop.

16.1 km — Northeast on Regional Road 35. Turn-off on right (north). Outcrop 90m west of parking area on north side of highway.

Caution: Turn-off is a driveway.

Stop 2 – Chelmsford Formation, Whitewater Group

UTM co-ordinates 481688E, 5157337N latitude-longitude 46.5693493N, 81.2389532W

Well-defined Bouma sequences can be identified in these outcrops. These turbidite beds display excellent sedimentary structures indicating tops, including fining upward sequences, cross-bedding, flame structures and rip-ups. Distinctive, large, elongate "concretions" overprint bedding and can be traced within beds over many metres along the outcrop face. A well-developed, steeply dipping cleavage associated with the post-Sudbury event South Range Shear Zone (SRSZ) cuts bedding and appears to stretch the concretions into upright oval cross-sections.

23.5 km — Highway 144. Turn-off on right (NE). Outcrop 50m southeast of parking area on northeast side of the highway.

Caution: Outcrops adjacent to Highway 144.

Stop 3 – Levack Gneiss Complex with Matachewan dike and Pseudotachylite (Sudbury Breccia)

UTM co-ordinates 464642E, 5164239N latitude-longitude 46.6307803N, 81.4619083W

This location is part of what is considered the North Range footwall. It is outside the Sudbury Structure. Compositionally heterogeneous, complex gneiss of the Archean Levack gneiss complex is cut by pegmatitic dikes associated with the Archean Cartier Granite and a Paleoproterozoic Matachewan diabase dike. All these rocks are cut by narrow, impact-generated Sudbury Breccia veinlets. These outcrops display a complexity that is inherent in the footwall of the Sudbury Structure.



Stop 2. A) Bedded Chelmsford Formation with flame structures. **B)** 'Concretions' (outlined in dashed red line) in Chelmsford Formation.



Stop 3. Sudbury Breccia hosted in Levack gneiss. Ruler marked in inches on the top, centimeters on the bottom.

4.8 km — Return south on Highway 144 (4.7km). Turn left on onto Regional Road 8 (Onaping/Levack). Turn-off 160m from intersection on right. Outcrop 30m south of parking area; trail to outcrop.

Stop 4 – Quartz Gabbro and Granophyre

UTM co-ordinates 468456E, 5162737N latitude-longitude 46.6174539N, 81.4119820W

This outcrop shows the gradational contact between the Quartz Gabbro and the Granophyre units of the Sudbury Igneous Complex.

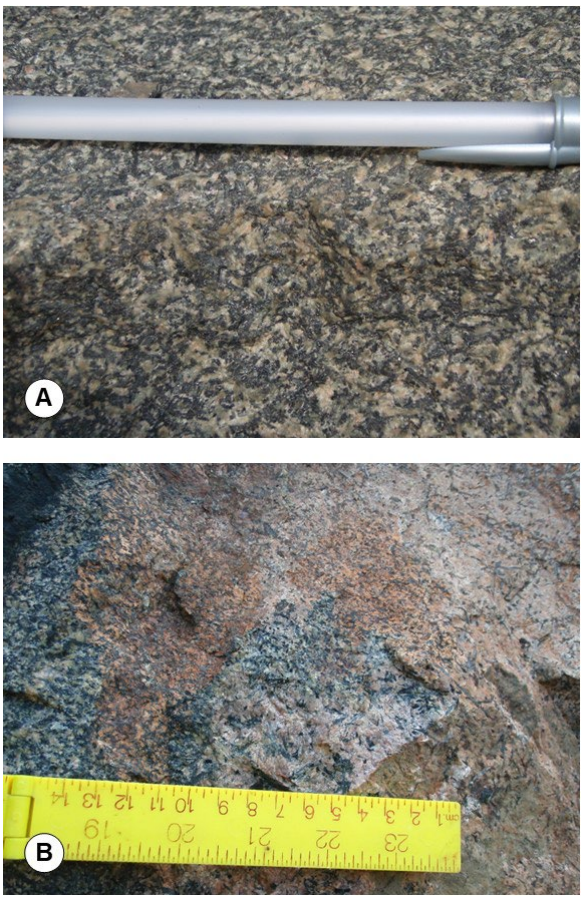
Rousell and Brown (2009) describe the granophyre on the southern part of the outcrop as being "approximately three parts micrographic intergrowth (potassium -feldspar and quartz) to one part tabular, plagioclase phenocrysts". The contact between the granophyre and the quartz-gabbro to the north is described as "a gradual change in the micrographic intergrowth to plagioclase ratio. The contact is arbitrarily placed where the modal plagioclase exceeds that of intergrowth".

1.9 km — North on Regional Road 8 (450m). Turn left on onto Onaping Drive (1.3km). Turn right onto unnamed road to gate for rehabilitated Onaping Landfill (150m). Outcrop is 250m north of parking area; walk on road to outcrop.

Stop 5 – SIC Contact: Levack gneiss, Sublayer and Footwall Breccia
(access requires permission from the City of Greater Sudbury)

UTM co-ordinates 467459E, 5163351N latitude-longitude 46.5892202N, 81.4298083W

This stop is at the west end of the highly productive Levack–Onaping embayment structure that hosts the currently operating Fraser, Coleman, and McCreedy West Cu-Ni-PGE mines (Figure 5). Two large outcrop faces expose Contact Sublayer rocks (western exposure) and Footwall Breccia rocks (eastern exposure). The differences between them are subtle and are primarily based on the fragment composition, and the presence or absence

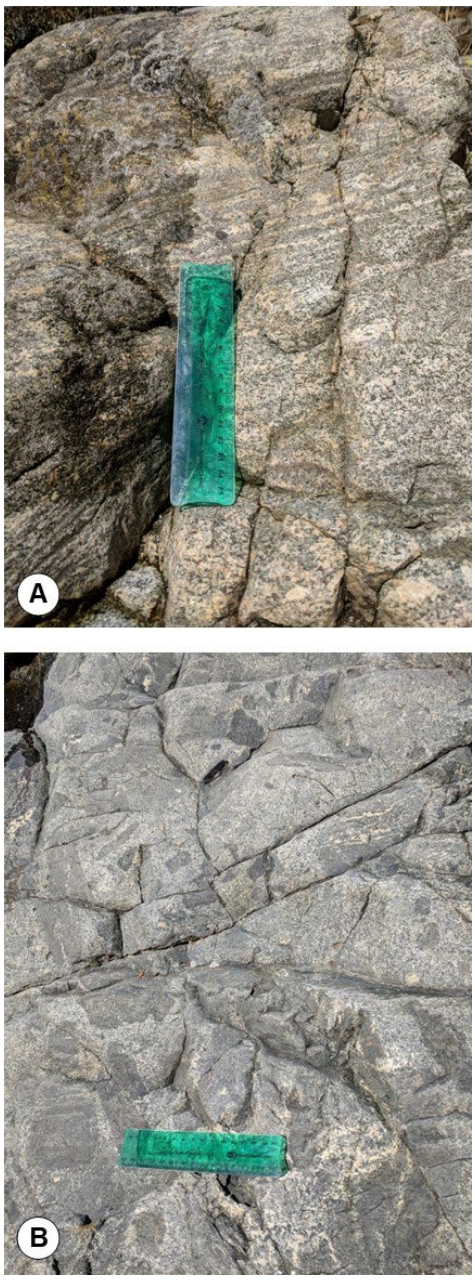


Stop 4. A) Sudbury Igneous Complex, quartz gabbro with pink felspars**. B)** micropegmatite (granophyre) dike.

of partial melt pods. Differences are best seen from a distance.

The outcrop on the west exemplifies barren Contact Sublayer. The exposure at the Discovery Site would look like this if it were not strongly mineralized. This rare, nearly barren outcrop of Sublayer shows fragments of norite > ultramafic >> felsic gneiss in an igneous crystalline matrix.

The outcrop on the east has been interpreted as unmineralized to weakly mineralized Footwall Breccia. Here, unlike the Sublayer outcrop, the fragments are all Levack gneiss. The Footwall Breccia is locally cut by zones of partial melting characterized by acicular amphibole crystals (that may originally have been pyroxenes) in a pink matrix.



Stop 5. A) Levack Gneiss Complex**. B)** Sublayer (dark norite breccia).

8.0 km — Return along unnamed road to Onaping Drive (150m). Turn left. Turn right (south) onto Regional Road 8 (1.3km). Turn left (south) on onto Highway 144 (500m). Turn left into A.Y. Jackson Lookout (6.2km). Outcrop is along west side of highway; 230m north of turn-off; 160m from parking area.

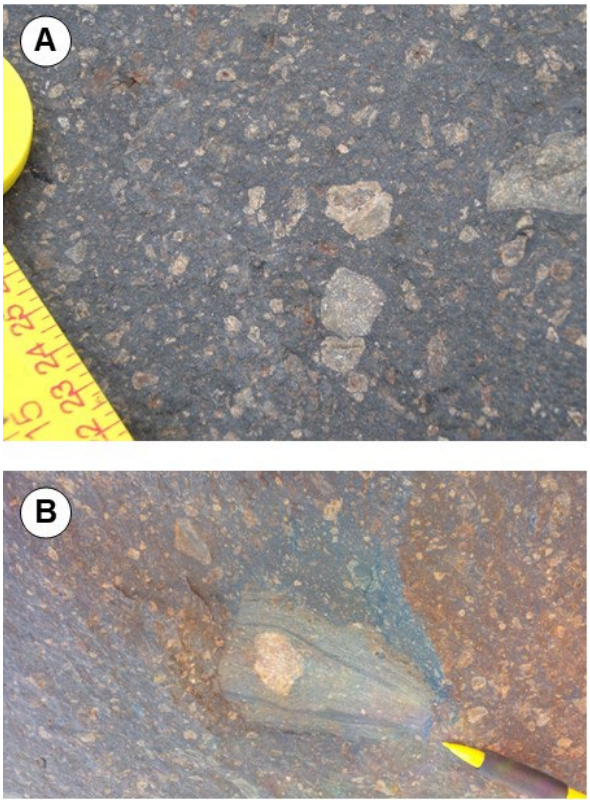
Stop 6 – Onaping Formation, Whitewater Group

UTM co-ordinates 470724E, 5159607N latitude-longitude 46.583893N, 81.3821632W

Please watch for highway traffic at this site!

Caution: Stop requires walking across Highway 144. Traffic can be very heavy and fast-moving.

This is the A.Y. Jackson Lookout, a scenic spot to enjoy High Falls. The outcrops to be examined are on the far side of Highway 144 and with the curves in the road and the speed of the traffic it is a dangerous crossing.



Stop 6. Breccia of the Onaping Formation, Sandcherry Member. **A)** General aspect. **B)** Fragment within a streamlined glass rim.

The Onaping Formation is the lowermost formation of the Whitewater Group. These rocks belong to the Sandcherry Member, a chaotic fragmental basin fill created from the fall-back of ejecta from the impact. Originally described as volcanic in origin, a few features identified in this

unit clearly distinguishes these rocks from the tuffs they closely resemble. The most compelling evidence is geochemical or microscopic: presence of an iridium anomaly (Mungall et al. 2004), presence of "impact diamonds" (Masaitis et al. 1999), and fragments containing shock metamorphosed quartz (French 1967; Peredery 1972). The planar deformation features in the quartz have only ever been identified in rocks affected by an impact and rocks affected by an atomic blast.

Look for lithic cored "bombs" and fragments with glassy rims. These rocks do look very much like volcaniclastic material.

5.4 km — Turn right (south) out of A.Y. Jackson Lookout onto Highway 144. Continue 5.4 km to stop on right (south). Park on the highway shoulder.

Stop 7 – Onwatin Formation, Whitewater Group

UTM co-ordinates 475298E, 5159222N latitude-longitude 46.5861084N, 81.3224356W

The Onwatin Formation is approximately 600 m thick, comprising carbonaceous and pyritic argillite and minor wacke. The formation is thought to have been deposited in a deep restricted basin with stagnant and anoxic bottom waters. Estimates of the range of total carbon content of the Onwatin Formation vary, and the carbonaceous material may have originated as floating algal mats (Rousell 1984; Arengi 1977). Coleman (1905) estimated a range of 6.8 to 10% carbon in the Onwatin slate, whereas Arengi (1977) calculated 0.26 to 4.05% free carbon in the Onwatin Formation. Arengi (1977) concluded that the carbon occurs as elongated segmented structures, either as individuals or in clots, which resemble modern and fossil algal and fungal filaments.



Stop 7. Onwatin Formation slate. Bedding is near horizontal (white dashed line); cleavage is vertical.

31.4 km — East on Highway 144 to Highway 144S (Lively/Highway 17; 11.7km). Turn right to stay on Highway 144S. Turn left onto Regional Road 24 (Lively; 13.6km). Turn right onto Anderson Drive (5.9km). Turn left into Tom Davies Community Centre (150m). NOTE: the entrance to the community centre is after the building (one-way). Outcrop is on the east side of the building next to the highway.

Stop 8 – Pseudotachylite: Sudbury Breccia

UTM co-ordinates 488791E, 5141810N latitude-longitude 46.4297697N, 81.1458923W

The glaciated outcrop shows fragment-rich pseudotachylite that is part of the South Range Breccia Belt. Fragments consist of lower Huronian Supergroup metasedimentary and metavolcanic rocks, which form the host rocks to the pseudotachylite dike. More specifically, the fragments consist of mafic metavolcanic rocks of the Elsie Mountain Formation, metarhyolite of the Copper Cliff Formation, and metapelite and metaquartzite of the McKim Formation. Fragments may show reaction rims and incipient marginal fragmentation. This suggests that the matrix was a melt.



Stop 8. Polymictic Sudbury Breccia–Pseudotachylite. (GPS is 15 x 6.5 cm).

10.0 km — Turn right onto Main Street from Anderson Drive (130m). Turn left onto Old Highway 17 (Regional Road 55 to Sudbury; 1.3km). Turn left onto Power Street (7.2km). Turn left onto Godfrey Drive (950m). Parking for stop on right (400m). Stop 85m east along powerline.



Stop 9. Copper Cliff No. 1 Mine site.

Stop 9 – Copper Cliff Offset: Copper Cliff No. 1 Mine site

UTM co-ordinates 494819E, 5146946N latitude-longitude 46.476050N, 81.0674912W

The Copper Cliff Offset is one of the Sudbury Offset Dikes (also referred to as Quartz Diorite Dikes or QD) that host much of the economic copper-nickel-PGM mineralization in the Sudbury Camp. The offset dikes are part of the noritic sublayer, and the Copper Cliff Offset merges with the Main Mass norite in a funnel-shaped embayment to the north. This segment of the offset is about 8 km in length. At this site, the dike lies along the contact between the supracrustal rocks of the Elsie Mountain and Stobie Formations to the east and the Creighton Granite to the west (Cochrane 1984). The dike is cored by the subunit "IQD" or Inclusion Quartz Diorite. This subunit is the host of economic mineralization in the offset environment.

The Copper Cliff No. 1 Mine was the first underground mine in Sudbury. Production began in 1886. The Copper Cliff North Mine continues production on the same Offset.

Of other historical interest, the park on the other side of Godfrey Street was a roast bed. The area has been re-greened.

9.5 km — Return south on Godfrey Drive and turn left onto Balsam Street (130m). Turn left onto Regional Road 55 (1.3km). Continue on Regional Road 55, keeping left at cloverleaf for Big Nickel Drive (Regional Road 34). Regional Road 55 becomes Lorne Street after Big Nickel Drive cloverleaf. Turn left onto Martindale Road (3.6km). Left turn to remain on Martindale (450m). Slight left onto Walford Road at Regent Street intersection (1.4km). Turn left onto Paris Street (750m). Turn right onto Ramsey Lake Road (700m), Stop on right (1.5km). Outcrop begins 50m back (SW) from parking spot along footpath.



Stop 10. Shattercones. Pen is 13 cm long.

Stop 10– Shatter Cones

UTM co-ordinates 501535E, 5146083N latitude-longitude 46.4683165N, 80.9800069W

Protected site: NO HAMMERS

This outcrop has some of the best preserved and abundant shatter cones found in the Sudbury area. The host rock here is quartzite of the Mississagi Formation. The shatter cones appear as conical striated features whose surfaces are often micaceous and shiny. They range in length from a few centimetres to about a metre. Large cones may have numerous small cones along their flanks. Cones are exposed only where they control the outcrop surface. On other surfaces intersecting crescent-shaped fractures give the characteristic shattered appearance to the rock. They are best seen here when obliquely illuminated by the late afternoon sun.

The distinctive fractures, termed shatter cones, form by passage of shock waves through rock, and are found at many astroblemes and "crypto-explosion" structures. They have also been reported from localities with no known explosive associations. Geological mapping has shown that the SIC is surrounded by a belt of rocks more than 16 km wide containing shatter cones. In some locations, if the rocks are returned to their hypothetical orientation during the Sudbury Event, the apices of the shatter cones appear to point inward toward the basin (French 1972). Other workers, however, have noted that at other locations around the SIC, shatter cone orientations are in opposing directions (Dressler 1984a).

The shatter cones, together with planar features in quartz, are believed to be evidence of a meteorite impact origin for the Sudbury Structure (e.g., Dietz 1964, 1972; French 1972).

End of road log.

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Field Trip 5 – An Overview of the Huronian Supergroup in the Elliot Lake area

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Introduction

The field trip uses road accessible outcrops. All of the road stops can be accessed using a 2-wheel drive vehicle. Unless otherwise stated, all UTM co-ordinates are in Zone 17, datum NAD 83, which is essentially equivalent to NAD WGS84.

Safety

Many of the field trip stops are located on highways that are especially busy during the summer season. Care should always be exercised when parking, exiting vehicles, and crossing the roads. Use of safety vests and/or bright clothing is recommended, in order to improve your visibility to motorists.

Most of the trip routes are on Crown land or public roadways, but access is on or near private property in some cases. As in all such situations, please respect the property rights of others, so as to maintain good relationships, so that future access for geologists is not adversely affected.

Purpose

The transect through the Huronian Supergroup at Elliot Lake used by this field trip provides an opportunity to examine nearly all of the major units of the supergroup in a single day.

These first pages are intended to give participants new to the Huronian Supergroup of Ontario a summary of what we think we know of these ancient rocks. Much of the material is borrowed from prior ILSG guidebooks (Bennett 2006; Bennett et al. 1997), but with updating and the inclusion of additional stops by R.M. Easton.

The Huronian Supergroup is one of the Earth's most studied sequences of rocks. Since the turn of the century the results of hundreds of studies of Huronian rocks have been published in scientific journals and government publications. These studies have led geoscientists, to present evidence for the Earth's earliest glacial periods, the development of free oxygen in the atmosphere of the early Earth, the deposition of paleoplacer deposits of uranium, and evidence for plate tectonic activity during the Paleoproterozoic. Much of the evidence is based on rock exposures which will be visited during this field trip.

Proterozoic rocks of the Canadian Shield in the Sudbury to Elliot Lake area are assigned to either the Paleoproterozoic Southern Province or the Mesoproterozoic Grenville Province (cf. Card et al. 1972; Wynne-Edwards 1972). The Southern Province in Ontario comprises Paleoproterozoic metasedimentary and metavolcanic rocks of the Huronian Supergroup and gabbroic intrusions of the Nipissing gabbro suite. Also included in the Southern Province are the Sudbury Igneous Complex (SIC), the Whitewater Group; plutonic and minor volcanic rocks of the Killarney Magmatic Belt; and rocks of the Sudbury mafic dike swarm (*see* Figure 2; Bennett et al. 1991). Table 1 summarizes the major geological events affecting the Superior, Southern and Grenville provinces in the Sault Ste. Marie to Sudbury area.

The Huronian Supergroup (Robertson et al. 1969a; *see* Figure 3) is a sequence of variably metamorphosed Paleoproterozoic sedimentary and minor volcanic rocks that lie unconformably upon

Archean rocks of the Superior Province. The Huronian rocks extend eastward from Lake Superior, along the north shore of Lake Huron to Sudbury, and then northward to the Noranda area of Quebec; a distance of approximately 450 km (Figures 1, 2).

The Huronian Supergroup attains its greatest thickness of 12,000 metres southeast of Sudbury (Debicki 1990). The sequence thins northward due to the wedging out of basal units, the thinning of the siliciclastic units, and erosion within the sequence (Roscoe 1969; Frarey and Roscoe 1970).

The *circa* 2310 Ma U/Pb age from zircons in tuffaceous layers in the Bar River Formation (Hill et al. 2018; Rasmussen et al. 2013) places an upper age limit on the age of deposition of the bulk of Huronian Supergroup, with all sedimentation being completed well before emplacement of the Nipissing gabbroic intrusions at *circa* 2217 Ma (Davey et al. 2019; Corfu and Andrews 1986; Noble and Lightfoot 1992). The age of the rhyolites of the Copper Cliff Formation (*circa* 2450 Ma; Bleeker et al. 2015; Ketchum et al. 2013) is probably close to the start of initial deposition of the Huronian Supergroup.

The Archean Basement

The basement to the Huronian Supergroup consists predominantly of rocks of the Ramsey–Algoma granitoid complex (Card 1979), which includes several large felsic batholiths, including the Cartier (2642 Ma: Meldrum et al. 1997) and Birch Lake (2651 Ma: Kamo 2006) granite batholiths. The batholiths were emplaced into a slightly older, granodiorite and quartz diorite intrusive and gneissic complex, which have ages of 2700 to 2675 Ma (Easton 2013a; Prevec 1993; Ontario Geological Survey, unpublished data).

These intrusive phases are younger than the few greenstone belts present in the Sudbury to Elliot Lake area; the largest of which is the Whiskey Lake greenstone belt, located south and southeast of Elliot Lake. Limited geochronology from the Whiskey Lake greenstone belt indicates 2 main periods of volcanism, an earlier, tholeiitic mafic dominated phase at 2724.9±1.4 Ma (Optional Stop C) and a younger, calc-alkaline phase at 2686.5±1.1 Ma (Easton 2010, 2013a). Volcanism had ceased by 2674.8±0.8 Ma, the emplacement age of a granodiorite intrusion into the greenstone belt (Stop 1).

The ages of both the greenstone belts and the Ramsey-Algoma granitoid complex are predominantly in the range of 2695 to 2650 Ma, in contrast to ages from the Abitibi greenstone belt which are typically in the range of 2740 to 2690 Ma. This distinction is important, as it allows for discrimination between potential source regions for the Huronian Supergroup strata, as described in detail in the section on "What Detrital Zircon studies tell us about the Source Region for the Huronian Supergroup".

Huronian Magmatism

Introduction

Four distinct, more-or-less coeval (2480 to 2460 Ma), post-Kenoran igneous rock sequences are associated with the Huronian Supergroup:

- Mafic dikes in the basement rocks but which do not cut the Huronian Supergroup (Matachewan and Hearst dike swarms).
- Igneous complexes, typically layered, of gabbro, gabbronorite and anorthosite (East Bull Lake intrusive suite).
- Mafic to felsic volcanic flows (Elliot Lake Group).
- Felsic plutons in the Southern and Grenville Provinces in the Sudbury area (2475 to 2460 Ma).

Basement Dikes

The granitoid rocks of the Ramsey-Algoma granitoid complex are intruded by mafic dikes of the Matachewan–Hearst swarm, which was likely emplaced in 2 main pulses, the first, earlier pulse at *circa* 2480 Ma is believed to have been coincident with emplacement of the East Bull Lake intrusive suite of layered intrusions (Krogh, et al. 1984; James et al. 2002a; Bleeker et al. 2015; Heaman 1997). The second and "main pulse" of the Matachewan–Hearst dike swarm occurred at *circa* 2460 Ma (Bleeker et al. 2015).

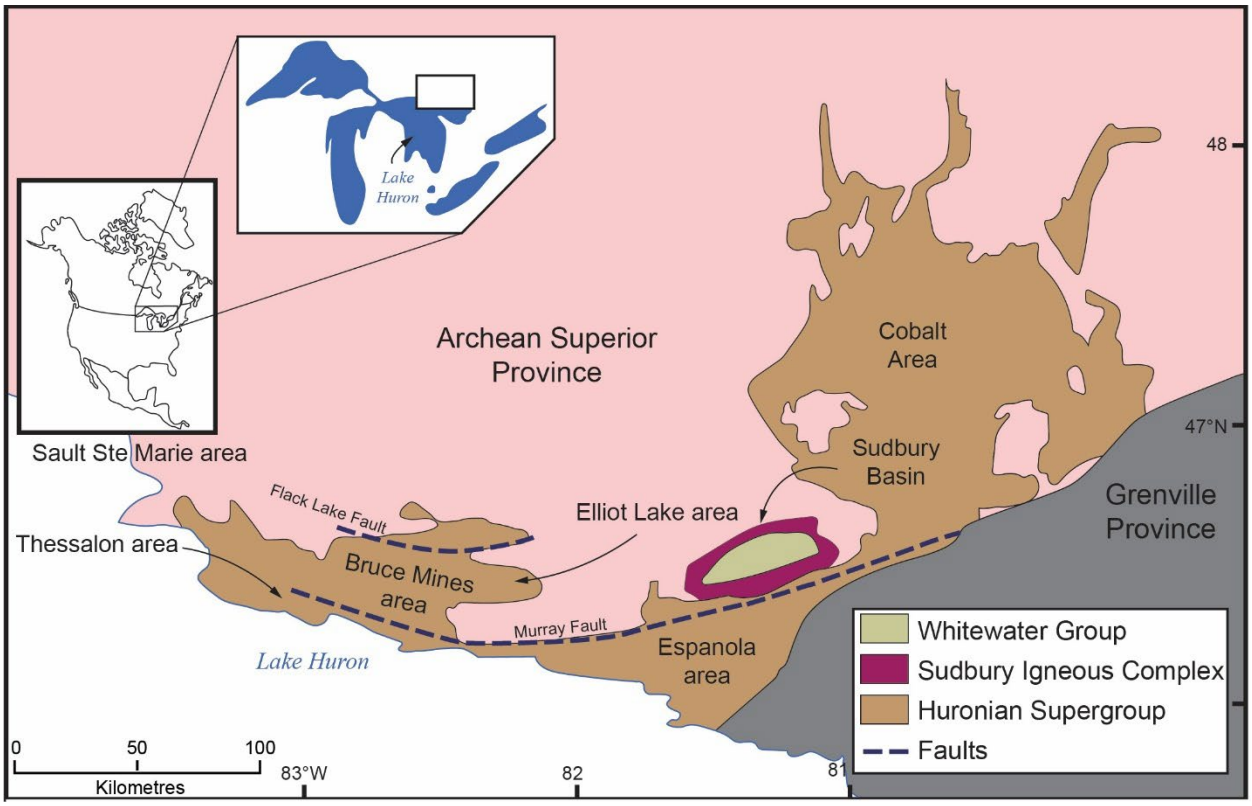


Figure 1. Sketch map showing the regional setting of the Sudbury Igneous Complex and the Huronian Supergroup (*modified from* Young et al. 2001).

Layered Gabbro, Gabbronorite and Anorthosite Complexes

At the base of the Huronian Supergroup in the Elliot Lake, Agnew Lake and Sudbury areas are several layered gabbro to anorthosite intrusions referred to as the East Bull Lake intrusive suite (Peck et al. 1995; James et al. 2002a, 2002b). These bodies have ages of *circa* 2475 Ma (Clough and Hamilton 2017; Krogh et al. 1984) and appear to be slightly older than the rocks of the Elliot Lake Group.

Intrusions of the East Bull Lake intrusive suite are characterized by the presence of anorthositic phases, and locally by a well-developed, primary rhythmic layering of alternating anorthositic and gabbroic layers (cf. James et al. 2002a, 2002b; Easton et al. 2010). Major intrusions of the East Bull Lake intrusive suite occur at Agnew Lake and East Bull Lake, with the largest body, the River Valley intrusion being found in the Grenville Province adjacent to the Grenville Front (cf. James et al. 2002a, 2002b; Easton et al. 2010). The three largest bodies contain platinum-group element mineralization near their basal contacts (cf. Easton et al. 2010).

U/Pb chemical-abraded thermal-ionization-mass-spectrometry (CA-TIMS) zircon ages from the River Valley, Agnew and East Bull Lake intrusions cluster at 2475 Ma (Clough and Hamilton 2017; Easton et al. 2010). Although similar in age to the older phase of the Matachewan dike swarm, there are numerous occurrences of Matachewan dikes cutting intrusive rocks of the East Bull Lake intrusive suite (cf. Easton 2003, 2009; Easton et al. 2010) indicating that emplacement of all these mafic rocks was coeval. All the East Bull Lake intrusive suite bodies found to date have been emplaced into the Archean basement at, or just below, the Archean–Huronian Supergroup boundary.

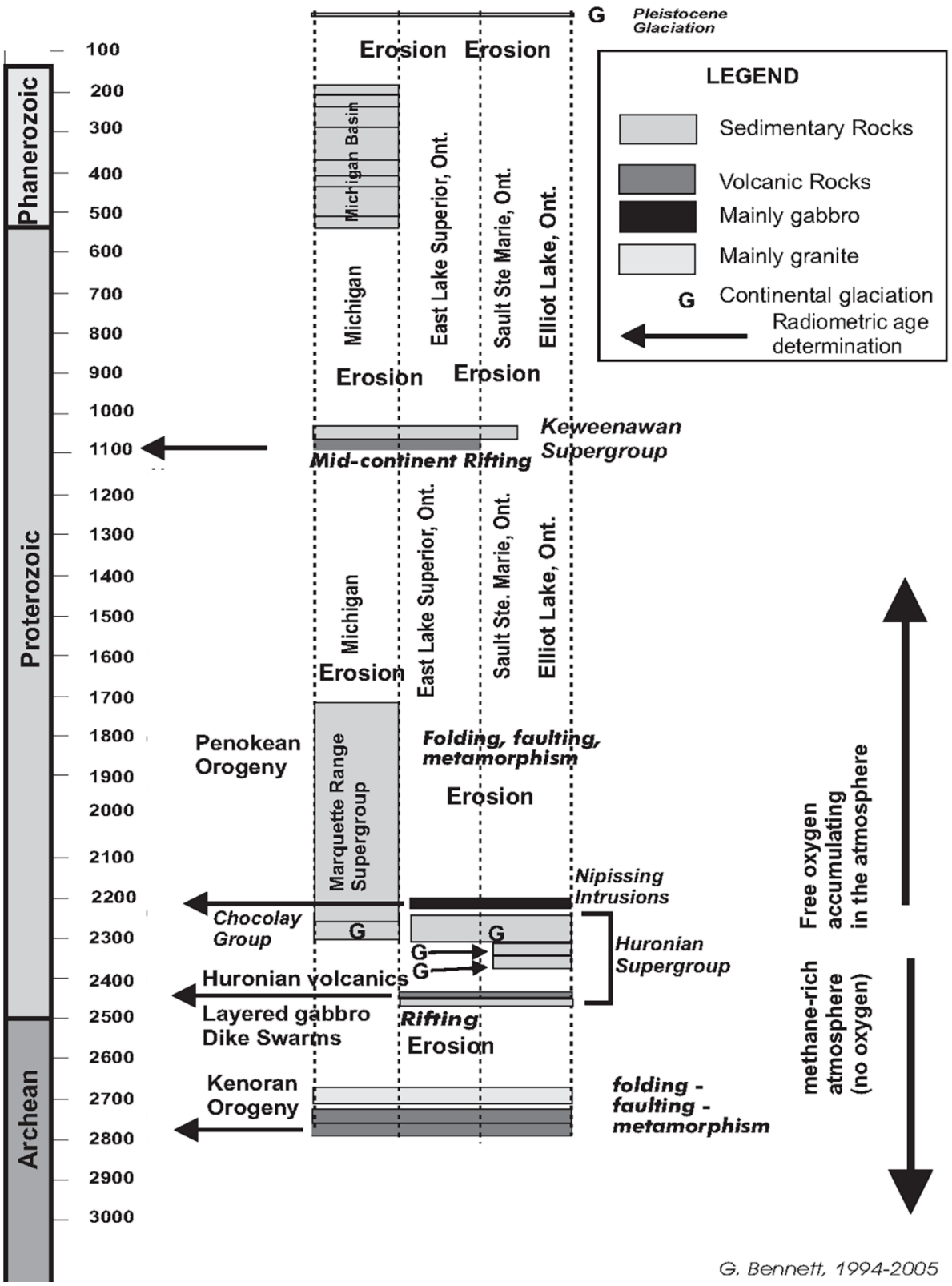


Figure 2. A time-rock chart for the southeast Lake Superior region (*from* Bennett 2006).

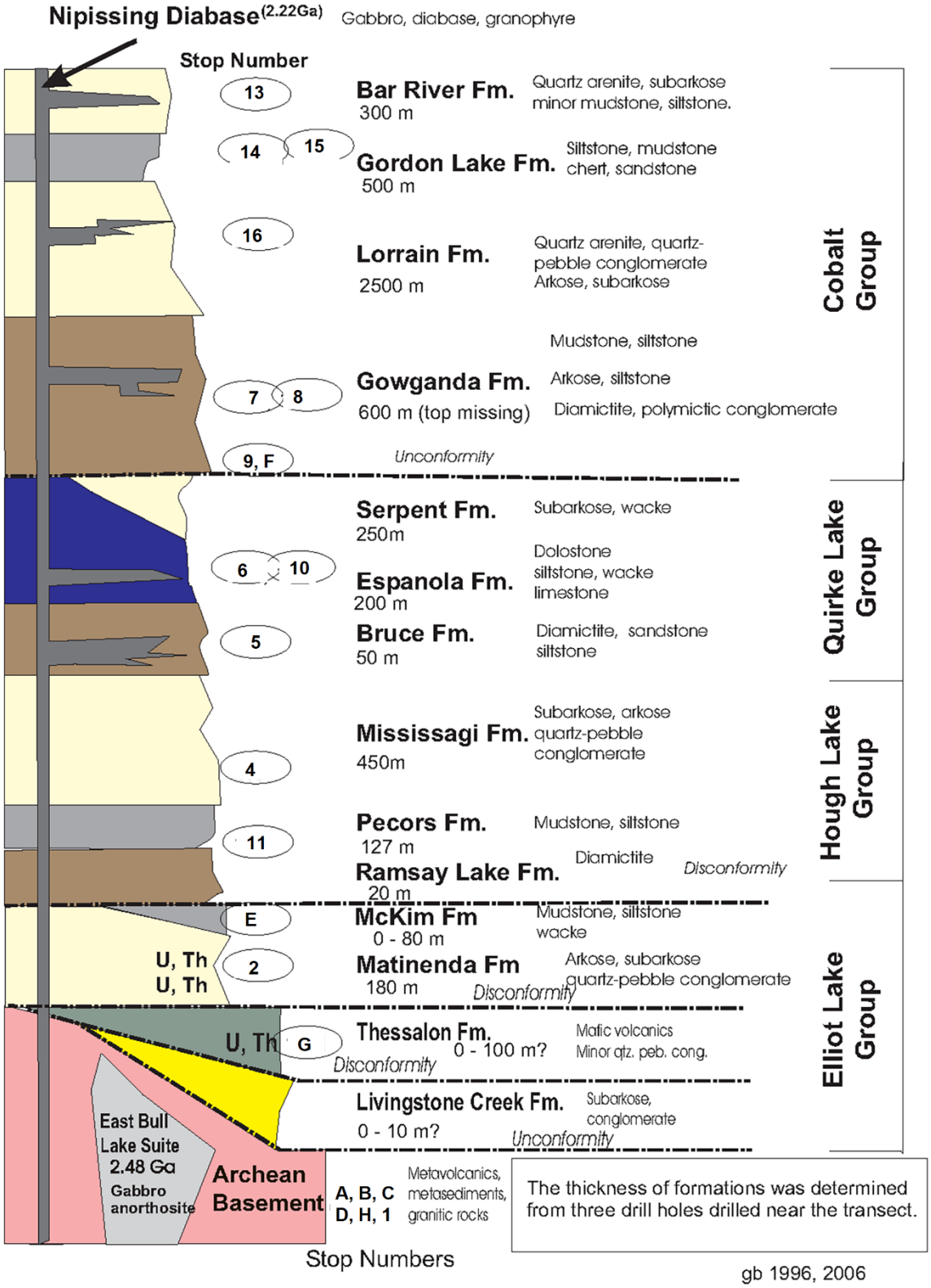


Figure 3. A stratigraphic column for the Elliot Lake fieldtrip transect (*from* Bennett 2006).

Table 1. Timing of major geological events and summary of age constraints on the main rock units present in the Sudbury to Elliot Lake area.

Event and/or Map Unit	Age Constraint (Ma)	Comment and/or Source
Grenville dike swarm	586±4	Kamo, Krogh and Kumarapeli (1995)
Pegmatite vein emplacement	989±2	Corfu and Easton (2000)
Age of peak metamorphism in the hanging-wall of the Grenville Front tectonic zone	1000 to 990	Corfu and Easton (2000)
Age of peak Grenvillian metamorphism in the Central Gneiss Belt	1040 to 1030	Carr et al. (2000)
Sudbury mafic dike swarm	1238±4	emplaced in or along northwest-trending faults in the Southern Province, deformed and metamorphosed in the Grenville Province. Krogh et al. (1987).
Killarney magmatic belt second-stage magmatism, coincident with magmatism in the Eastern Granite Rhyolite Province and in the Central Gneiss Belt	1471±3	van Breemen and Davidson (1988)
Regional albitization metasomatic event	1701±4	U/Pb monazite, Schandl, Gorton and Davis (1994); fluid focussed along northwest faults
Killarney magmatic belt volcanism and high-level plutonism	1740, 1747±3, 1749±12	van Breemen and Davidson (1988); Sullivan and Davidson (1993); Davidson and van Breemen (1994)
Northwest-trending regional faults	Pre-1700, post-1850	Faults cut Sudbury Structure
Penokean orogeny (folding and metamorphism of Huronian Supergroup rocks)	1775±10 ~1835	Peak deformation. Zi et al. (2022) Peak metamorphism. Holm et al. (2001)
Impact event and formation of Sudbury breccia	1850±1	Krogh, Davis and Corfu (1984); Davis (2008)
Penokean arc formation and magmatism	1890-1860, 1845-1830	Zi et al. (2022)
Thrust faulting	post-F_2 pre-regional faulting	Sudbury breccia localized along these faults, suggesting they are pre-Sudbury Structure
F_2 folding	post-2200, pre-1700, pre 1850?	Pre-regional faulting, Nipissing sills axial planar to folds
F_1 folding	pre-2200	Nipissing sills folded of intruded along folds
Emplacement of Nipissing gabbro sills	2217±4	Davey et al. (2019); Corfu and Andrews (1986); Noble and Lightfoot (1992)
Huronian Supergroup sedimentation	>2220 but <2460	Youngest detrital grains in Bar River Fm are 2306 Ma (Hill, Davis and Cochran 2018)
Huronian Supergroup felsic volcanism and related plutonic rocks, including the Matachewan dike swarm	~2477 to 2375 (2450±25, 2460±20, 2477±9, 2415±5	Krogh, Davis and Corfu (1984), Heaman (1997); Corfu and Easton (2000), Krogh, Kamo and Bohor (1996), Smith (2002); Bleeker et al. (2015)
Emplacement of East Bull Lake intrusive suite rocks	2475±2	Heaman (geochronologist, University of Alberta, personal communication, 1999); Clough and Hamilton (2017)
Emplacement of orthopyroxene hornblendite bodies (East Bull Lake suite)	2468±5	Corfu and Easton (2000)
Emplacement of alkali feldspar granite and megacrystic granodiorite near River Valley	2660 to 2665	Bodies intrude Crerar and Pardo gneiss, Easton (2003)
High-grade Archean metamorphism and migmatization	2647±4	Krogh, Davis and Corfu (1984); Wodicka and Card (1995); Ames et al. (2005)
Emplacement ages of Archean units in the Sudbury area	2711±7 to 2642±1	Krogh, Davis and Corfu (1984); Wodicka and Card (1995); Chen, Krogh and Lumbers (1995); Meldrum et al. (1997); Ames et al. (2005)

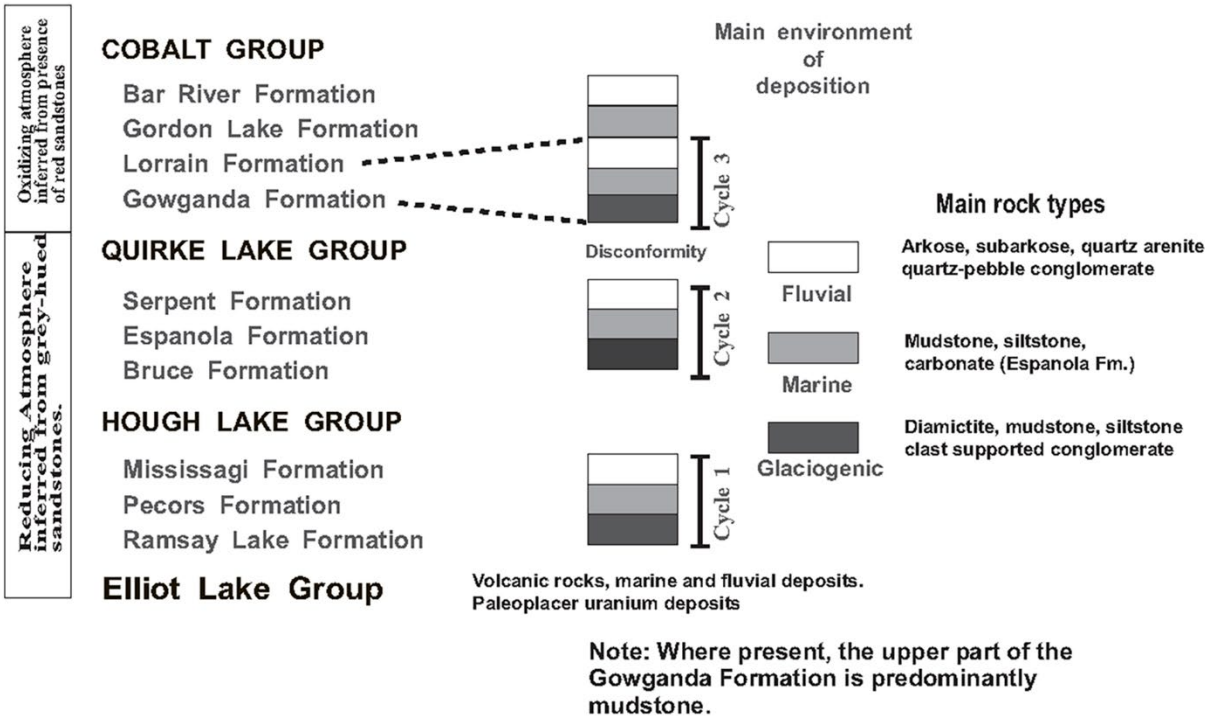


Figure 4. The cyclicity of Huronian Supergroup rocks (*from* Bennett 2006).

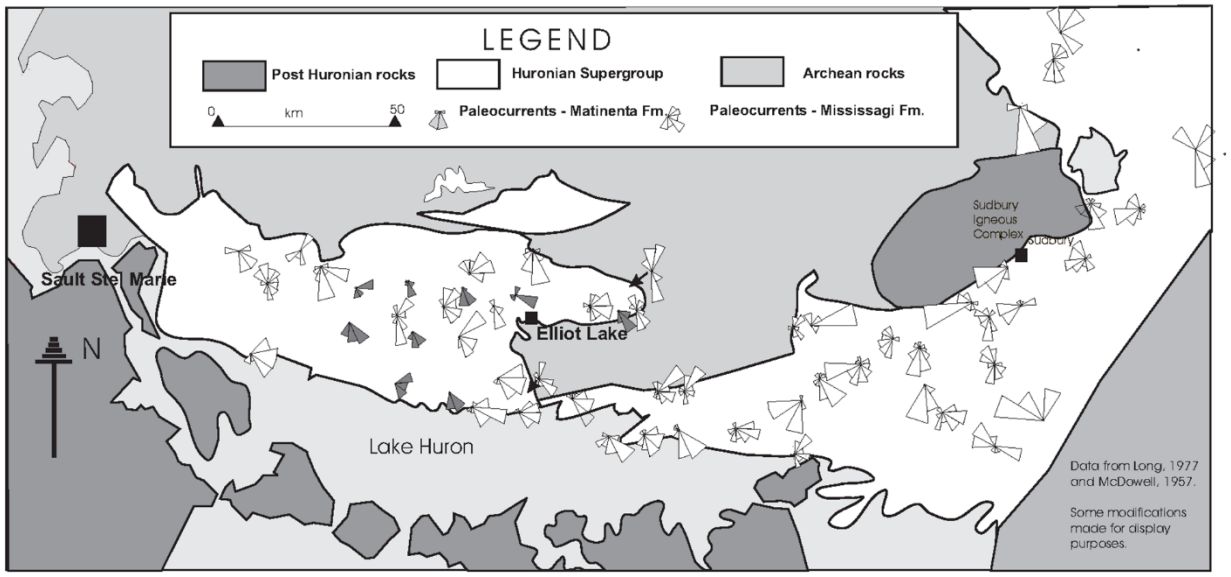


Figure 5. Paleocurrent directions in the Matinenda and the Mississagi Formations (*from* Bennett 2006).

The Huronian Supergroup

Introduction

The Huronian Supergroup is subdivided into 4 groups (Robertson et al. 1969a, 1969b), which in ascending stratigraphic order are: the Elliot Lake Group, Hough Lake Group, Quirke Lake Group and Cobalt Group (Figure 3). Formations of the 3 upper groups, with the exception of the Serpent Formation of the Quirke Lake Group, show regional stratigraphic continuity, and display a remarkable cyclicity of lithological units (Figure 4). Each cycle begins with matrix-supported conglomerate (diamictite), followed by mudstone, siltstone and/or limestone, and capped by a thick sequence of crossbedded, coarse sandstone (Bennett et al. 1991). Paleocurrent studies (cf. Long 1976, 1978; McDowell 1957) have shown flow to the south to southeast, with southeast being the predominant direction (Figure 5).

Conglomerate units (e.g., Ramsey Lake, Bruce and Gowganda formations) in each of the cycles have been interpreted as being glaciogenic in origin, likely deposited in a marine environment adjacent to an ice shelf. The siltstone and sandstone units are interpreted to represent deposition during warmer intraglacial or post-glacial periods in either fluvial or marine environments (cf. Junnila and Young 1995; Fralick and Miall 1989).

The Elliot Lake Group

The Elliot Lake Group differs from the overlying Huronian groups in that:

- its internal stratigraphy is generally discontinuous and less extensive.
- it does not have the diamictite-mudstone-sandstone sequence of the overlying groups.
- it contains the only important uranium deposits and the only volcanic rocks of the Huronian Supergroup.
- st formations have disconformable surfaces.

The Livingstone Creek Formation

Conglomerates and sandstones of the Livingstone Creek Formation (Frarey 1967, 1977) form the lowermost Huronian Supergroup unit The formation is at least 400 m thick in the Sault Ste. Marie area and between 110-300 m thick in the Thessalon area. It consists of 2 distinctive rock types: an upper, well-sorted, grey sandstone and a clast-supported polymictic conglomerate (Bennett et al. 1991). (Figures 3, 6, 7). The Livingstone Creek Formation has not been recognized east of the Quirke Lake Syncline (Bennett 2006).

In most areas, clast-supported, polymictic conglomerate is predominant in the lower sections of the Livingstone Creek Formation. Cobble- to boulder-sized clasts, generally of grey granitic rocks and minor mafic plutonic and metamorphic rocks, are set in a sparse matrix of grey coarse arkose or arkosic grit. Bennett (2006) notes that he had not observed clasts of Huronian Supergroup volcanic rocks in these conglomerates. Locally, thin units of crossbedded, grey arkose are interbedded with the conglomerate (Frarey 1977, Bennett et al. 1991). The granitic mega-clasts of the conglomerate member are predominately pale grey in contrast with the predominantly reddish hues of the underlying Archean basement rocks. Some of the granitic megaclasts in the predominately grey conglomerate near the south end of Pine Ridge Road near Thessalon show the distinct texture of the typical Archean, massive, pink, potassium feldspar-megacrystic granite – but with only a hint of the pink color in the phenocrysts. The grey colour of the Livingstone Creek Formation conglomerates appears to be due to the reduction of ferric iron in the feldspars of the granitic clasts, and not a result of differing provenance as some have suggested. This conclusion is supported by Bennett's (2006) observation that granitic rocks in a "paleosol zone" a few metres to a few tens of metres below the base of the Livingstone Creek Formation commonly are grey as well.

The grey, sandstone member of the Livingstone Creek Formation can be distinguished from most Huronian Supergroup sandstones by its uniform grain size (fine- to medium-sand). In addition, mudstone and pebbly units are lacking in the sandstone member of the Livingstone Creek

Formation, and carbonate occurs along the foreset beds of the well-developed trough crossbeds.

In addition to the well-known exposures in the Thessalon and Sault Ste Marie areas, other possible occurrences of the Livingstone Creek Formation include:

- grey sandstone and conglomerate near Crazy Lake in Nicholas Township (Bennett 1978; Bottrill 1971).
- a basal grey sandstone unit directly underlying the Matinenda Formation in Haughton Township (Bennett et al. 1991).
- an area of clast-supported, grey granite-cobble conglomerate near Samried Lake (Jackson 2001).

The clast size, local source and low stratigraphic position of the Livingstone Creek Formation conglomerates are consistent with deposition as an alluvial fan(s). The uniform, fine- to medium-grained sand of the trough cross-bedded sandstone member suggests a different, although likely related, more distal depositional environment than that of the conglomerate. The sandstone member may represent deposition by median streams flowing in a fault-bounded valley with walls of Archean rocks partly covered by alluvial fans (Bennett et al. 1991). The well-sorted nature of the sandstone suggests an aeolian component or even aeolian deposition as proposed by Meyer (1983).

The Huronian Supergroup Volcanic Rocks – Overview

As noted by Easton (2013a), the transition between dominantly subaerial and dominantly submarine deposition of metavolcanic rocks of the Huronian Supergroup occurs in the Elliot Lake area, and this change in depositional environment may have been significant with respect to sedimentary depositional environments in the Elliot Lake area itself. Thus, it may be no coincidence that Huronian Supergroup mafic metavolcanic rocks are found in proximity to all the past-producing uranium mines and current prospects in the Elliot Lake area.

The Huronian Supergroup Volcanic Rocks of the Sault Ste Marie-Elliot Lake area

Frarey (1967) named the Huronian volcanic rocks in the Thessalon and Sault Ste. Marie areas that overlie the Livingstone Creek Formation as the "Thessalon Formation" (Figures 3, 6).

Bennett (2006), based on his examination of all known exposures of Huronian Supergroup volcanic rocks as well as all available drill-core and drill-hole logs reporting volcanic rocks in the Elliot Lake–Sault Ste Marie area, concluded that there is no credible evidence for more than one period of Huronian Supergroup volcanism in the Elliot Lake–Sault Ste. Marie area and that all the Huronian Supergroup volcanic rocks west of the nose of the Quirke Lake Syncline are stratigraphically correlative with the Thessalon Formation (Figure 6) (Bennett 1978, 2006; Bennett et al. 1991).

Unfortunately, none of the many attempts to obtain an absolute age determination from rocks of the Thessalon Formation have been successful. Nonetheless, there is no reason to think that the age of the Thessalon Formation differs significantly from that of the Copper Cliff Formation (*circa* 2460 Ma).

The maximum thickness of the Thessalon Formation in the Sault Ste. Marie area is approximately 650 to 820 m (Frarey 1977). Diamond drilling has indicated at least 670 m of Thessalon Formation volcanic rock under Lake Huron south of the town of Thessalon, and the formation may be up to 1080 m thick north of Bass Lake in Aberdeen Township (Bennett 2006).

In the Sault Ste. Marie, Thessalon and Aberdeen Lake areas, the Thessalon Formation can be subdivided into an upper tholeiitic basalt unit and a lower complex or "mixed member" (Bennett et al. 1991), which includes fractionated rocks, including basaltic andesite, tholeiitic andesite, mugearite, hawaiite and rhyolite.

Magnesium-rich basalts with some of the chemical characteristics of komatiites are present in the Dollyberry Lake, Pecors Lake and Thessalon

areas. The lower flow sequences show much lower concentrations of Ni, Cr, and contain higher amounts of Ti and P than do the upper flows (cf. Ketchum et al. 2013). In the Dollyberry Lake area, the upper basalts appear to be missing, possibly due to erosion (Bennett 2006).

A comprehensive analysis of the geochemistry of the Thessalon Formation volcanic rocks between Sault Ste. Marie and Thessalon, concluded that the lavas are divisible into 7 distinct units based on mapping, petrography and major and trace element geochemistry (Figure 8) (Tomlinson 1996; Ketchum et al. 2013). The 7 units were grouped into 2 "lava series". The upper lava series (unit 6) is equivalent to the upper tholeiitic basalt sequence of Bennett et al. (1991). The lower lava series (units 1-5, of Tomlinson 1996 and Ketchum et al. 2013) consists mainly of basaltic andesite with subordinate, local rhyolite, mugearite, andesite and high magnesium basalt flows; and corresponds to the "diverse member"of Bennett et al. (1991).

Bennett et al. (1991) proposed that the upper, basaltic flows of the Thessalon Formation (upper lava series) probably represent part of a continental flood basalt sequence, whereas the diverse member (lower lava series) appears to have erupted from central vents. The volcanic rocks of the Quirke Lake Syncline display lithological and geochemical similarities to the lower lava series of the Thessalon Formation west of the Quirke Lake Syncline (Bennett 2006).

The upper, tholeiitic basalt flows of the Thessalon Formation (upper lava series of Tomlinson 1996; Ketchum et al. 2013) are almost uniformly greenish-grey fine- to medium-grained tholeiitic metabasalt. The essential minerals are albite, actinolite, chlorite, clinozoisite, epidote and Fe-Ti oxide. Primary clinopyroxene is present in only a few samples of basalt from the Sault Ste. Marie area. The andesitic rocks of the lower lava series are typically darker and contain stilpnomelane and biotite with green pleochroism (Fe^{+3} rich?) and albite and actinolite. Quartz is a minor component of the basaltic and andesitic types (Bennett 2006).

In most areas the metamorphic grade of the Thessalon Formation flows is lower greenschist facies, although the presence of albite and primary pyroxene, along with elevated sodium contents, indicates sub-greenschist facies at the northern end of the Duncan volcanic belt near Sault Ste. Marie (Bennett 2006).

Amygdules of epidote, chlorite, calcite, quartz and stilpnomelane in complex zonal arrangements are common. Flattened chlorite-filled amygdules a centimetre or less across are a distinctive feature of most mafic flows of the Thessalon Formation. Pillow structures are rare but are present in most areas. Scoriaceous flow-tops and crosscutting breccias are commonly filled with a fine-grained mixture of quartz and grey to red secondary albite (Bennett 2006).

With regard to the geochemistry and tectonic setting of the Thessalon Formation, Tomlinson (1996) stated "that the source of the lavas was metasomatized upper mantle rather than a deep mantle or plume component. Structural subsidence patterns in the Archean basement (Zolnai et al. 1984) are thought to be responsible for lithospheric stretching, in-turn causing mantle upwelling, episodic partial melting and volcanism."

Syndepositional features present where sedimentary rocks of the Livingstone Creek Formation infill fractures in the underlying Archean basement indicate that initially volcanism was a consequence of rifting. In active rifts, a single uplift and melting event occurs as a plume impacts the lithosphere, but in passive rifts uplift and melting are episodic. In addition, the presence of multiple erosional surfaces in the Elliot Lake Group indicate that many episodes of uplift occurred (Bennett 2006). Therefore, the Huronian rifting event can best be characterized as a typical passive rifting event (Tomlinson 1996). This is consistent with Jolly's (1987) conclusion that the Thessalon Formation is a continental flood basalt sequence related to continental rifting.

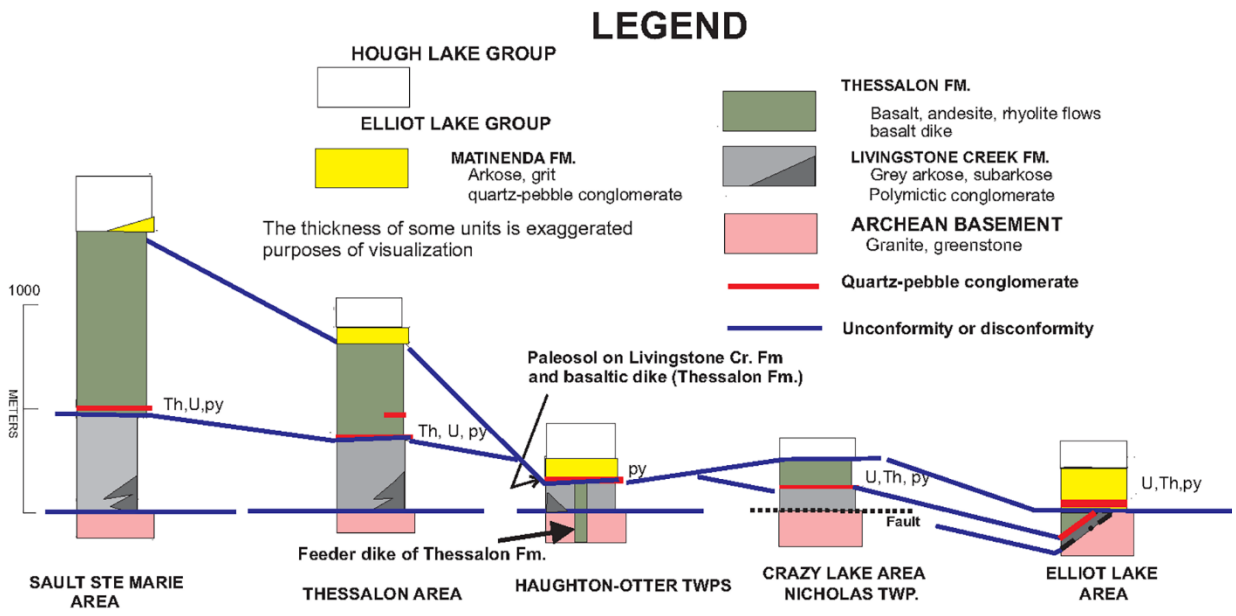


Figure 7. Stratigraphic Relationships in the Elliot Lake Group (*from* Bennett 2006).

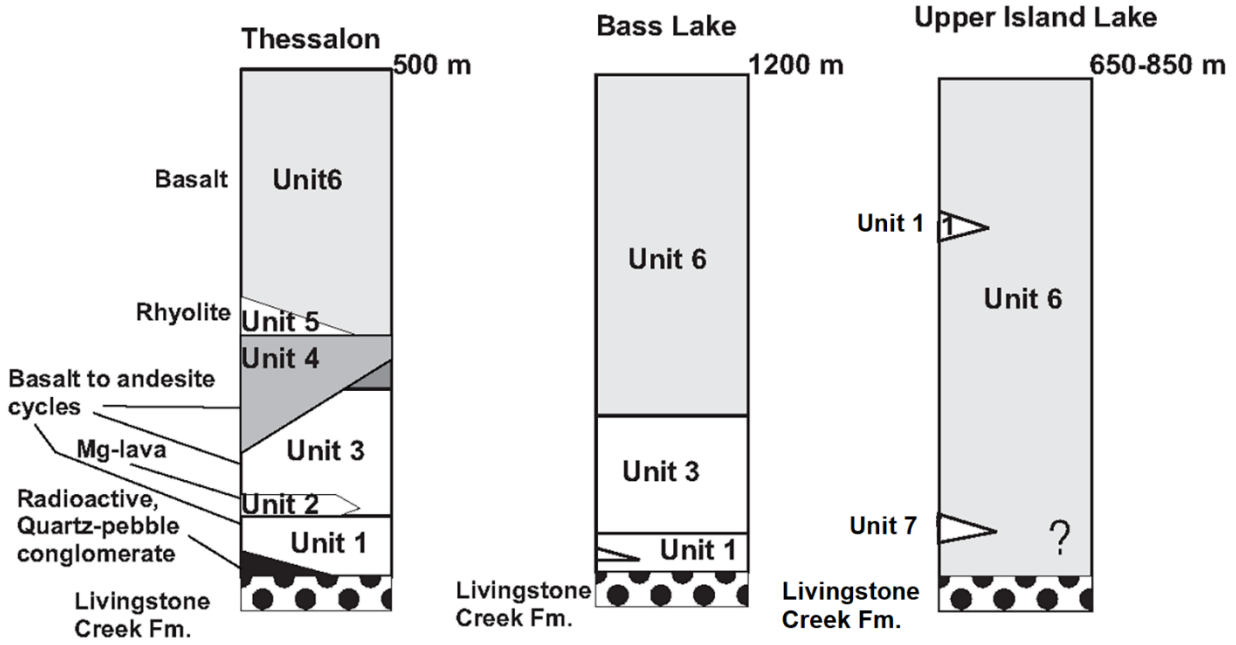


Figure 8. Internal stratigraphy of the Thessalon Formation, Elliot Lake Group, in the Sault Ste. Marie to Elliot Lake area (*from* Bennett 2006, *modified from* Tomlinson 1996)).

Huronian Supergroup volcanic rocks of the Sudbury Area

The volcanic rocks of the Sudbury area differ in terms of internal stratigraphy, overall thickness, and depositional environment from the Huronian Supergoup volcanic rocks in the Sault Ste. Marie–Elliot Lake area.

The Huronian Supergroup volcanic sequence in the Sudbury area has been subdivided into the predominately mafic Elsie Mountain (1000 m thick) and Stobie Formations (1500 m thick), and the felsic Copper Cliff Formation (760 m thick). The depositional age of the Copper Cliff Formation is *circa* 2460 Ma (Beeker et al. 2015; Ketchum et al. 2013; Krogh et al. 1984), and it is likely that the Creighton Granite was coeval with the Copper Cliff Formation (Beeker et al. 2015). The volcanic rocks in the Sudbury area show evidence of submarine eruption from fault-controlled vents along the edge of a depositional basin into which arkosic sandstones were transported from the Archean granitic terrain to the north, with turbidites being deposited from the basin margins (Card 1978a).

Based on recent mapping in the Sudbury area, a preliminary geochemical characterization of the mafic volcanic rocks of the Elsie Mountain and Stobie Formations has been presented (Gordon (2021). Mafic volcanic rocks of the Elsie Mountain Formation are divided into EMF-1 and EMF-2, which are geochemically distinct from each other. Mafic rocks of EMF-1, along with those in the Stobie Formation, are high-Fe tholeiitic basalts. EMF-1 samples in the Elsie Mountain Formation represent the basal lavas, but they are not the most primitive lavas as they have lower MgO, Ni and Cr and higher SiO_2 compared mafic lavas in the overlying Stobie Formation. Mafic rocks of EMF-1 and the Stobie Formation exhibit similar primitive mantle-normalized trace element profiles, characterized by LREE enrichment relative to HREE and negative Nb-Ta-Ti anomalies. EMF-2 mafic volcanic rocks are tholeiitic andesites with distinct primitive mantle-normalized REE profiles characterized by strongly depleted HREE. EMF-2 mafic rocks also have higher Ni and lower V contents compared to EMF-1. On a Nb/Yb versus Th/Yb diagram (Pearce element diagram), the EMF-2 mafic rocks define a distinct trend separate to that of the EMF-1 and Stobie Formation mafic volcanic rocks. Comparing data from Gordon (2021) with that of Ketchum et al. (2013) on volcanic rocks in the Thessalon area (~200 km to the west), the mafic rocks of EMF-1 and the Stobie Formation are comparable to a Thessalon upper basalt-basaltic andesite unit. The rocks of EMF-2 closely resemble a stratigraphically lower basalt-andesite Thessalon unit. Based on the geochemical similarity with the Thessalon volcanic rocks, similar magmatic processes were likely responsible for the generation of the mafic volcanic rocks of the Thessalon, Elsie Mountain and Stobie Formations.

Sedimentary rocks associated with the Thessalon Formation

Some early reports referred to the presence of quartz-pebble conglomerate in the Livingstone Creek Formation, however, this could not be confirmed by Bennett (2006). At many locations, however, a thin unit (< 1 m) of radioactive, pyritic, quartz-pebble conglomerate overlain by a few metres of coarse arkose sand was found to lie upon the Archean basement, or directly atop the Livingstone Creek Formation, where the latter is present. In Duncan Township in the Sault Ste. Marie–Thessalon area, this quartz-pebble conglomerate-arkose sequence occurs in the lower flows of the Thessalon Formation. (Hay 1963; Bennett et al. 1978; Meyer 1983) (Figures 6, 7).

Bennett et al. (1991) proposed that the conglomerate-arkose units indicate the presence of a disconformity between the volcanic rocks of the Thessalon Formation and the Livingstone Creek Formation. The wide distribution of these units (Figure 6) suggests that they reflect an early erosional period of regional extent. The resistant nature of the mineral assemblage in the conglomerate (assuming an oxygen deficient atmosphere) points to a period of extreme weathering. Some of this quartz-rich regolith may still be visible as a quartz breccia atop the granitic basement west of Highway 639 (Optional Stop G).

The Matinenda Formation, Elliot Lake Group

The Matinenda Formation of the Elliot Lake Group is a sequence of arenites and intercalated quartz-pebble conglomerates which host the once strategically important uranium deposits of the Elliot Lake camp where it lies on Huronian Supergroup volcanic rocks and/or the Archean basement (Roscoe 1969; Robertson 1968, 1976). In the Thessalon, Sault Ste. Marie, and Sudbury areas, it consists predominantly of fine-to medium-grained, subarkose to subwacke, and is probably less than 50 m thick (Bennett 1978). In Haughton Township, the Matinenda Formation lies upon grey sandstones equated with the Livingstone Creek Formation (Bennett 2006) (Figure 7).

In the Sudbury area, clastic units correlated with the Matinenda Formation thin rapidly eastward and are intercalated with the mainly metavolcanic rocks of the Stobie Formation and mudstones of the McKim Formation (Card 1978a).

The most abundant rock type of the Matinenda Formation in the Elliot Lake area is generally described as medium- to coarse-grained subarkose, arkose and grit consisting of poorly-sorted quartz and feldspar grains set in a matrix of sericite and comminuted rock and mineral and the fragments. The ratio of potassium feldspar to plagioclase feldspar is about 8:1. Minor constituents are pyrite, calcite, chlorite, zircon and rarely, leucoxene-coated iron oxide and monazite. Varied amounts of sericite give the sandstones a green, apple green or greenish-yellow colouration. Well-sorted, quartz-pebble conglomerate beds, with well-rounded pebbles and cobbles of quartz and chert, and pebbly subarkose units, are scattered throughout the coarse subarkose of the Matinenda Formation, but are more common near the base, in what has been termed the "floater-reef zone" (Robertson 1968; Pienaar 1963) (Stop 3).

Trough crossbedding, scour, and fill structures are common in the subarkose units (Robertson 1968; Roscoe 1969). Paleocurrent studies have established a northwest source area for the sediment of the Matinenda Formation (McDowell 1957; Long 1978; Figure 5). Fralick and Miall (1989) suggested that the Matinenda Formation was deposited from shallow braided streams flowing down a south dipping paleoslope which underwent tilting to the southeast during deposition. Kimberly et al. (1980) reported that the uraniferous conglomerates contained almost no magnetite-ilmenite and had very high K/Na ratios. These are also features of the paleosols beneath the Huronian Supergroup and suggest that the sediment of the Matinenda Formation was formed by the intense weathering of a granitic source terrain, as proposed by Roscoe (1969).

Two southeast trending ore zones were recognized since the early days of uranium mining in the Elliot Lake camp. The Nordic zone, east of the City of Elliot Lake is about 1.6 km (1 mi) wide and 5.6 km (3.6 mi) long. The Quirke Zone, in the Quirke Lake area, is about 3.2 km (2mi) wide and 9 km (6 mi) long. Basement paleotopography is thought to have had a determining influence on the position and orientation of the zones. Ore grade (approximately 0.1 % U_3O_8 (850 ppm U)) conglomerate occurs as persistent lenses with individual units up to 4.5 m thick. The uraniferous quartz-pebble conglomerates are commonly well developed at the base of the Matinenda Formation but also occur in the arkose up to 45 m above the base (Roscoe 1969, Robertson 1968). Total mine production from 1955 to 1990 was 160 million tonnes of ore averaging 896 g/t U_3O_8 for a total uranium metal production of 164,000 tonnes.

The quartz-pebble conglomerate consists mainly of well-rounded, pale- to dark-grey, quartz and chert pebbles in a matrix of pyrite, quartz and/or feldspar grit and sericite. Minable units contain about 15% pyrite. Radioactive minerals include uraninite, brannerite. and uranothorite (Roscoe 1969). Monazite and zircon are common heavy minerals.

The sedimentological and mineralogical features of the uranium-bearing zones of the Elliot Lake camp are generally believed to support a modified paleoplacer origin of the ores as outlined by Roscoe (1969). Advocates of this hypothesis propose that prior to the accumulation of significant free oxygen

in the Earth's atmosphere, southeastward flowing streams carried quartz, pyrite and uraniferous minerals released by the extensive weathering of the Ramsey-Algoma granitoid terrain and deposited them in southeast-trending units constrained by the basement topography (Bennett 2006).

The McKim Formation

The McKim Formation is the uppermost formation of the Elliot Lake Group. In diamond drill core, the contact between the Matinenda Formation and the McKim Formation is interfingering over 1 to 3 metres, consisting of clean sandstone of the Matinenda Formation and mudstone and wacke of the McKim Formation.

Robertson (1968) gives a thickness of 0 to 100 m for the McKim Formation on the south limb of the Quirke Lake syncline. It is missing on the north limb. The McKim Formation is thickest in the Sudbury area, where it is up to 2400 m thick. Card et al. (1977) recognized 3 facies within the McKim Formation:

- the "quartz sandstone" facies is equivalent to the Matinenda Formation, and represents thin interbeds of meta-quartz arenite and minor metaconglomerate in the 2 other main facies of the McKim Formation.
- the "greywacke" facies of interbedded metawacke, metasiltstone and thin bedded mudstone and siltstone. Ripple marks, cross-laminations, graded beds and Bouma cycles are common. Bedding varies from a few centimetres to 50 cm thick.
- the "laminated argillite facies" of thin-bedded mudstone and siltstone, with occasional beds of fine-grained wacke. Bedding is commonly less than 1 cm, and rarely exceeds 10 cm.

Where more highly metamorphosed, rocks of the laminated argillite facies, are best described as metapelites. The metapelites are characterized by high Al_2O_3 contents (20-25 weight %, Easton 2006b; Card et al. 1977), and moderate Fe/Mg ratios (~2), which is probably why metamorphic porphyroblast development is generally restricted to the laminated argillite facies.

The Murray Fault appears to have exerted an important influence on the deposition of the McKim Formation. North of the Murray Fault the McKim Formation rarely exceeds a few hundred metres in thickness, whereas south it is at least 2400 metres thick (Debicki 1990). Card (1978) suggested that the change from laminated siltstone in the west to more wacke in the east indicated a change from more distal to proximal facies, in turn suggesting more tectonic activity and possibly a source for the McKim Formation sediments from the east. Fralick and Miall (1989) concluded that the McKim Formation in the Elliot Lake area represented a marine transgression that gradually drowned the Matinenda Formation fluvial plain.

Aweres Formation

In the Sault Ste. Marie area, the Aweres Formation, a 1700 m thick sequence of conglomerate and sandstone (McConnell 1927), unconformably overlies mafic volcanic rocks of the Thessalon Formation. The internal stratigraphy and rock types of the Aweres Formation are consistent with deposition as an alluvial fan (Bennett 2006).

The base of the formation consists almost entirely of mafic volcanic clasts whereas higher levels show a progressive increase in granitic clasts. The uppermost rocks of the Aweres Formation south of Aweres Lake are mainly arkose with thin pebble conglomerate beds. The lithological variation with stratigraphic height indicates the continual erosion of an uplifted, fault-bounded, plateau of Huronian Supergoup volcanic rocks (Bennett 2006).

The distinct lithology of the Aweres Formation prevents its direct correlation with other Huronian Supergroup rocks. The upper surface is partly fault-bounded. but is unconformably overlain by the Gowganda Formation on Highway 556. It is possible that the Aweres Formation is an erosional remnant of a more extensive alluvial fan system that extended in a more-or-less northeast direction beyond the present northern limit of the Hough Lake and Quirke Lake Groups. The Mississagi Formation may represent a distal depositional environment compared to that of the Aweres Formation (Bennett 2006).

Stratigraphic Relationships within the Elliot Lake Group, Sault Ste Marie-Elliot Lake area

The stratigraphic relationship between the Matinenda, Thessalon, and Livingstone Creek Formations is revealed on a rock face near the northern boundary of Haughton Township about 30 km (18 miles) north of the town of Thessalon (Bennett 2006). Here pyritic quartz-pebble conglomerate of the Matinenda Formation directly overlies an apple-green paleosol on grey, fine-grained sandstone correlated with the Livingstone Creek Formation (Bennett 2006).

About 600 m northwest of the aforementioned occurrence, arkose and quartz-pebble conglomerate of the Matinenda Formation disconformably overlie a steeply dipping, east-striking, mafic dike; the upper few metres of which is a sericite-leucoxene paleosol. The dike cuts grey sandstone and apple-green paleosol of the Livingstone Creek Formation (Bennett 2006).

Less than 2 km south of the above location Chandler (1976) identified a fault-bounded block of Thessalon Formation volcanic rocks with a minimum thickness of approximately 500 m. The mafic dike referred to above was a feeder for Thessalon flows, since the Thessalon Formation is the only known igneous activity at this stratigraphic level (Bennett 2006).

The above observations show that there was a period of volcanic activity, and a period of erosion, separating the Matinenda and the Livingstone Creek Formations. Since paleosols occur upon Huronian Supergroup flows in the Elliot Lake area, the sub-Matinenda unconformity seen in Haughton Township likely extends east to the Quirke Lake Syncline. In addition, the outcrop pattern of the volcanic rocks on geological maps also suggests that the volcanic rocks are erosional remnants preserved in basement depressions (Bennett 2006). The Thessalon Formation may have once extended beyond the limit suggested from its present outcrop distribution (Figure 6), especially if it were a continental flood basalt sequence.

Hough Lake Group

Introduction

The Hough Lake Group (Robertson et al. 1969a; Roscoe 1969) is lowest of the 3 groups of the Huronian Supergroup that display the cyclic deposition of diamictite; mudstone-siltstone and/or carbonate; and arenite. Each cycle is generally thought to represent a sequence of glaciogenic–marine–fluvial and/or shallow marine deposition (Figure 4).

Ramsay Lake Formation

The Ramsay Lake Formation is the lowermost unit of the Hough Lake Group and is the oldest of 3 such conglomerate units that define the base of Hough Lake, Quirke Lake and Cobalt Groups (Roscoe 1969; Pienaar 1963) (Figure 3, 4).

The Ramsay Lake Formation is a widespread, but relatively thin unit. In the Elliot Lake area, the Ramsay Lake Formation ranges from zero to just over 30 m thick (based on diamond drill logs from the Sault Ste. Marie District Geologist's Office). The Ramsay Lake Formation is 70 to 170 m thick in the Sudbury-Manitoulin area (Card 1978a).

Matrix-supported polymictic conglomerate (diamictite) is the most abundant rock type in the formation, especially near the base. Cobbles in the lowermost few metres usually reflect the underlying rock type (Robertson 1968; Parviainen 1973). Locally, minor amounts of mudstone, wacke and arenite are present. Subround to well-rounded pebbles and cobbles of grey granitic rocks and angular to rounded clasts of dark green to black volcanic rocks generally form less than 30 volume percent of the diamictite. The dark matrix consists of quartz, feldspar, chlorite, muscovite-sericite-illite and pyrite (Parvianen 1973).

Although some writers have argued for a debris flow origin, most writers now accept the Ramsay Lake Formation as having a significant glaciogenic component (cf. Roscoe 1969; Robertson 1976). Fralick and Maill (1989) identified an ice-proximal association of pebbly sandstone and diamictite; subaqueous gravity flows and ice rainout deposits; and ice-proximal, fluvial outwash deposits.

Pecors Formation

The Ramsay Lake Formation is conformably overlain by a sequence of generally dark, bedded and laminated wacke, mudstone, siltstone and sandstone (Roscoe 1969). The Pecors Formation is 30 m thick at Quirke Lake (Robertson 1968) but is as much as 900 m thick south of the Murray Fault in the Sudbury area (Card 1978a). It was not identified in the area between Thessalon and Sault Ste. Marie (Frarey 1977). Ripple marks, graded bedding, cross-laminations parallel laminations, ball and pillow structures, clastic dikes and slumpage features have been reported in the formation. The basal part of the formation is commonly laminated, resembling varves, and in places has dropstones (Robertson 1968; Parvianen 1973). Partial Bouma sequences are common (Card 1978a; Robertson 1976). The Pecors Formation is the result of transgressive units formed in deep water by turbidity currents (Card 1978a). The presence of dropstones is evidence of a cold paleoclimate and provides supporting evidence for the glaciogenic origin of the underlying Ramsay Lake Formation.

Mississagi Formation

The Mississagi Formation is a thick sequence of predominantly grey, arenitic rocks extending most of the length of the Huronian Supergroup outcrop belt. In the Quirke Lake syncline, the Mississagi Formation is 344 to 704 m thick. South of the Murray Fault the formation is notably thicker, being more than 3000 m thick in the Sudbury area (Card 1978a; Long 1978).

By far the most dominant rock type in the Mississagi Formation is moderately well-sorted, medium- to coarse-grained subarkose and arkose. Small to medium quartz and/or chert pebble conglomerate is a minor component of the formation; but is more common in the western and northeastern parts of the Huronian belt. Fine-grained pyrite along forsets commonly results in rusty staining of outcrops. Greenish, sericitic units form relatively thin planar-bedded units between crossbedded sandstones. Palonen (1973) provided evidence supporting a marine origin for the Mississagi Formation; but Long (1978) argued that the abundance of mud-grade matrix in the immature arenites, the predominance of unimodal paleocurrent directions, and the lack of quartz arenites argued against a marine environment for the Mississagi Formation. Long (1978) concluded that the Mississagi Formation was deposited from braided streams with low to intermediate sinuosity and high width to depth ratios.

Beds are commonly about a metre thick but can range from a few centimetres to over 4 m thick. Trough cross-stratification and ripple cross-stratification are common sedimentary structures (Long 1978). Cross-stratified beds may show grain-size gradation (McDowell 1957).

Long (1978) measured over 2500 cross-stratified units in the Mississagi Formation (Figure 5) and recognized 2 major stream systems: a stream system flowing southeast to east from the Sault Ste. Marie area, which joined a stream system flowing southwest from the Cobalt Plain, thereby forming a southward flowing system southwest of the Sudbury area. These observations suggest that the area now occupied by the Sudbury Igneous Complex was elevated during the time of Mississagi Formation deposition (Long 1978).

Quirke Lake Group

Bruce Formation

The Bruce Formation extends from the Garden River Indian Reserve near Sault Ste. Marie to about 70 km northeast of Sudbury. It consists mainly of matrix-supported and minor clast-supported conglomerate. Pebbly wacke, arkose, wacke and siltstone are locally present.

The Bruce Formation is from 79 to 12 m thick in the Elliot Lake area and is 26 to37 m thick under the main part of the Quirke Lake Syncline (Robertson 1968).

Pebble- to boulder-sized, angular to subrounded clasts generally consist of pale-grey granitic rocks, Archean supracrustal rocks and fine-grained mafic clasts. The upper parts of the formation may contain up to 5% carbonate (Robertson 1968).

The Bruce Formation is generally interpreted as a tillite with minor beds and lenses of glacially derived sandstone. Dropstones have been observed in laminated units (Robertson 1968).

Casshyap (1969) concluded that the formation was deposited from terrestrial wet-base glaciers. Sims et al. (1981), however, proposed that the Bruce Formation represents an accumulation of debris flows released by normal faulting, a sudden increase in paleoslope, and a sudden increase in water depth. This is consistent with observations made in Porter and Vernon townships showing considerable down-cutting, ranging from 5 to 30 m, of the Bruce Formation into the underlying Mississagi Formation (Easton 2005).

Espanola Formation

The Espanola Formation is the only widespread carbonate unit of the Huronian Supergroup. It is a present from Sault Ste. Marie to the Maple Mountain area, approximately 70 km northeast of Sudbury. Its widespread distribution and distinctive lithology make it the most useful stratigraphic marker unit in the Huronian Supergroup. In the Elliot Lake area, the Espanola Formation can be subdivided into 3 members: a lower limestone member, a middle siltstone- arenite member and an upper dolomite member (Robertson 1968). The latter generally contains 3% to 4% total iron which gives it a distinct brownish hue on weathered surfaces. Contacts between members tend to be gradational. All 3 members are thinly bedded to laminated. The threefold subdivision is less well developed south of the Murray Fault (Young 1982).

Intraformational breccias, mud cracks, ripple-marks, flame structures and ball-and-pillow structures are common sedimentary features. Hofmann et al. (1980) described stromatolites in the Espanola Formation on Quirke Lake. All these features suggest deposition in quiet shallow waters with carbonate deposition being interrupted by influx of fine-grained sediment.

Serpent Formation

The Serpent Formation occurs throughout much of the Huronian Supergroup outcrop belt; however, it was locally removed by erosion during a period of tectonic activity preceding deposition of the Gowganda Formation of the Cobalt Group. Thickness estimates range from 150 to 1500 m (Bennett et al. 1991). According to Robertson (1968), nowhere in the Blind River–Elliot Lake area is there evidence that the total thickness of the Serpent Formation has been preserved.

The Serpent Formation is mainly fine- to medium-grained, quartz arenite and arkose. Conglomeratic units have been noted, especially near its base. Carbonate is a significant component near the base of the formation in the Elliot Lake area (Robertson 1968). Planar and festoon crossbedding, rip-up clasts, fine-laminations, and mud cracks have been reported. Long (1976) proposed that the Serpent Formation was deposited in a distal braided stream environment with calcareous units representing a sabkha environment. Young (1982) noted that the presence of very large crossbeds and a bimodal size distribution suggest aeolian processes may have been active, at least locally.

Cobalt Group

Gowganda Formation

The Gowganda Formation is a complex sequence of conglomerates, sandstones, siltstones and mudstones, and is the lowermost formation of the Cobalt Group. Its thickness ranges from 1070 m in the Sault Ste. Marie area; to 970 to 1150 m around Whitefish Falls on the north shore of Lake Huron; and from 950 to 2700 m near Sudbury. Near Dunlop Lake, in the Elliot Lake area, the Gowganda Formation is about 600 m thick.

Matrix-supported conglomerates are common, especially in the lower parts of the formation. However, these are commonly intercalated with clast-supported conglomerates and sandstone units. Laminated mudstones and siltstone are especially prominent in the upper parts of the Gowganda Formation. Many occurrences of ice-rafted

dropstones have been reported in laminated mudstone-siltstone units. Individual units are generally relatively thin and discontinuous making subdivision of the Gowganda Formation difficult except in well-exposed areas.

Most granitic clasts in Gowganda Formation conglomerates have a distinctly pinkish or reddish hue, in comparison to the grey, granitic clasts in the matrix-supported conglomerates of the stratigraphically lower Ramsay Lake and Bruce Formations. Pink- and red-hued sandstones also first make their appearance in the formation. Roscoe (1969) pointed out the appearance of red coloration (i.e. ferric iron) just above the basal units of the Gowganda Formation, and argued that it represents the appearance of free oxygen in the Earth's atmosphere, and a change from the previously reducing atmospheric conditions that allowed the accumulation of easily oxidized minerals such as pyrite and uraninite. Roscoe (1969) did, however, emphasize that glaciation is only one of several processes likely responsible for the deposition of the Gowganda Formation.

The depositional environment of the diamictites in the Gowganda Formation have been the subject of discussion since Coleman (1905) proposed a glacial origin for these matrix-supported conglomerates. Many subsequent writers including Ovenshine (1965), Casshyap (1969), Lindsay (1971) and Young and Nesbitt (1985) also have supported a glacial, glacial-marine, or glaciolacustrine, origin for the Gowganda Formation diamictites. Card (1968) concluded that, although glaciation may have supplied coarse detritus to the basin initially, debris flows and turbidity currents, related to vertical tectonic movement, may better explain the thickness variations, rock associations and distribution of units in the Gowganda Formation in the Sudbury–Manitoulin area.

Lorrain Formation

The Lorrain Formation is generally well-exposed throughout most of the Huronian Supergroup outcrop belt, where it commonly forms the background to some of the most scenic views in northern Ontario. It is overwhelmingly an arenite sequence, with local siltstone units present in lower parts of the formation. It is up to 2500 m thick near Sault Ste. Marie and in the LaCloche Syncline, southwest of Sudbury. It is up to 2300 m thick in the Cobalt Basin.

In general, the lower part of the Lorrain Formation is dominated by pink, arkosic sandstone; the middle by hematite-rich subarkose and quartz-arenite; and the upper part by pale grey to white mature, quartz-arenite.

A distinctive jasper-pebble conglomerate found in the Sault Ste. Marie area is a popular decorative stone, known locally as "pudding stone". Previously these jasper clasts were thought to be derived from banded iron formations from the Abitibi greenstone belt, however, Bleeker (2018) observed that the jasper clasts of the puddingstone are often angular (i.e. more or less proximal), that they suddenly become a dominant clast type (again suggesting proximal); that there are few if any real banded iron formation clasts; and that the jasper clasts are extremely fine-grained and delicately textured and they do not contain magnetite, unlike banded iron formation samples from the Abitibi greenstone belt which are noticeably more recrystallized. Thus, Bleeker (2018) concluded that the jasper clasts of the Lorrain Formation puddingstone are not of Archean derivation, but represent penecontemporaneous reworking of otherwise poorly-preserved Huronian jasper deposits, possibly associated with a minor volcanic or hydrothermal centre that has not yet been identified. Given that the occurrence of puddingstone is strongly concentrated in the area around Bruce Mines, the source jasper beds were likely local deposits restricted to that part of the Huronian basin, possibly the fine-grained siliceous siltstone and associated layers that have been referred to in some of the early papers on the Huronian as "Bruce Mines Jasper" (Collins 1925).

The presence of aluminous minerals is a characteristic feature of the uppermost quartz-arenites of the Lorrain Formation. Diaspore and kaolinite are common in the Sault Ste. Marie area

and north of Elliot Lake (Wood 1973) whereas kyanite, andalusite and kaolinite occur as metamorphosed equivalents in the LaCloche Lake–Killarney area (Card 1978a). Young (1973) and Wood (1973) interpreted the presence of diaspore and kaolinite as the result of post-depositional, in-situ, alteration of feldspar under hot and humid climatic conditions.

The presence of abundant detrital hematite in the Lorrain Formation and the occurrence of monazite-bearing quartz-pebble conglomerate north of Elliot Lake, have been interpreted by Frarey and Roscoe (1970) as indicating an oxidizing environment.

Planar and trough crossbedding are common, as are ripple marks and other primary depositional structures. There is no consensus as to the depositional environment of the Lorrain Formation. Most of the sedimentary structures present can be found in either shallow marine or fluvial environments. Wood (1973), Young (1973) and Frarey (1977) favored a fluviatile setting, whereas Pettijohn (1970) supported a marine setting. Card (1976) proposed that the Lorrain Formation resulted from near-shore coastal shelf deposition during episodic marine transgression and regression.

Gordon Lake Formation

The Gordon Lake Formation displays a gradational contact with the underlying Lorrain Formation. It is composed predominantly of variegated mudstone, siltstone, chert and minor fine-grained sandstone. The Gordon Lake Formation in the Flack Lake area is subdivided into a lower member of reddish arenite, siltstone, and chert with anhydrite and gypsum nodules; a middle member of green siltstone and mudstone; and an upper member of reddish mudstone, siltstone and chert (Robertson 1986). Sedimentary features include small-scale crossbeds, ripple marks and desiccation cracks.

Some features of the Gordon Lake Formation are unique in the Huronian Supergroup. Wood (1973) noted the abundance of feldspar in marked contrast to rocks of the immediately underlying Lorrain Formation. He also described hematite ooliths and the abundance of grains in the 0.02 to 0.05 mm range, a relatively uncommon grain size in sedimentary rocks. Since this size is found in loess deposits, Wood (1973) proposed that the quartz silt of the Gordon Lake Formation was formed by glacial action, and then carried by the wind and deposited in a tidal flat environment.

Bar River Formation

The Bar River Formation is the uppermost formation of the Huronian Supergroup. It is characterized by quartz-arenite with minor ferruginous arenite and siltstone. It is approximately 300 m thick in the Flack Lake area, north of Elliot Lake. Wright and Rust (1985) concluded that the Bar River Formation was deposited in a tidal environment.

Nipissing Intrusions

Sills, and minor dikes and cone sheets, of gabbro, diabase and granophyre, commonly referred in the older literature (pre-1995) as "Nipissing diabase", are the most widespread igneous rocks associated with the Huronian Supergroup. Nipissing intrusions are widely and evenly distributed throughout the Huronian Supergroup outcrop belt but, and with few exceptions, are not recognized in the Archean Ramsey-Algoma granitoid terrane. Individual intrusions may be up to several hundred metres thick and extend over a strike-length area of 10s of kilometres. There is no current consensus on the tectonic setting for emplacement of the Nipissing intrusions.

Olivine-bearing hypersthene gabbro, gabbro, feldspathic pyroxenite, two-pyroxene quartz gabbro, hornblende gabbro, granophyric gabbro and granophyre have been identified in Nipissing intrusions. Many Nipissing sills are characterized by chilled margins 50 cm to 5 m wide, overlain by 10-20 m of quartz gabbro, then 100-500 m of hypersthene-poor gabbro-norite and vari-textured diabase (Lightfoot and Naldrett 1996).

Baddeleyite and/or zircon from Nipissing gabbro sills in the Gowganda area, the Sudbury area, the

Agnew Lake area and north of Thessalon, have all given ages between 2214 and 2219 Ma (Davey et al. 2019; Noble and Lightfoot 1992, Corfu and Andrews 1986; Easton, unpublished data 2021). Buchan and Card (1985) report that paleomagnetic data suggests at least 2 periods of Nipissing intrusive activity. If so, then the 2 paleomagnetic poles formed in a short timespan of *circa* 5 million years.

Lightfoot and Naldrett (1996) discuss the geochemical characteristics of the Nipissing magmas and the potential for platinum group metal deposits. They concluded that parental magmas of remarkably uniform composition underwent in-situ contamination and differentiation in the intrusions. In addition to nickel-copper-PGE mineralization (Jobin-Bevans 2014, 2016; Jobin-Bevans et al. 1998), a spatial association between Nipissing intrusions and 5-element vein-type mineralization has long been recognized, especially in the Cobalt area (cf. Fyon et al. 1992).

Huronian Paleosols and Evidence for Oxygen Accumulation in the Huronian Atmosphere

Paleosol Evidence

It has long been recognized that the study of paleosols (ancient soil profiles) beneath the Huronian Supergroup could provide information pertaining to the development the Earth's atmosphere and climate during the Proterozoic. Since iron is much less soluble in the ferric state than when in the ferrous state, the behavior of iron in paleosols should provide some indication of the oxygen partial pressure of the environment. Many of the best descriptions of Precambrian paleosols have been from those associated with the Huronian Supergroup unconformity (Gall 1992).

Grandstaff et al. (1986) identified 8 features of paleosols; most of which have been described in paleosols beneath the Huronian Supergroup. These features are:

- stratiform
- relatively thin (<20 m)
- transitional lower boundary-sharp upper boundary
- colour variations
- destruction of primary rock textures accompanied by the development of soil textures
- destruction of primary minerals with formation of clay minerals or metamorphic equivalents
- dikes of material from overlying sediment washed down into desiccation cracks in the soil
- rip-up clasts of overlying sediments

Well-preserved paleosols below the Matinenda Formation in the Elliot Lake area have been described by many workers (Roscoe 1969; Pienaar 1963; Robertson 1968; Frarey and Roscoe 1970; Gay and Grandstaff 1980; Kimberly et al. 1984; G-Farrow and Mossman 1988; Prasad and Roscoe 1991; Sutton and Maynard 1992, 1993; Easton 2013b).

Bennett et al. (1991) proposed that there are 3 disconformities or unconformities in the Elliot Lake Group, which all have the potential for paleosol development (Figure 7). These are in descending stratigraphic order the sub-Matinenda disconformity, the sub-Thessalon Formation disconformity, and the sub-Livingstone Creek Formation unconformity.

The sub-Livingstone Creek Formation unconformity is the lowest unconformity and is the only entirely sub-Huronian Supergroup unconformity (Figure 3, 7). This unconformity is exposed in the Thessalon area, where the upper few metres of the Archean granitic rocks can be seen to progress from angular, slightly rotated blocks, separated by grey grit and fine-grained sandstone, upward, to more rounded boulders with a higher proportion of finer clastic material (Collins 1925). This zone may be termed a "paleo-regolith", since there is little or no obvious development of the yellow, sericitic paleosol commonly found in the younger, sub-Matinenda paleosols.

Prasad and Roscoe (1996) described 2 paleosols in the same diamond drill core from the Denison Mine at Elliot Lake. One was found above Huronian Supergoup volcanic rocks and another, less well-developed paleosol, was found upon Archean tonalite below a short section of quartz-pebble conglomerate and grit below the 9 m thick

volcanic unit (Prasad, personal communication, 1997 in Bennett 2006).

The best developed, and most studied, Huronian paleosols have been found directly below the Matinenda Formation. On mafic rocks, the sub-Matinenda paleosols can generally be recognized by the presence of an upper, distinctly apple-green to yellowish, sericitic zone which grades downward, over a few centimetres to several metres, to a black, fine-grained, chlorite-rich eluvial zone up to several metres thick. Abundant pseudomorphs of titanium oxide after ilmenite are a feature of paleosols on mafic igneous rocks. Rip-up clasts of sericitic paleosol are commonly found in the lower few metres of the overlying Matinenda Formation. Prasad and Roscoe (1996) report significant amounts of carbonate and pyrite in sub-Matinenda paleosols in the Elliot Lake area.

The uppermost sections of sub-Matinenda Formation paleosols developed on Archean granitic rocks is generally an apple-green to yellowish rock composed mainly of quartz and sericite (Robertson 1968; Gay and Grandstaff 1980; Sutton and Maynard 1992), Where the texture of the protolith is well preserved, but the original mineralogy is replaced, the paleosol may be termed a saprolith (Rainbird et al, 1990). The chlorite-rich eluvial zone of paleosols on granitic rocks is generally lacking or relatively thin.

Sub-Matinenda paleosols commonly show the pronounced loss of sodium typical of most paleosols. Calcium and magnesium are also depleted, but there is generally a large increase in potassium content (Gay and Grandstaff, 1980). In most cases, iron and manganese are depleted in the upper parts of the paleosol. This is held to provide evidence of weathering in a reducing environment. Gay and Grandstaff (1980), however, noted an upward increase in total iron in paleosol from the Pronto Mine area. Easton (2013b) locally reported chemical compositions approaching that of a bauxite developed over a mafic substrate in diamond drill core from Elliot Lake.

Gay and Grandstaff (1980) concluded that the upward increase in iron content indicated the presence of free oxygen in early Huronian atmosphere, although at approximately 1% of the present level. They also suggested that the loss of iron in most Huronian paleosols could be due to local reducing environments. Some writers have concluded that the increase in potash (as sericite) in Huronian paleosols is due largely to diagenetic and metamorphic processes that may mask the environmental and hydrologic conditions operative during paleosol development (cf. Gay and Grandstaff 1980; G-Farrow and Mossman 1988).

The mineralogy and geochemistry of sub-Lorrain Formation paleosols described by Rainbird et al. (1990) and Sutton and Maynard (1992, 1993) commonly show an enrichment of Fe^{+3} relative to Fe^{+2} without a significant loss of total iron. Hematite is a common mineral in the upper parts of sub-Lorrain paleosols, in contrast to the presence of pyrite in sub-Matinenda paleosols. In this respect, the sub-Lorrain paleosols resemble many post-Geon 23 paleosols and are consistent with weathering in an oxidizing atmosphere (Prasad and Roscoe 1996; Rainbird et al. 1990).

Other Evidence

Since pyrite and uraninite are unstable under oxidizing conditions, the abundance of detrital pyrite and uraninite in the paleoplacer uranium ore zones in the Matinenda Formation provide evidence for an oxygen deficient atmosphere during weathering, transport and deposition of early Huronian Supergroup sediments.

In contrast to the common red beds of more modern clastic sequences (post-Geon 23), sandstones and most granitic clasts below the Cobalt Group are almost all drab coloured despite the abundance of red and pink granitic rocks in the source area (Roscoe 1969, 1973). Frarey and Roscoe (1970) proposed that the drab colour of lower Huronian Supergroup clastic rocks is due to the lack of free oxygen in the atmosphere during their deposition.

Red-hued, hematite-bearing rocks, which Roscoe (1969) proposed mark the presence of an oxidizing atmosphere, make an appearance with the Gowganda Formation of the Cobalt Group, and are important in parts of the Lorrain and Gordon Lake Formations.

Not all workers, however, accept the above explanation for the preservation of uraninite and pyrite, and the observed change in colour with stratigraphic position. For example, Ohmoto (1996) has stated "the loss of total iron in paleosols of all ages is not due to a reducing atmosphere but to the reductive dissolution of ferric hydroxides under an oxic atmosphere".

Regional Tectonic Patterns and Metamorphism

Major structures in the Huronian Supergroup outcrop belt follow 2 trends: 1) west-northwest trending folds and faults in the Sault Ste. Marie-Elliot Lake area; and 2) east to northeast striking folds and faults in the Sudbury-Manitoulin area south of the Murray fault. These 2 orientations are associated with differing fold styles, metamorphic grade and metamorphic fabric.

In the Sault Ste Marie–Elliot Lake area, fault and fold structures generally trend west-northwest to northwest. Folds are generally upright, and open, with gentle, variably-plunging hinges. There is only weak development of minor tectonic structures; metamorphic grade is subgreenschist (Figure 9, Card 1978b). The major structural features of the Elliot Lake area include a gently south-dipping homocline south of the Flack Lake fault, the open fold of the Quirke Lake Syncline, and the Chiblow Anticline to the south of the Quirke Lake Syncline. In the Elliot Lake area, neither Jackson (2001) nor Easton (2009, 2013a) found any evidence of a detachment at, or near, the basement-cover interface.

Notable changes occur across the northeast-trending faults of the Murray Fault system, the most significant structural feature of the Huronian Supergroup outcrop belt. Because many formations show a significant increase in thickness south of the fault, it is generally interpreted as an inverted growth fault; i.e. an early listric normal fault active during sedimentation; which during later compression became a thrust or reverse fault (Card 1978a; Jackson 2001; Zolnai et al. 1984).

The rocks of the Huronian Supergroup have been subjected to several deformational events (Table 1). This is particularly evident south of the Murray Fault. In the Whitefish Falls area, Young and Nesbitt (1985) concluded that some large-scale folding was related to syn-depositional and/or post-depositional deformation of unconsolidated sediment. Early syndepositional deformation is indicated also the unconformity beneath the Gowganda Formation; and the presence of ragged, slumped contacts and large slump blocks along major faults (Card 1978a; Young 1983).

Convincing evidence of at least one important pre-Nipissing (*circa* 2217 Ma) deformational event, historical assigned to the apocryphal Blezardian orogeny (Stockwell 1982), comes from the observation that Nipissing bodies in the Sudbury-Whitefish Falls area transect axial surfaces of major folds (Card 1978a).

Such relationships are not observed north of the Murray Fault (Jackson 2001; Robertson 1964). North of the Murray Fault, Nipissing sills tend to occupy structures parallel to the axial plane of the Chiblow anticline, suggesting pre-Nipissing. folding. Easton (2006a), in the Porter-Vernon area north of Espanola and north of the Murray Fault, noted that at least 2 periods of folding are present, roughly orthogonal to one another — the resulting interference forms a dome-and-basin pattern (Figure 10). F_1 folds Nipissing gabbro intrusions in the lowermost part of the stratigraphy (in the Hough Lake and Quirke Lake Groups), whereas Nipissing gabbro appears to be emplaced along fractures related to F_2 axial planes (all groups). This suggests either multiple periods of gabbro emplacement, or more likely, that gabbro emplacement occurred syn-folding. In either case, folding cannot be significantly younger than the emplacement age of the Nipissing intrusions.

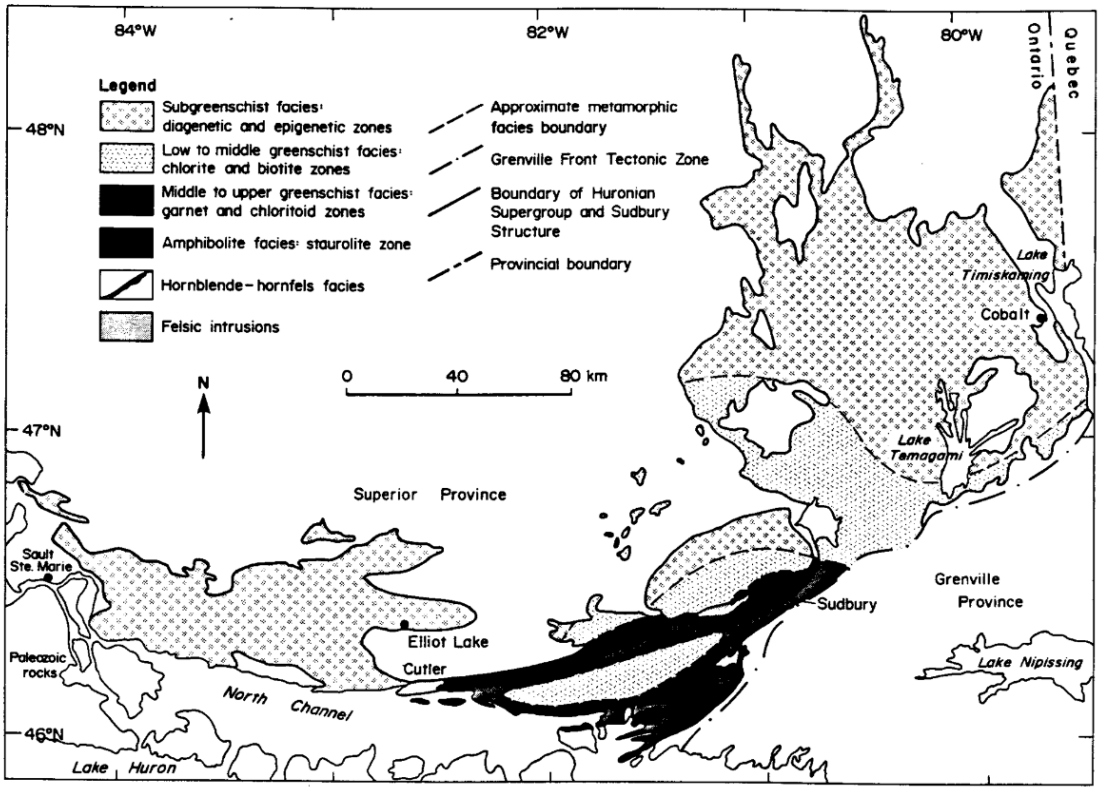


Figure 9. Metamorphism of the Huronian Supergroup. Figure from Card (1978b).

Jackson (2001) also noted evidence of pre-Nipissing faults north of the Murray Fault zone. Easton (2006a), again in the Porter-Vernon area, made the same observation, and recognized at least 5 major fault sets, 4 of which are post-folding.

- The earliest faults are north-trending and juxtapose Archean granitic basement against Huronian Supergroup strata. These faults appear to have been fluid conduits, as indicated by the presence of large quartz vein systems and microbrecciation in the Archean basement, and hydrothermal annealing of quartz in sedimentary rocks adjacent to the faults.
- East-northeast faults also juxtapose Huronian Supergroup strata against basement rocks, but are post- F_1 folding, with both vertical and lateral movement. They may be associated with a set of north to northeast, dominantly normal faults, which may have an older thrust component.
- Most significant in terms of map pattern, at least in the southern part of the Porter-Vernon area nearest the Murray fault system, are east to east-northeast normal faults across which major changes in stratigraphic level occur. There may be a thrust component to these faults, but if so, it has been obscured by subsequent vertical movement, and the fact that these faults are the loci for the development of extensive zones of Sudbury breccia. The localization of Sudbury breccia along this fault set suggests that it may have developed at *circa* 1850 Ma due to the Sudbury impact.
- Finally, significant vertical displacement, occurs along a major set of closely spaced northwest-trending faults. Some of these faults are the loci of Sudbury swarm diabase dikes (*circa* 1240 Ma), which are undeformed and unmetamorphosed, suggesting that this fault set formed between 1850 and 1240 Ma.

Following emplacement of the Nipissing intrusions, but prior to the emplacement of the Sudbury Igneous Complex (1850 Ma), there was further deformation and regional metamorphism. Rb/Sr isotopic studies of Huronian Supergroup metasedimentary rocks indicate that metamorphic resetting occurred at 1900-1850 Ma (Fairbairn et al. 1969). This age range is correlative with the Penokean Orogeny of Michigan and Minnesota (Sims et al. 1981), which has a peak metamorphic age of *circa* 1835 Ma (Holm et al. 2001).

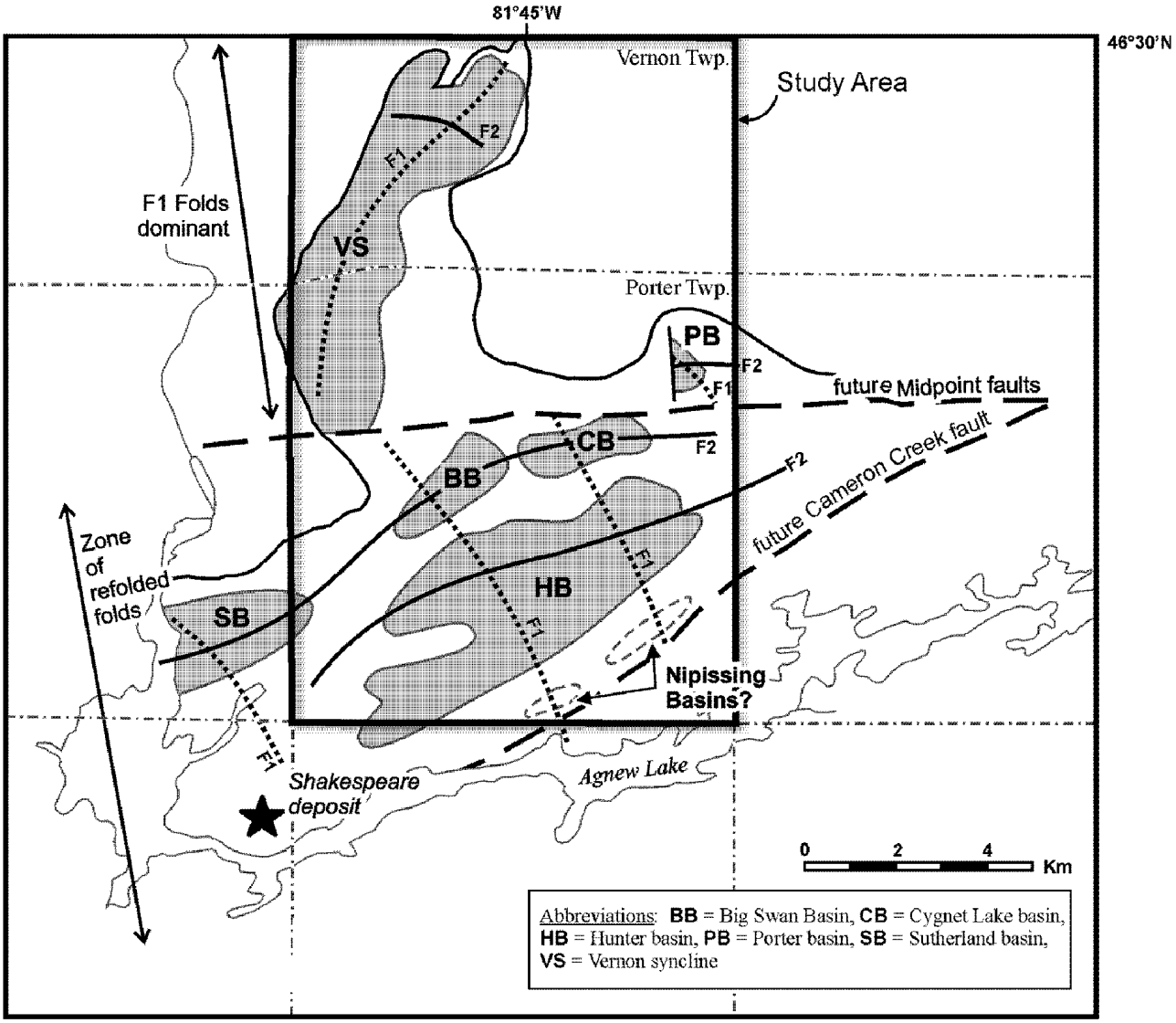


Figure 10. Simplified geological map of the northeast shore of Agnew Lake, showing the distribution of fold styles within Porter and southern Vernon townships. The contact between the Mississagi and Bruce formations has been highlighted to illustrate the fold pattern, and units stratigraphically above the Bruce Formation are shown by a pattern. Between the Cameron Creek and Midport faults, the area is dominated by a dome and basin geometry, indicating the presence of 2 fold generations, with approximately perpendicular axial planes. North of the Midport fault, the early, north-oriented fold style (F_1) dominates. Figure from Easton (2005).

Recently, increasing evidence suggests, that even in its type area, the Penokean orogeny is not as significant a regional event as had been previously thought. In fact, there is increasing evidence that the Yavapai orogeny (Geon 17) may be responsible for much of the deformation previously attributed to the Penokean (cf. Zi et al. 2022; Holm et al., 2018; Schulz and Bjornerud 2018; Raharimahefa et al. 2014). Clearly additional work on the timing and extent of deformation affecting the Huronian Supergroup is needed in Ontario, especially away from Sudbury where there has been considerable resetting of isotopic systems by the Sudbury impact event.

After the emplacement of the Sudbury Igneous Complex (SIC) at 1850 Ma (Davis 2008), and the deposition of the Whitewater Group, there is evidence of further deformation and low-grade metamorphism of the Huronian Supergroup, followed by intrusion of granite plutons at *circa* 1740 Ma and *circa* 1450 Ma, predominately in the Killarney area and in what is now the Grenville Province in the Sudbury area. The intensity of post-SIC deformation and retrograde meta-morphism increase south of the Murray Fault, especially in the area between the SIC and the Grenville Front.

A study of magnetic fabrics, strain patterns, and microstructures in granitoid rocks of the Creighton and Murray granites and their Huronian Super-group host rocks (Riller 1996) lent credence to the concept of a pre-2220 Ma "Blezardian orogeny" (Stockwell 1982). Riller (1996) concluded that major folding and amphibolite facies regional metamorphism in the Sudbury area was coeval with the emplacement of the Creighton the Murray granites, which at the time yielded discordant upper intercept ages of $2333^{+33}/_{-22}$ Ma (Frarey et al. 1982) and $2388^{+20}/_{-13}$ Ma (Krogh et al. 1984). Subsequent work on the Creighton granite by Bleeker et al. (2015), Kenny et al. (2017) and Smith (2002), and on the Murray granite by Krogh et al. (1996), have yielded ages of 2460±20 Ma and 2477±9 Ma, respectively, suggesting that the Blezardian orogeny was not widespread, nor could it have deformed Huronian Supergroup sedimentary rocks that were not yet deposited.

Jackson (2001) also proposed that the "inverted growth-fault" model, as applied by Zolnai et al. (1984) to structural-stratigraphic relationships in the Huronian Supergroup outcrop belt may, in some cases, be interpreted as thrust faults with flats following depositional boundaries, and ramps that cut up through the stratigraphic section. Given the data available, neither model could be rejected for major northwest-trending faults in the Sault Ste. Marie area (Jackson 2001). Jackson (1994) points out that the curvature of the Flack Lake fault is in the opposite direction to that expected if it is a thrust fault, as proposed by Zolnai et al. (1984).

The Murray Fault system separates moderately deformed, low grade metamorphic rocks to the north from multi-deformed, higher-grade rocks of the Sudbury–Manitoulin area to the south. The Sudbury–Manitoulin area is characterized by open to sub-isoclinal, flattened buckle folds with upright to northward overturned axial surfaces. Elongate domes and basin are formed by reversals in plunge. Penetrative axial place cleavage and steeply plunging rodding and/or mineral lineations are well developed. More than one age of major and minor structures can be discerned south of the Murray Fault (Jackson 2001).

Metamorphism south of the Murray Fault ranges from lower greenschist to lower amphibolite facies (Figure 9). Rocks of higher metamorphic grade occur in 2 zones or nodes, one along the Murray Fault system itself and another northwest of the Grenville Front. Both zones coincide with major anticlinoria, although in detail, metamorphic isograds transect fold axes (Jackson 2001). Higher-grade metamorphic nodes do not coincide with the few granitic intrusions that intrude the Huronian Supergroup rocks south of the Murray Fault. The inferred 1900 to 1850 Ma age of metamorphism is much younger than the age of the Creighton and Murray granites (*circa* 2460 Ma) yet older than the *circa* 1740 Ma and 1450 Ma Cutler and Chief Lake granites.

Jackson (2001) considered the origin of the high-grade staurolite-biotite assemblages of the McKim Formation in the hanging wall of the Murray Fault as one of the most enigmatic aspects of the tectonic history of the Southern Province. He concluded that geobarometry indicates a relatively low- pressure metamorphism (2-3 kbar, bathozone 2 of Carmichael 1978) at high temperature (Figure 11). These conditions differ significantly from the 6-7 kbar pressures (bathozone 5 of Carmichael 1978) estimated for the Penokean metamorphism in Minnesota as determined by Holm and Selverstone (1990). Jackson (2001) concluded that the high-temperature metamorphism was at or below pressure corresponding to the thickness of the Huronian Supergroup rock column, thereby precluding crustal thickening as the origin of the metamorphism. Jackson (2001) concluded that a high heat flow regime, such as that developed in areas of crustal extension and related mantle upwelling, was the cause. Such a model is compatible with Card's (1964) view that the high-grade metamorphism may be the result of rapid, focused heat flow.

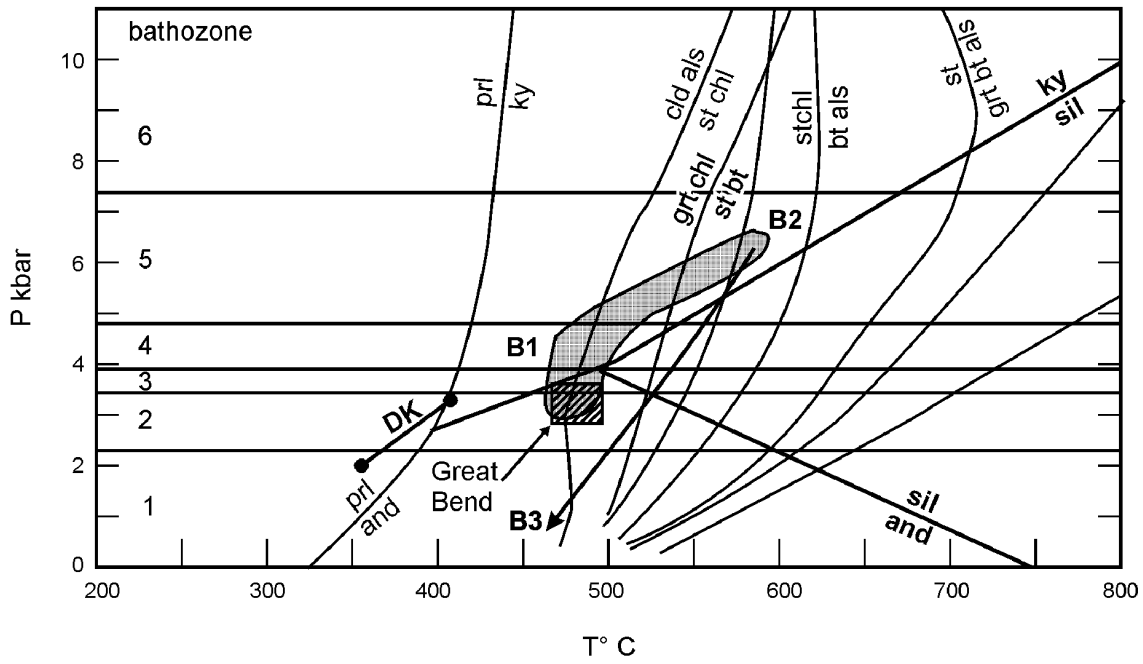


Figure 11. Pressure-temperature (P–T) grid showing the location of major mineral assemblages in the system KFMASH, *after* Spear (1993). Bathozones from Carmichael (1978). Also indicated are possible P–T paths for different parts of the Southern Province. Abbreviations: and = andalusite, as = aluminosilicate, bt = biotite, chl = chlorite, cld = chloritoid; crd = cordierite, grt = garnet, kfs = potassium feldspar, ky = kyanite, ms = muscovite, prl = pyrophyllite, qtz = quartz, sil = sillimanite, st = staurolite; B1 = Baldwin Township initial conditions, B2 = Baldwin Township peak conditions, B3 = Baldwin Township retrograde path; DK = diaspore to kyanite path. Figure from Easton (2006b).

In contrast, at the Great Bend of the Spanish River, interpreted as a high-grade metamorphic node (Card 1978b), and which has been the subject of detailed metamorphic studies (Card 1964; Fox 1971), the co-existing assemblage staurolite-chloritoid in metapelitic rocks of the McKim Formation, along with the local presence of andalusite, places the area in bathozone 3 of Carmichael (1978) (Figure 11). Furthermore, work on McKim Formation metapelites in Baldwin Township that contain relict kyanite, north of the Murray Fault near Espanola, Easton (2006b) concluded that minimum metamorphic conditions corresponded to bathozone 4 (>4 kbar, ≥500°C; Carmichael 1978). Upon reaching peak temperatures, the metapelites cooled quickly to lower grade, likely in a fluid-rich environment, which retrogressed most minerals except for kyanite (Figure 11). Minerals such as chloritoid, andalusite and staurolite, were especially susceptible to retrograde alteration, as they would have already started to break down during the period of increasing temperature (Easton 2006b).

Easton (2006b) argued that tectonic thickening is the most common explanation used to account for the transition from andalusite to kyanite and provides an explanation for the syn-kinematic character of the metamorphic porphyroblasts. It also can account for the differences between Baldwin Township and the Great Bend area.

The model of Easton (2006b) explains other metamorphic mineralogical anomalies in the Southern Province. For example, at low temperatures, but similar pressures, the reaction of diaspore to kyanite occurs, which would account for the presence of reported occurrences of diaspore and kyanite (Card 1978a, 1978b; Church 1967; Chandler et al. 1969). It accounts for the folding of metamorphic isograds, as reported by Jackson (2001) in the May Township area. It also provides an alternate explanation for the high-grade metamorphic nodes in the Southern Province other than the presence of focussed heat and fluid zones proposed by Card (1978b). The resulting tectonic history of the Southern Province, based on the observations in Baldwin Township and the proposed tectonic thickening model (Easton 2006b), presents a considerably more complex metamorphic history for the south-central Southern Province than has been previously envisaged (e.g., Card 1978b; Bennett et al. 1991; Bennett 2006). If nothing else, it emphasizes the need for additional, modern, metamorphic studies of the Huronian Supergroup throughout the Sault Ste. Marie to Sudbury area.

What detrital zircon studies tell us about the source region for the Huronian Supergroup

The publication of the first detrital zircon analyses from the Huronian Supergroup (Rainbird and Davis 2006) took place at the same time as this field trip was last run back in May 2006. Since then, detrital zircon work has been completed on more than 25 samples of the Huronian Supergroup (Table 2), and from almost every unit (except for the Pecors, Espanola and Bruce formations) (Craddock et al. 2013; Davis et al. 2018; Easton and Heaman 2008, 2011; Hill et al. 2018; Kenny et al. 2017; Long et al. 2011; Ménard 2017, 2019; Petrus et al. 2016; Rasmussen et al. 2013). Most of this work occurred in the area between Sudbury and Sault Ste. Marie, all north of the Murray fault, with only 2 samples studied so far from the Cobalt basin northwest of Sudbury. These data are summarized in Table 2, with age ranges and averages based on grains that are < 5% discordant, a lower cutoff than used in many studies. Key observations are:

- Zircons between 2450 and 2490 Ma, likely derived from either Huronian Supergroup volcanic rocks and/or related mafic and felsic intrusions, so far have been reported only from the Matinenda or the Mississagi Formations, generally from sample sites near the base of the formations.
- Samples from the lower Huronian Supergroup (Elliot Lake and Hough Lake Groups) are dominated by Geon 26 detritus (*see* Table 2), consistent with provenance dominated by local sources characteristic of the Ramsay-Algoma granitoid complex. Where detailed stratigraphic sampling has occurred, the lowermost units have unimodal populations,

becoming more diverse with increasing stratigraphic height (e.g., Easton and Heaman 2011). The only exceptions are the 2 samples from the Cobalt basin, which are dominated by Geon 27 populations, consistent with more Geon 27 basement in that area.

- Above the Mississagi Formation, Geon 27 populations are dominant, but Geon 28, 29 and Geon 30 grains are also commonplace (*see* Table 2). This may reflect a change in sedimentation style, and/or increased erosion of the hinterland resulting in a wider range of source material becoming available.
- The uppermost Huronian Supergroup units have ages of *circa* 2310 Ma (Hill et al. 2018; Rasmussen et al. 2013), meaning deposition of the entire supergroup occurred between 2460 to 2310 Ma.
- Persistent throughout the sequence are occasional Geon 25 grains, typically with ages of 2550-2590; these grains become somewhat more abundant in the upper 2 groups. These grains have no known local source, and as suggested by Bleeker (personal communication, 2019). may have a source region to the south, such as the Kaapvall craton, that was subsequently rifted away from North America.
- Currently it is not possible to determine if the detrital zircon populations differ between glaciogenic (e.g., Ramsay Lake) and non-glaciogenic sandstone units. In the Elliot Lake area, the Ramsay Lake Formation has a zircon population consisting only of Geon 26 and Geon 27 grains, similar to the population present in the underlying Matinenda Formation (Easton and Heaman 2011; Ménard 2019).
- Grains >3000 Ma occur sporadically throughout the Huronian Supergroup, mainly in the Matinenda and Mississagi Formations, and could be sourced locally from Michigan (*see* Ayuso et al. 2017). More difficult to explain is the population of 29 ancient grains, 3000-3600 Ma, in the Gowganda Formation sample from Cobalt. Is this sourced locally in the Cobalt area, or have these grains been transported from sources currently exposed on the northeast shore of Hudson's Bay? It is unclear if the sampled unit is glaciogenic or not, as the sampled rock type was not specified by Kenny et al. (2017).

Nd isotope data reported by Easton (2012) supports the conclusions based on the detrital zircon studies. Archean felsic volcanic and granodiorite samples from the Whiskey Lake greenstone belt, have positive εNd^T values close to the depleted mantle evolution curve. Paleoproterozoic rocks of the Thessalon Formation and the East Bull Lake intrusive suite have εNd^T values ranging from 2.58 to -2.28 (Easton 2012; Prevec 1993), suggesting derivation from a primary magma originating from a depleted mantle source, which locally was affected by minor amounts of crustal contamination.

In contrast, the Matinenda Formation sandstone samples have negative εNd^T, ranging from -0.52 to -9.21 (Easton 2012). Samples with the highest negative εNd^T were also enriched in Th, most likely due to the presence of monazite. The magnitude of the negative εNd^T values in the Matinenda Formation indicates a negligible contribution from the volcanic and intrusive rocks of the Whiskey Lake greenstone belt, all of which have positive εNd^T. The Matinenda Formation data can be explained if the sandstones contain a significant component derived from a suite of radiogenic granites located 30 to 40 km north and northwest of Elliot Lake which have εNd^T of -6.19 (Easton 2012). These radiogenic granites typically contain 8 to 33 ppm U and 30 to 50 ppm Th (Easton 2010), thus they are also a potential source of uranium. In contrast, uranium contents of other Archean felsic intrusions in the Elliot Lake area are 1 to 4 ppm (Easton 2010). It may be no coincidence that the most negative of the Matinenda Formation samples was collected only a few metres below the mineralized Main Conglomerate Bed.

In summary, the neodymium data, and the detrital zircon and geochemical data reported by Easton and Heaman (2011), are all consistent with a local source region that included radiogenic granites found north and northwest of Elliot Lake. The absence of similar radiogenic granites north of the Huronian Supergroup between Elliot Lake and Sault Ste. Marie may explain why the Matinenda Formation west of Elliot Lake contains no significant uranium occurrences.

Table 2. Summary of data for all Huronian Supergroup samples based on grains ≤ 5% discordant, in most studies many more grains were analyzed. For samples with significant discordance, the lower numbers shown are for grains ≤ 10% discordant. Also indicated are grains per Geon. All samples are sandstones unless otherwise noted. Samples from the Cobalt Basin are in italics. Abbreviations: cong, conglomerate; EL, Elliot Lake area; MCB, main conglomerate bed; S, Sudbury area; TH, Thessalon area. Table updated from Easton (2019).

Formation	Number	Range (Ma)	Main Peak (Ma)	24	25	26	27	28	29	>3.0
Bar River mudstone	n=16	2279-2745	**2344**		5	1	2			
Bar River EL	n=62	2523-3074	2706 (27>>26)		5	18	30	2	4	3
Gordon Lake EL	n=57	**2284**-2840	**2317,** 2702 (26≈27)	2	3	20	22	1		
Gordon Lake EL		3 sites	**2308, 2308, 2311**							
Lorrain EL	n=30	2684-2890	2713 (27>26)			6	18	6		
Gowganda	*n=172*	*2520-3614*	*2705, 2857, 2965, 3076, 3316 (27>26)*		*2*	*34*	*51*	***38***	***18***	***29***
Serpent EL	n=46	2549-3576	2719 (27>>26)		1	9	22	11		3
Serpent EL-S	n=10	2531-3317	2688 (5%)		1	4	4		1	1
	n=19	2531-3317	2688 (10%)		2	8	8			
Mississagi EL	n=125	2420-3499	**2450**, 2679 (26>27)	**19**	3	45	34	10	1	**13**
Mississagi EL	n=63	2443-3617	**2466,** 2692 (26>27)	**4**	5	24	13	4	7	**9**
Mississagi EL-S	n=22	2591-2832	2663 (26>>27)		2	18	1	1		
Mississagi (upper) S	n=130	2388-3286	**2477,** 2697 (26≈27)	**2**	2	57	57	9		2
Mississagi S	n=117	2414-2978	**2490**, 2560, 2689	**3**	10	47	39	11	7	
Mississagi S	n=72	2544-2949	2683 (26>>27)		4	44	22	2		
Ramsay Lake EL	n=84	2658-2781	2678 (26>>27)			54	30			
Ramsay Lake EL-S	n=65	2656-2887	2697 (26>27)			35	28	2		
Ramsay Lake S	n=25	2526-2719	2659 (26>>27)		3	19	3			
Ramsay Lake S cong	n=37	2607-2821	2677 (26>>27)			27	8	2		
McKim S	n=36	2533-2752	2670 (26>>27)		3	26	7			
Mississagi cong	*n=210*	*2366-2906*	***2459,** 2703, 2771*	***1***	*3*	*78*	*122*	*4*	*1*	
Matinenda S	n=39	2505-3774	2557, 2661 (26>>27)		11	22	5			**1**
Matinenda EL-S	n=27	2451-2714	**2457,** 2671 (26>>27)	**3**		20	4			
Matinenda EL	n=30	2650-2742	2680 (26>>>27)			25	5			
Matinenda (upper) EL	n=47	2620-2897	2664 (26>>>27)			47	3	1		
Matinenda EL	n=36	2617-2776	2649 (26>>>27)			33	3			
Matinenda above	n=5	2634-2651	2641 (5%)			5				
MCB EL	n=15	2621-2684	2643 (10%)			15				
Matinenda below MCB EL	n=28	2546-2838	2641 (26>>>27)		3	24		1		
Livingstone Creek TH	n=37	2507-2890	2698 (26≈27)		1	17	16	3		

Tectonic Models for the Development of the Huronian Basin

Various tectonic models have been proposed for the early development and later deformation of the Huronian basin. Many reconstructions are essentially modifications of the model put forth by Dietz and Holden (1966), which stated that the Huronian Supergroup represents a rift and passive margin sequence that was compressed, partly tectonically buried, and metamorphosed during a collision with the Superior craton and another mass which overrode its southern edge. Zolnai et al. (1984) and Bennett et al. (1991) accepted the essential aspects of the Dietz and Holden (1966) model.

The model proposes that rifting and continental break-up was coeval with Huronian Supergroup volcanism (2475-2450 Ma) and that the much later regional deformation occurred coincident with the Penokean Orogeny (1860-1835 Ma). This model does not attempt to account for the multiple deformation events affecting Huronian Supergroup rocks or the origin of the Nipissing intrusions.

Young (1982) proposed that the Huronian Supergroup was deposited in an aulocogen, an easterly-trending fault-bounded trough, which opened toward an ocean in the area now occupied by the Grenville Province. Sims et al. (1981) concluded that the Huronian Supergroup, the Marquette Range Supergroup, and Animikie Group rocks were deposited as intra-continental, fault-controlled basins along a major, Neoarchean structure, the Great Lakes Tectonic Zone.

More recently, Roscoe and Card (1992), noting the close stratigraphic correlation between the Paleoproterozoic sequences of the Wyoming craton and the Huronian Supergroup, proposed that the Superior and Wyoming cratons are rifted portions of what was once a single continental land mass. They suggest the direction of the Matachewan-Hearst dike swarm (2475-2460 Ma) indicates an east-west tensional regime, which resulted in, a Huronian basin elongated in a north-south direction. On this larger craton, sediment was deposited in a southward-deepening intracratonic basin. Roscoe and Card (1992) proposed that it was during the Nipissing igneous event (2217 Ma) that successful rifting of the Superior Province took place with the eventual drifting of part of the missing Superior Province to its present location as Wyoming craton. They attribute pre-Nipissing folding to the Blezardian orogeny of Stockwell (1982) and the later, more important, deformation to be coeval with the Penokean orogeny of Michigan, Wisconsin and Minnesota (Roscoe and Card 1992).

Jackson (2001) supported the model of Roscoe and Card (1992) since the high heat flow, which he considers necessary to give the observed features of the high-grade metamorphic rocks, would be a necessary effect of mantle upwelling during continental break-up. He also interpreted some of the early high-strain deformation as being consistent with a Nipissing-age break-up of the Superior craton.

Acknowledgements

The senior author wishes to thank Mike Hailstone, former Sault Ste. Marie Resident Geologist (after Gerry Bennett) for introducing him to this trip back in 2008. This was followed by a mapping program in the Elliot Lake to Pecors Lake area between 2009 and 2011.

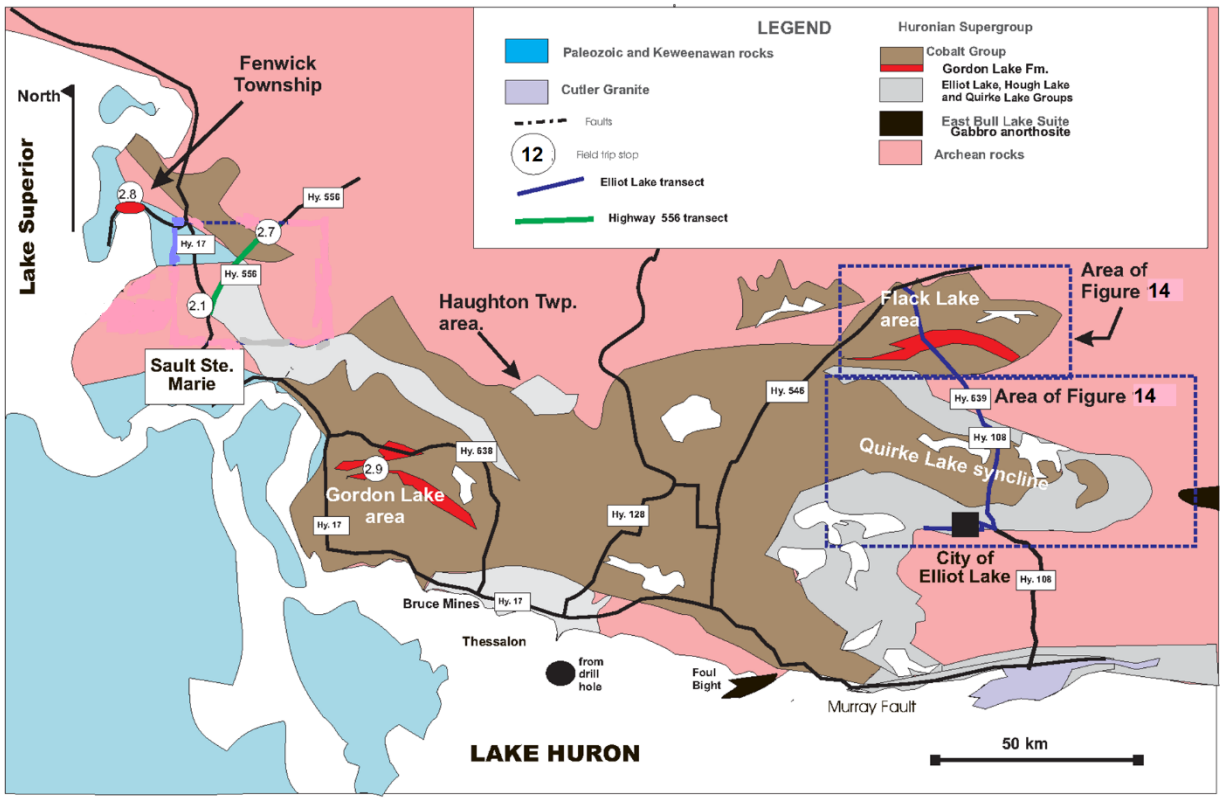
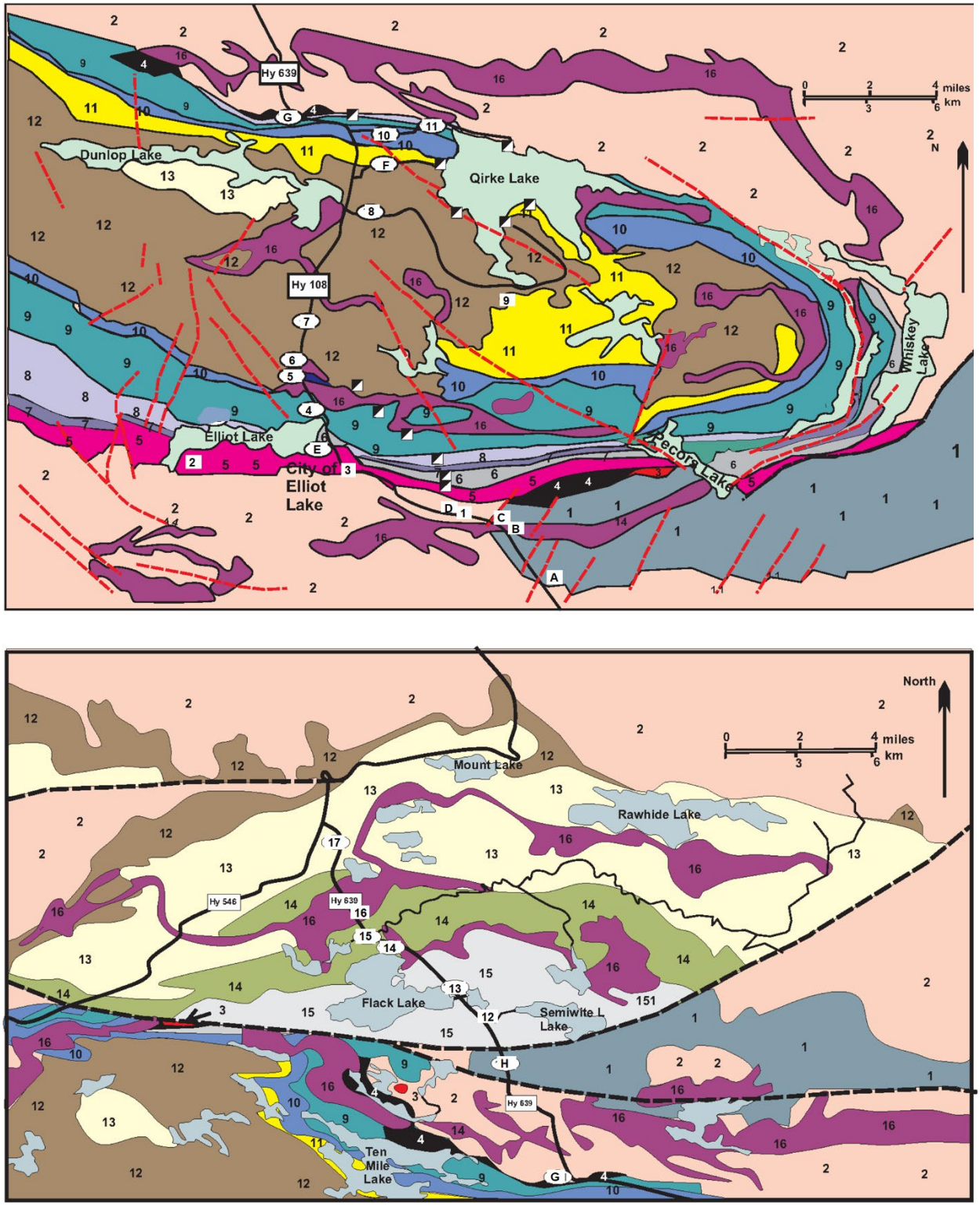


Figure 12. Index map for included geological maps and some areas mentioned in the text. Figure *modified from* Bennett (2006).



Figure 13. Legend for Figure 14. Figure *modified from* Bennett (2006)



Figures 14. Upper. Geological map of the Quirke Lake syncline with stop locations. **Lower.** Geological map of the Flack Like area with stop locations. Both *modified from* Bennett (2006).

FIELD TRIP DETAILS

Geological Maps

Geological compilation maps covering all or parts of the area of the field trip include Giblin and Leahy (1979); Johns, McIlraith and Muir (2003); and Easton (2013a). The southern part of the field trip area, including optional Stops A-C and Stop 1, as well as the Whiskey Lake greenstone belt and the eastern part of the Quirke Lake syncline, are depicted on Easton (2013a).

ROAD LOG

Note: Caution should be taken when parking vehicles on the shoulder of the highway and when examining outcrops located along Highways108 and 639 and on other roads along the field trip route. All UTM co-ordinates are given in NAD 83 datum, zone 17, which is essentially equivalent to NAD WGS84.

Note: This guidebook describes a total of 25 stops (17 stops and 8 optional stops). If one is starting in Elliot Lake, it is possible to do all of the stops in one day. If coming from Sudbury, then some stops will need to be omitted (mainly the optional stops). All 16 stops from Bennett (2006) are included in the road log, along with an additional 9 stops added by the senior author.

Leave from Science North at the junction of Paris Street and Ramsey Lake Road in Sudbury. Head to Highway 17 and proceed west toward Sault. Ste. Marie. At the junction of Highway 17 and 108, turn right onto Highway 108 and head north to Elliot Lake (29 km (18 mi) from, Highway 17). Set odometer to zero at this point.

From the junction with Highway 17 to the bridge at Depot Lake, the highway passes through the Ramsey-Algoma granitoid complex. The rocks along the highway have not been studied in detail. The granitoid rocks include xenoliths of mafic rock and are cut by numerous dikes of the Matachewan dike swarm (*circa* 2460 Ma), which are medium-grained, medium-green, and locally plagioclase porphyritic. Also present are fine-grained, flinty, mafic dikes that may represent feeders to Huronian Supergroup volcanic rocks.

0.0 km. Bridge across Depot Lake (approximately 15.4 km from visitor centre). Enter into the Whiskey Lake greenstone belt (Archean). Set odometer to zero.

1.0 km. Approximately a kilometre north of the bridge, pull over on the right shoulder on the passing lane up the hill. Examine exposures on the large rock face on the right side of the road. Highway 108. This is a new stop.

Optional STOP A: Archean metasedimentary rocks of the Whiskey Lake greenstone belt

UTM co-ordinates 381165E 5133640N

Here we see thin- to medium-bedded turbidites of the Archean Whiskey Lake greenstone belt. Partial Bouma sequences are present in the thicker turbidite beds. The turbidites likely have a high felsic volcaniclastic component, as a sample from this outcrop for detrital zircon study yielded no zircons or titanite (Photo1).



Photo 1. Thick volcaniclastic bed in turbidites at Optional Stop A. Hammer handle is 33 cm long.

Although not observed in this outcrop, as one heads along the highway and up stratigraphy, the turbidites pass into thinly bedded mudstones and then into a magnetite facies iron formation. These metasedimentary rocks lie atop the older (*circa* 2740 Ma) of the 2 volcanic sequences that comprise the Whiskey Lake greenstone belt (Easton 2013a).

Return to vehicles, continue on Highway 108

2.3 km (1.44 mi). Turnoff to Elliot Lake airport on the left. Continue straight.

3.6 km (2.25 mi). Pull over on the shoulder of the road. Examine long roadcut on the right side of the highway. This is a new stop.

Optional STOP B: Stone Ridge intrusion

UTM co-ordinates 379693E 5135505N

The roadcut exposes part of an east-trending metagabbro intrusion, termed the Stone Ridge intrusion. The intrusion is 700 to 1000 m wide, with a minimum strike length of 15 km. It lies 1 to 2 km south of, and roughly parallels, the Archean–Proterozoic unconformity (Easton 2009, 2013a). As seen in this roadcut, large parts of the intrusion contain preserved primary mineralogy. The predominant rock type is a weakly metamorphosed, grey to light grey weathering, medium-grained, leuconorite to leucogabbro-norite. Where recrystallized, orthopyroxene is altered to amphibole, and the rock takes on a greener colour. Texturally, the body is remarkably uniform, but coarse-grained to pegmatitic patches of gabbro occur along the northern margin of the intrusion. Matachewan dikes (*circa* 2460 Ma), which were observed to intrude the body (Easton 2009), would preclude the Stone Ridge Intrusion being part of the Nipissing intrusive suite, which was not emplaced until *circa* 2217 Ma, suggesting that the Stone Ridge intrusion is more likely part of the East Bull Lake intrusive suite.

Return to vehicles, continue northward of Highway 108 to the junction with Nordic Road and the golf course.

5.3 km (3.31 mi.) Junction with Nordic Road. Park safely and examine outcrops on the northeast and northwest sides of the intersection. This is a new stop.

Optional STOP C: Archean metavolcanic rocks of the Whiskey Lake greenstone belt

UTM co-ordinates 378483E 5136436N geo-chronology site; 378486E 5136443 flattened tuff

The northeastern outcrop consists of reversely graded felsic tuffs and felsic tuff-breccias of the Archean Whiskey Lake greenstone belt (Photo 2). A U/Pb zircon sample from this outcrop yielded a CA-TIMS age of 2724.9±1.4 Ma (Hamilton *in* Easton 2010) indicating that these rocks are among the oldest in the greenstone belt.

The northwestern outcrop consists of extremely flattened tuff-breccia (Photo 2). The higher strain in this outcrop may reflect its location in the contract strain aureole of the large granodiorite body that we will examine at Stop 1. Return to vehicles, continue straight (westward) on Highway 108.

6.1km (3.81 mi). Turnoff to the left takes you to the Discovery Site lookout. Several large boulders representative of the uranium-bearing "Main Conglomerate Bed" are present in the parking area of the lookout. Looking northeast from the highway and/or the lookout, you can see a large ridge of greenish sandstone of the Matinenda Formation. The first mine in the Elliot Lake Camp, the Buckles Mine, was located at the base of this ridge opposite the turnoff.

6.8 km (4.25 mi). Pull over and park near the middle of a large roadcut on the north side of the highway This is a new stop.

STOP 1. Thessalon Formation feeder dike and Archean granodiorite

UTM co-ordinates 377092E 5136767N

The roadcut contains a 15 m wide, near-vertical mafic dike which is vesicular (Photo 3). The vesicular nature of the dike is best observed on the roadcut on the south side of the Highway (Photo 4). The mafic dike has the composition of a tholeiitic andesite on a Jensen discrimination diagram (54.9 wt.% SiO_2, 2.38 wt.% K_2O, magnesium number of 36 (data in Easton (2013b)) and is a high-K basaltic andesite in the IUGS total-alkali silica classification.



Photo 2. Felsic volcanic rocks at Optional Stop C. **Upper.** Lapilli tuff in the northeastern outcrop which was sampled for geochronology. Knife is 9 cm long. **Middle.** Moderately flattened tuff breccia from the northwestern outcrop. **Lower.** Strongly flattened tuff breccia from the northwestern outcrop. Hammer handle is 33 cm long.



Photo 3. Roadcut on north side of Highway 108 at Stop 1. The dark, 10 m wide, vertical, steep-walled mafic dike is intruded Archean granodiorite, which has a U/Pb zircon age of 2674.8±0.8 Ma (Easton 2013a). This dike likely feed Thessalon Formation flows.

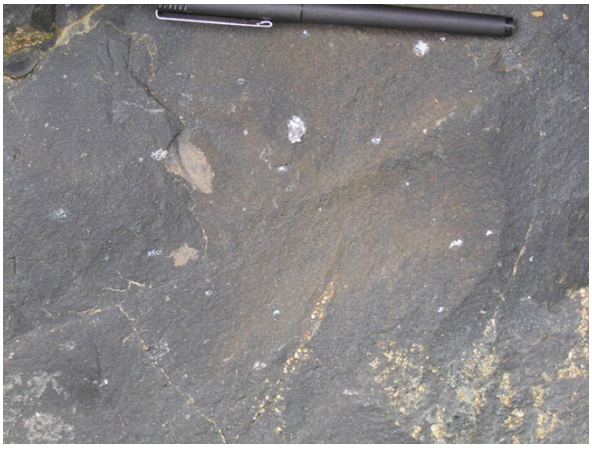


Photo 4 Close-up of mafic dike rock on the south side of Highway 108 at Stop 1 with irregular, white vesicles in the dike. Pen is 13.5 cm long.

The medium-grained granodiorite that forms the bulk of the roadcut is typical of the younger intrusive bodies within the Ramsey-Algoma granitoid belt. A sample collected from the roadcut on the south side of the road yielded a CA-TIMS age of 2674.8±0.8 Ma (Easton 2013a). The youngest age reported so far from metavolcanics of the Whiskey Lake is 2685.5±1.1 Ma (Easton 2013a).

Return to vehicles, continue straight (westward) on Highway 108.



Photo 5. Paleoweathering in Archean granodiorite at Optional Stop D. **Upper**. Grey, unweathered granodiorite. **Middle.** red-green weathering granodiorite. **Lower.** Green weathering granodiorite. See Table 3 for chemistry on each type. Pen is 9 cm long.

7.1 km (4.44 mi). Pull over and park on the shoulder of the highway. Cross over to the former route of the highway and walk west toward the intersection of the new and the old highways. This is a new stop.

Optional STOP D: Paleoweathering of Archean granodiorite

UTM co-ordinates 376567E 5136603N

Proceeding west along the outcrops present on the north side of the former highway, you can observe the progression from grey granodiorite to reddish-weathering granodiorite to greenish granodiorite, with the change in coloration reflecting increasing paleoweathering of the granodiorite (Photo 5). Key elemental changes between the 3 phases are summarized in Table 3, and include increasing iron, manganese, magnesium, potassium and aluminum contents, with decreasing silica content.

Table 3. Element changes related to paleo-weathering in granodiorite at Optional Stop D. Data *from* Easton (2013b). Major elements are in weight percent. Abbreviations. LOI, loss on ignition; CIA, chemical index of alteration.

Element Photo 5	Grey, not weathered upper	Red weathered middle	Green weathered lower
SiO_2	69.1	60.1	52.7
Al_2O_3	15.7	19.4	19.6
$Fe_2O_3^{total}$	3.2	5.0	9.9
MnO	0.02	0.06	0.12
MgO	1.9	2.7	5.5
CaO	1.0	0.5	0.7
Na_2O	6.8	7.9	6.9
K_2O	1.0	2.5	0.8
LOI	1.3	2.0	3.5
CIA	89	87	70
Th (ppm)	7	12	13
TiO_2, P_2O_5, U, Zr nearly constant in all 3 samples			

Return to vehicles, continue straight (west) on Highway 108. Continue straight on the highway past Hillside Drive South.

9.2 km (5.75 mi). Roadcut to the right of the highway just west of the junction with Esten Drive north consists of coarse sandstone and pebblestone of the Matinenda Formation similar to what we will see at Stop 3. The pink colouration in this roadcut is related to its proximity to a fault located approximately along the highway route.

Continue straight on the highway and past the first few stoplights to Hillside Drive North. Turn left (west) onto Hillside Drive North.

If going directly to Stop 2, once on Hillside Drive North, continue west for approximately 1 km to Spine Road. Turn right (west) onto Spine Road and drive past the hospital to Lawrence Avenue at the far west end of Spine Road (~2.1 km). Park in the turn-around at the end of Spine Road for Stop 2. This is Stop 15 of Bennett et al. (1997) and Stop 1.2 of Bennett (2006).

If going to Optional Stop E, continue west on Hillside Drive North for about 400 to Spruce Avenue. Turn right on Spruce Avenue and continue approximately 100 m to Valley Crescent, turn right onto Valley Crescent and follow it for approximately 350 m to Balsam Place. Turn right on to Balsam Place and stop at the end of the cull-de-sac. This is Stop 1.1 in Bennett (2006).

Optional STOP E: McKim Formation and Nipissing Diabase

UTM co-ordinates 372751E, 5138695N

Outcrops on the east side of the cul-de-sac are mudstone and grey sandstone of the Mckim Formation. Note the deflection of the axial plane cleavage in the mudstone units. The movement of adjacent beds inferred from the deflection of the cleavage indicates the south limb of a syncline. A gabbroic dike is separated from the sedimentary rocks by a zone of sheared and fractured rocks. The McKim Formation is missing on the north limb of the syncline.

STOP 2. Radioactive quartz-pebble conglomerate, Matinenda Formation

UTM co-ordinates 370462E, 5137991N

The low outcrops on north side of Spine Road are grey, buff and dark-grey sandstone and radioactive, pyritic, quartz-pebble conglomerate of the Matinenda Formation. Pebble units, located near the top of the outcrop, are rusty-weathering, about 20-30 cm thick, and dip about 10 degrees to the north (Photo 6). Pebbles in this outcrop are 1-2 cm across and are generally much smaller than the typical pebbles in the ore zones of the Elliot Lake mines. Scintillometer readings from the pebble beds, collected by R.M. Easton between 2009 and 2022, range from 15,000 to 18,000 counts per second, with 600 to 800 ppm U and 600 to 800 ppm Th, with the U content being typical of the ore grades from the Elliot Lake camp.



Photo 6. Matinenda Formation at Stop 2. Hammer head is resting on radioactive, rusty-weathering pebble beds that are the focus of this stop.

Ruzicka and LeCheminant, (1984) reported the radioactive conglomerate contains “rare-earth-element-bearing uranothorite (?), large zircons, a Ti-U-Si-Fe phase (brannerite?), chalcopyrite and chromite. The distribution of radioactive minerals in the conglomerate displays layering thus indicating a detrital origin of these grains”.

The dark-grey areas in the radioactive beds are due to the presence of minor amounts radioactive carbon generally known in the Elliot Lake area as "thucolite" also referred to as a hydrocarbon kerogen. Ruzicka and LeCheminant, (1984) noted that several generations of carbon occur in the conglomerates of the Matinenda Formation. The earliest generation occurs as layers concordant with the bedding or as a component of the matrix and appears to have been deposited in areas of quiescent sedimentation during the last phase of an upward-fining sedimentary cycle.

Later generations of thucolite are probably remobilized phases of the first generation. The carbonaceous matter in the Elliot Lake ores is comparable in occurrence and composition to hydrocarbon in the Witwatersrand gold reefs; interestingly Ruzicka and LeCheminant (1984) report elevated gold content (1000-2000 ppb) in the carbonaceous matter of the Elliot Lake ore beds. The radioactive carbon at this site is reported to be auriferous, although the gold content is not available.

In 1955 Rio Algom Mines Limited completed a diamond drill hole about 30 m south of this location. The drill log shows that the radioactive beds exposed here are about 35 metres above the Archean basement rocks. This drilling indicated that there are no ore-grade units in this area.

Grab samples collected by G. Bennett in 1982 and reported in Bennett (2006) returned up to 0.80 lbs U_3O_8/ton and 0.78 lbs ThO_2/ton (340 ppm U, 340 ppm Th). A continuous chip sample returned 0.31 lbs U_3O_8/ton and 0.53 lbs ThO_2/ton (130 ppm U, 232 ppm Th).

Retrace route on Spine Road to Hillside Drive. At the traffic light, turn right onto Ontario Drive heading south. Vacant lot on the right is the site of the former Algo Mall, which had a catastrophic collapse on June 23, 2012. Continue for approximately 450 m on Ontario Avenue and where the road bends, continue straight into the retail mall parking lot. Park and examine the large rock face on the west side of the mall parking lot. This is a new stop.

STOP 3. Coarse sandstone, floater reef zone, Matinenda Formation, Elliot Lake Group

UTM co-ordinates 373385E 5137830N

Created in 2014, this rock face exposes greenish coarse sandstone and pebblestone of the Matinenda Formation. Compared to Stop 2, we are now only 25 m higher in the stratigraphy above the "Main Conglomerate Bed" in what is locally referred to as the "Floater Reef Zone". The name is derived from the fact that many of the pebblestone beds are variably radioactive (typically 2,000 to 5,000 counts per second). Scintillometer data collected by the senior author from this Stop give 20 to 25 ppm U and 130 to 150 ppm Th.

At the north-end of the rock face is a 5 m wide medium-grained mafic dike cutting the Matinenda Formation. The affinity of the dike is unknown, and no geochemical data are available for this dike. It cannot be a Matachewan dike as it cuts the Matinenda Formation, and it does not have the scintillometer characteristics of a Nipissing gabbro (potassium is too low). It could be related to a suite of east-trending dikes in the Elliot Lake area, such as the one at Stop 14.

Return to vehicles. Exit the south end of the mall parking lot, turn left onto Hillside Drive South, continue east on Hillside to the traffic lights (Highway 108), approximately 300 m.

Turn left onto Highway 108 and head north. Reset odometer to zero

1.1 km (0.69 mi). Miners Memorial Park and Horne Lake on the right. Cliff on the east side of Horne Lake consists of Mississagi Formation sandstone (Photo 7).

2.1 km (1.31 mi), Westview Park on Elliot Lake on the left.

2.4 km (1.5 mi). Stanleigh Road on the right,



Photo 7. View from the Miners Memorial across Horne Lake showing cliff of Mississagi Formation sandstone.

3.5 km (2.19 mi). Pull-over where snowmobile-ATV trail intersects Highway 108. Cross Highway to outcrop on the west side. This is Stop 2.3 of Young (1991), Stop 16 of Bennett et al. (1997) and Stop 1.3 of Bennett (2006).

STOP 4. Mississagi Formation, Hough Lake Group

UTM co-ordinates 371717E, 5140426N

One-metre-thick beds of grey sandstone of the Mississagi Formation on the west side of the highway display the rusty staining on the face of the outcrop reflecting the minor pyrite content along the foreset beds of trough cross-beds (Photo 8). The paleocurrent direction (from the west) can be best observed on the upper surface of the outcrop. — ***Please exercise caution when walking on smooth, wet rock surfaces*** — The grey colour of these sandstones and the presence of apparent detrital pyrite are held by most geoscientists to indicate the very low partial pressure of free oxygen of the atmosphere during the deposition of the Mississagi Formation.

J.A. Ménard and R.M. Easton collected a sample for detrital zircon geochronology from this stop in 2017 (Photo 8). The zircon population ranged in age from 2420 to 3499 Ma, with notable peaks at 2450 Ma (13 grains) and 2679 Ma (45 grains), and with 10 Geon 28 grains and 13 grains ≥3000 Ma (Ménard 2019). The abundance of *circa* 2450 Ma grains compared to other Mississagi Formation samples (*see* Table 2) is not unexpected given the abundance of Elliot Lake Group metavolcanic rocks to the north of this site near Dunlop Lake (R.M. Easton, unpublished data). The broad age range of zircons in this sample contrasts greatly with the limited age range (2658-2781 Ma) present in a conglomerate of the underlying Ramsay Lake Formation, which was sampled by Ménard and Easton in 2017, 13 km to the east-southeast of the field trip stop (Ménard 2019). The population in the Ramsay Lake Formation sample is almost identical to the 4 Matinenda Formation samples studied by Easton and Heaman (2011), only a few hundred metres to the south of the conglomerate.



Photo 8. Crossbedding in Mississagi Formation sandstone at Stop 4.

Return to vehicles, continue north on Highway 108 for 1.4 km.

4.1 km (2.56 mi). Pull over on the right shoulder of the passing lane part way up the hill. This is Stop 2.4 of Young (1991); Stop 17 of Bennett et al. (1997) and Stop 1.4 of Bennett (2006).

STOP 5. Nipissing gabbro, altered Mississagi Formation, Bruce Formation

UTM co-ordinates 371652E, 5140948N

We will start by examining the low outcrops on the west side of Highway 108. ***Please exercise caution when crossing the highway.***

Low, rounded outcrops at the south end are Mississagi Formation sandstones similar to those we saw at Stop 4, but slightly pink in colour, likely due to the formation of albite by hydrothermal fluids from the adjacent Nipissing gabbro intrusion (east roadcut). After an outcrop gap, conglomerate

of the Bruce Formation is well-exposed. In particular note a clast with a gneissic fabric, as well as numerous, subrounded to subangular, white, granitoid clasts typical of the formation (Photo 9). The abundant matrix is dark-grey to black. Large, glassy, quartz grains are abundant on fresh surfaces of the conglomerate, another feature typical of the formation. The glassy, black, appearance results from the dark matrix behind the clear quartz (Bennett 2006).



Photo 9. Ramsay Lake Formation conglomerate at Stop 4. Note a gneissic clast just below the rusty spot at the centre of the photo. Knife is 9 cm long.

A sill-like body of Nipissing gabbro is exposed at the south end of the roadcut on the east side of the highway. Rhythmic, compositional layering is visible on the vertical face of the roadcut. Additional evidence of hydrothermal activity along the intrusion contact is seen by dark-green to black chlorite deposited along fractures in the gabbro (Bennett 2006). Near the north end of the Nipissing outcrop face, note the relatively planar, striated, surface is truncated by more irregular surface that has been interpreted as the result of erosion by high-pressure, waterborne sediment, presumably by the melting of an adjacent Pleistocene ice sheet (Bennett 2006).

North of gabbro sill, the upper portion of the Mississagi Formation is exposed along the east side of the highway and is also pinkish.

A few tens of metres northward, the Mississagi Formation is overlain by diamictite of the Bruce Formation (UTM 371668E, 5141245N). The dispersed megaclasts in the Bruce Formation are predominantly grey granitic rocks with smaller mafic clasts, predominantly Thessalon Formation volcanic rocks. At this locale, there is no evidence of a significant disconformity at the base of the Bruce Formation.

Return to vehicles and proceed approximately 300 m to near the top of the hill and park on the right shoulder.

4.4 km (2.75 mi). Roadcuts are present on both sides of the highway. This is Stop 18 of Bennett et al. (1997) and Stop 1.5 of Bennett (2006).

STOP 6. Espanola Formation, Quirke Lake Group and Nipissing gabbro sills

The base of the Espanola Formation is a green, laminated unit about a metre or so thick. Laminated silty limestones and minor thin, chert beds of the limestone member of the Espanola Formation (Photo 10), overlie the green unit. At this location, the proximity of Nipissing gabbro sills (Photo 11) has led to the development of calc-silicate minerals including: grossular garnet, diopside, idocrase (vesuvianite), and wollastonite typical of a skarn (Robertson 1968; Bennett 2006). Wollastonite (identified by X-ray diffraction) is found just below the north-dipping gabbro sill near the north end of the exposure, where it occurs as sub-parallel groups of pale grey to white prismatic crystals about 1 mm wide and up to a cm long. The pink coating on joint surfaces is apophyllite ($KFCa_4[Si_8O_{20}]_8H_20$) an uncommon mineral (identified by X-ray diffraction), sometimes found in amygdules in basalts, but which is also

associated with calc-silicates. Young (1991) states that the small scale thrust faults and folds in the limestone on the west side of the highway are probably the result of slumping during early tectonic activity.

The upper, ferruginous dolostone-bearing member of the Espanola Formation and the overlying Serpent Formation are not present at this location but are well represented on the north limb of the syncline. Young (1991) and Bennett (2006) both suggested that the missing Serpent Formation and ferruginous dolostone member were removed during a period of pre-Gowganda Formation erosion, which are visible in roadcuts along the next stretch of the highway. Alternatively, a fault could be present in the linear valley at the end of the outcrop, which has truncated stratigraphy.

Return to vehicles and continue north on Highway 108.



Photo 10. Espanola Formation at Stop 5. Note white recessive weathering limestone beds and thinly laminated, darker, calc-silicate and mudstone beds. Hammer handle is 33 cm long.



Photo 11. Nipissing sill (dark) in contact with Espanola Formation marbles (white) at Stop 5. Note discoloration of the marbles near the sill contact. View west across Highway 108.

4.5-5.0 km (2.81-3.13 mi). Diamictite and minor sandstone of the Gowganda Formation are exposed in a near-continuous roadcut along Highway 108. In these exposures, megaclasts of pink granite, grey granite and granitic gneiss and mafic rocks are widely distributed in a dark green matrix. A typical example can be seen at 4.8 km. Most geologists now consider at least some of the diamictites in the Gowganda Formation to be tillites, although a debris-flow origin, either glaciogenic or as submarine debris flows, is a more reasonable interpretation at specific localities. Roscoe (1969) places the appearance of free oxygen in the atmosphere ("oxyatmoversion") as coinciding with the appearance of the reddish hue of hematite just above the base of the Gowganda Formation.

6.5 km (4.06 mi). The stop is at a large roadcut at the top of a hill, near a communication tower. Park at the south end of the roadcut on the east side of the highway. This is Stop 2.5 of Young (1991), Stop 19 of Bennett et al. (1997) and Stop 1.6 of Bennett (2006).

This is an impressive exposure though a stratified sequence of diamictites, clast-supported conglomerates and sandstones of the Gowganda Formation.

STOP 7. Stratified Gowganda Formation, Cobalt Group

UTM co-ordinates 371841E, 5143180N

At the south end of the roadcut on the west side of the highway, massive diamictite is exposed. It is overlain by about 50 cm of laminated mudstone with dropstones, in turn overlain by a thick succession of lenticular beds of coarse pink and pink-grey arkosic sandstone interbedded with distinct beds of diamictite (Photo 12), pebbly sandstone and clast-supported polymictic conglomerate (Photo 13). Some conglomerate units display normal and reverse grading suggestive of debris flows. Some sandstones contain large clasts. Some clasts in the diamictite show striations, suggestive of a glacial origin. Clasts in the conglomerate are mainly well-rounded fragments, but some rip-up clasts of sandstone are also present. The rocks displayed here may be interpreted as debris and mass flows, possibly formed in an ice-proximal setting by resedimentation of glacial debris at a retreating glacial margin (Young 1991; Bennett 2006).

Note the predominance of red and pink granitic clasts, in marked contrast to the pale grey clasts of the Bruce Formation seen earlier at Stop 5. There is also a significant proportion of black pebble to cobble-sized clasts. The mineral assemblage and metamorphic grade of a few mafic clasts examined by G. Bennett many years ago indicated that the clasts were probably from Thessalon Formation basaltic flows (Bennett 2006).

Return to vehicles and continue north on Highway 108.

7.4 km (4.63 mi). Pink sandstone and diamictite of the Gowganda Formation.

9.4 km (5.88 mi). Diamictite with large boulder, Gowganda Formation.

12.0 km (7.5 mi). Stanrock Road. Reset odometer. Turn east onto Stanrock Road.



Photo 12. Gowganda Formation at Stop 7 showing pink-grey arkosic sandstone (lower) overlain by matrix- to clast-supported conglomerate. Hammer handle is 33 cm long.



Photo 13. Gowganda Formation matrix-supported conglomerate (lower) overlain by clast-supported conglomerate at Stop 7. Hammer handle is 33 cm long.

2.0 km (1.25 mi). park on the south side (right) of the road opposite a roadcut on the north side. This is Stop 2.6 of Young (1991) and Stop 1.7 of Bennett (2006).

STOP 8. Laminated varvite? Gowganda Formation, Cobalt Group

UTM co-ordinates 374911E, 5147007N

Laminated siltstone/mudstone of the Gowganda Formation on the north side of the road. This unit has been interpreted as being similar to the varves found as deposits in Pleistocene glacial lakes.

Continue east on Stanrock Road.

9.5 km (5.94 mi). Turn right onto Popeye Lake Road.

9.8 km (6.13 mi). Park and examine outcrops on the west side of the road.

STOP 9. Gowganda—Serpent Formation disconformity

UTM co-ordinates 381485E, 5144900N

The Serpent Formation consists of fine, well-sorted sandstone and siltstone, and is typically light grey. The exposure at this Stop shows typical sandstones of the formation (Photo 14), just below the disconformity with the overlying Gowganda Formation. The contact can be seen partway up the hill above the road level exposures of the Serpent Formation. The contact is sharp but irregular, and the Gowganda Formation consists of polymictic, matrix-supported, conglomerate

Return to vehicles, retrace route to Highway 108. Reset odometer to zero at the junction. Turn right onto Highway 108 and continue north.

3 km (1.9 mi). Denison Mine Road - Turn east. Reset odometer to 0.

1.5 km (0.95 mi). This is Stop 2.7 of Young (1991), Stop 20 of Bennett et al. (1997) and Stop 1.8 of Bennett (2006).



Photo 14. Serpent Formation. **Upper**. Crossbedding in sandstone at Stop 9. **Lower.** Indistinct, medium bedding in fine sandstone at Stop 9. Hammer handle is 33 cm long.

Optional STOP F. Gowganda—Serpent Formation disconformity

UTM co-ordinates 375102E, 5150499N

The Serpent Formation is not present in the south limb of the Quirke Lake Syncline and in the Blind River—Sault Ste. Marie area where it was probably removed during a period of pre-Gowganda Formation erosion. At this location, on the south side of the road, well-sorted sandstone of the Serpent Formation is overlain by polymictic conglomerate of the Gowganda Formation. The contact is sharp but irregular. Evidence of a sub-Gowganda Formation disconformity at this location is based on the presence of pebble and cobbles of the Serpent Formation near the base of the overlying Gowganda Formation.

Return to Highway 108. Reset odometer at Highway 108 and Denison Mine Road.

0.5 km (0.3 mi). Disseminated carbonate in sandstone of the Serpent Formation on the east side of Highway 108. The detrital zircon sample of the Serpent Formation reported by Rainbird and Davis (2006) and Craddock et al. (2013) came from this roadcut. The population ranged from 2549 to 3576 Ma, with the dominant population at Geon 27, but with 11 Geon 28 grains (*see* Table 2).

1.1 km (0.68 mi). Road to Quirke Lake and the former Panel Mine. Reset odometer to zero. Turn east onto Panel Mine Road.

0.6 km (0.38 mi) and 1.5 km (0.9 mi). Two outcrop areas are exposed on the west side of the road. The southern of the 2 areas is currently better-exposed. The second area to the north is Stop 2.8 of Young (1991), Stop 21 of Bennett et al. (1997) and Stop 1.9 of Bennett (2006).

STOP 10. Upper member of the Espanola Formation, Quirke Lake Group

UTM co-ordinates area 1, 374350E, 5151383N area 2, 375258E, 51511331N

Both areas exposure ferruginous dolomite and siltstone of the upper member of the Espanola Formation. At the first stop, ripple marks are visible on some bedding surfaces.

The upper member is the uppermost of the 3 members of the Espanola Formation recognized in the Elliot Lake area (Robertson 1968). It is characterized by intercalated siltstone and reddish-brown weathering, ferruginous dolostone beds containing 3-4% FeO. Intraformational breccia, ripple marks, small-scale crossbedding, and a variety of soft sediment features are present, but only faintly visible on the south outcrop (Photo 15). Near the east end of the second outcrop a grey clastic dike crosses stratification at a high angle.

Return to vehicles and continue east on Panel Mine Road.

4.1 km (2.5 mi). This is Stop 2.9 of Young (1991), Stop 22 of Bennett et al. (1997) and Stop 1.10 of Bennett (2006).



Photo 15. Dropstone in laminated dolostone of the Espanola Formation at Stop 10. Pen is 13 cm long.

STOP 11. Ramsay Lake Formation overlain by Pecors Formation

UTM co-ordinates 377379E, 5152019N

Diamictites of the Ramsay Lake Formation contain cobbles of grey granitic rocks, mafic clasts of Huronian Supergroup metavolcanic rocks and Archean felsic metavolcanic clasts in an abundant dark-grey to black sandy matrix. The Ramsay Lake Formation is overlain by dark laminated siltstone and mudstone of the Pecors Formation (Photo 16). The latter contains a few dropstones (Photo 17). Note: the Matinenda Formation of the Elliot Lake Group, expected between the basement and the Ramsay Lake Formation, is truncated by the Ramsay Lake Formation in this area. The Matinenda Formation does occur in the mine workings down-dip from this location.

If you continue east to the end of the Panel Mine Road, you enter the rehabilitated area of the former Panel Mine. There is little evidence of the uranium mine and mill complex that was on this site until 1993.



Photo 16. Laminated mudstone of the Pecors Formation, Stop 11. Hammer handle is 33 cm long.



Photo 17. Laminated mudstone of the Pecors Formation containing a dropstone at Stop 11. Knife is 9 cm long.

Return to vehicles, retrace route back to Highway 108. Reset odometer to zero at highway. Turn right and continue north on Highway 108. Tailings dam of the Quirke Mine is visible west of the Highway.

0.8 km (0.5 mi). Highway 108 ends and Highway 639 begins.

1.0 km (0.6 mi). Diamictite of the Bruce Formation is exposed on the west side of the highway.

1.5 km (0.9 mi). Outcrops of Mississagi Formation are exposed along Highway 639. Note the yellowish colour characteristic of the Mississagi Formation where it lies directly on the Archean granitic basement (Robertson 1968).

1.9 km (1.19 mi). Near the top of the hill a fresh roadcut on the west side of Highway 639 exposes greenish, coarse-grained sandstone of the Matinenda Formation (UTM 372210E 5152225N. We are near the top of the floater-reef zone, so slightly higher stratigraphically than at Stop 3.

2.5 km (1.6 mi). This is Stop 23 of Bennett et al. (1997) and Stop 1.11 of Bennett (2006). We will not visit this stop due to logistical and accessibility reasons.

Optional STOP G. Huronian Supergroup volcanic rocks of the Thessalon Formation

UTM co-ordinates 371508E, 5152302N

A gated road leads west from Highway 639 to one of the Quirke Mine tailings dams. Park near the gate and walk a short distance along a rough road from the gate to the base of the tailings dam. Note the very dark green to black, flattened, chlorite amygdules characteristic of the Huronian Supergroup mafic volcanic rocks between Sault Ste. Marie and Elliot Lake. Cross the stream and proceed northward a short distance along a rough road to the crest to the low hill. The Huronian Supergroup volcanic rock at this location (UTM 371428E, 5152342N) include hawaiite and mugearite (Bennett 2006). The eastward-trending, south dipping unconformity between the Archean granitic basement rocks and Huronian Supergroup volcanic rocks is visible near the crest of the hill. There appears to be no paleosol development at this location. Near the west end of the outcrop, a thin, quartz-pebble conglomerate or breccia unit, consisting mainly of angular, quartz-clasts, overlies the granitic rocks at the base of the volcanic unit. Scattered, isolated, mainly cobble-sized clasts of quartz are also present along the unconformity.

Some visitors to this site have proposed that the contact between the Archean and Huronian Supergroup volcanic rocks is not an unconformity, but a fault contact. During a visit to the Stanleigh Mine in 1990, however, G. Bennett observed identical scattered, quartz pebbles along the contact of the Huronian Supergroup volcanic rocks

where they overlie Archean mafic volcanic rocks in a haulage drift on the south limb of the Quirke Lake Syncline (Bennett 2006). Bennett (2006) proposed that these quartz cobbles are lag deposits left behind when the finer sediment was washed off the surface. A few kilometres west of this stop, occurrences of this conglomerate unit contain more rounded quartz grains, and locally are overlain by a thin arkosic sandstone.

Return to vehicles and continue north on Highway 639.

2.6 to 10 km (1.63-6.25 mi). Highway 639 passes through Archean granitoid rocks cut by Matachewan and other diabase dikes before entering into dominantly mafic volcanic rocks of the Ompa Lake greenstone belt which has similar ages to the Whisky Lake greenstone belt, namely *circa* 2685 Ma (*see* summary in Easton 2010).

9.3 km (5.81 mi). This is Stop 24 of Bennett et al. (1997) and Stop 1.12 of Bennett (2006).

Optional STOP H. Pillowed Archean metavolcanic rocks

UTM co-ordinates 368530E, 515773N

Archean mafic metavolcanic rocks with well-developed pillow structures are exposed, on a north-sloping outcrop, on the east side of the highway. The pillows are deformed, however, facing directions can easily be determined. Small amygdules are concentrated near the upper surface of many pillows. Lichen growth since 2006 has rendered this stop less spectacular than as indicated in Photo 18.

Return to vehicles and continue north on Highway 639.

10.2 km (6.38 mi). Flack Lake fault occupies a valley near this point.

10.7 km (6.69 mi). Outcrops of hematite-stained sandstone of the Bar River Formation.



Photo 18. Pillow structures in Archean mafic metavolcanic rocks at Optional Stop H. Scale card is 10 cm long. Photo from Bennett (2006, p.40).

11.3 km (7.06 mi). Entrance to Mississagi Provincial Park.

11.6 km (7.25 mi). Large roadcut on the east side of the road. Park on the shoulder. This is an added stop.

STOP 12. Bar River Formation, Cobalt Group

UTM coordinates 367555E, 5159855N

Thin to medium bedded, pale grey sandstone of the Bar River Formation with herringbone crossbedding. Return to vehicles and continue north on Highway 639.

12.3 km (7.69 mi). Jim Christ Lake to the northeast (formerly Christman Lake). Park on the right shoulder beside a low ridge of partially blasted roadcuts on the north side of the highway. This is Stop 25 of Bennett et al. (1997) and Stop 1.13 of Bennett (2006). This once spectacular roadcut has suffered from needless road construction damage in recent years.

STOP 13. Bar River Formation, Cobalt Group

UTM co-ordinates 366742E, 5160652N

Sandstones and mudstones of the Bar River Formation at this stop display ripple marks, mud cracks and sinuous structures, which have been described as possible worm casts. Comparison with desiccation structures in the Gordon Lake Formation led Young (1969) to suggest that these features are the result of the transportation of consolidated desiccation fracture fillings.

This outcrop area was sampled by Rainbird and Davis (2006) and Craddock et al. (2013) for detrital zircon geochronology. Population range was 2523-3074 Ma, with peaks at 2531, 2705 and 2726 Ma.

Return to vehicles and continue north on Highway 639.

16.1 km (10.01 mi). This is Stop 26 of Bennett et al. (1997) and Stop 1.14 of Bennett (2006).

STOP 14. Red beds of the Gordon Lake Formation, Cobalt Group

UTM co-ordinates 364395E, 5162758N

Laminated, maroon buff or green siltstone and mudstone, and minor chert, represent the upper part of the Gordon Lake Formation. Desiccation cracks and ripple marks are present, as are reduction spots in the maroon beds.

It was near this stop that Hill et al. (2018) collected a green siltstone sample which had a limited zircon population (27 grains), but with the 4 youngest grains giving an age of 2302±19 Ma, and with another cluster of 5 grains at 2364±16 Ma. These ages, as well as those of Rasmussen et al. (2016), suggest deposition occurred at *circa* 2300 Ma. Other populations were 6 grains at 2525±15 Ma, with older grains ranging from 2674 to 3158, but dominated by Geon 27 grains (Hill et al. 2018).

Return to vehicles and continue north.

17.4 km (10.88 mi). Park on the shoulder of the road and examine outcrops on the east side of the road. This is Stop 27 of Bennett et al. (1997) and Stop 1.15 of Bennett (2006).

STOP 15. Gordon Lake Formation, Cobalt Group

UTM co-ordinates 363062E, 5163194N

Siltstones and sandstones of the Gordon Lake Formation display ripple marks, desiccation cracks, cross bedding and a late cleavage. Note: the presence of pyrite in contrast to hematitic nature of the Gordon Lake Formation near the top of the formation.

Return to vehicles and continue north on Highway 639.

19.6 km (12.25 mi). Park on the right shoulder roughly midway in a lengthy roadcut on a south-facing hill which has almost near continuous exposures of Nipissing gabbro. Examine outcrops on the east side of the road.



Photo 19. Mafic dike at Stop 16 showing sparse, small, plagioclase phenocrysts. Hammer handle is 33 cm long.

STOP 16. Nipissing gabbro and post-Nipissing dike

UTM co-ordinates 361850E, 5164825N

The medium-grained, slightly greenish gabbro is typical of the Nipissing intrusions in the Elliot Lake area. At the stop, the gabbro is cut by a 3 m wide, near-vertical, sharp-walled fine-grained mafic dike that is east-trending. Small plagioclase phenocrysts occur throughout the dike (Photo 19).

Similar dikes have been observed elsewhere in the Elliot Lake area (Easton 2013a, 2013b), but have yet to be assigned to any specific dike swarm. It is possible they could be related to the *circa* 1750 Ma Trap dike swarm, or the 2125-2105 Ma Marathon dike swarm. The senior author attempted to obtain an age on the dike, but no suitable phases were recovered for geochronology. Hunt and Roddick (1987) reported a K-Ar whole rock age of 1325 Ma for this dike.

Return to vehicles and continue north on Highway 639.

21.8 km (13.63 mi). This is Stop 28 of Bennett et al. (1997) and Stop 1.16 of Bennett (2006).

STOP 17. Lorrain Formation, Cobalt Group

UTM co-ordinates 361771E, 5166967N

White to pale pink quartz arenite of the upper Lorrain Formation is exposed on the east side of the highway. The detrital zircon sample reported by Rainbird and Davis (2006) and Craddock et al. (2013) came from this Stop. Here, Geon 27 zircons are twice as abundant as Geon 26 zircons, and there are also several Geon 28 zircons present (*see* Table 2).

23.5 km (15.69). Junction Highway 639 and 546 at Little White River Road. End of Field Trip. Retrace route back to Elliot Lake and return to Sudbury.

End of road log.

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