



INSTITUTE ON LAKE SUPERIOR GEOLOGY
Hibbing, Minnesota
May 14-17, 2014



ANNUAL MEETING

PROCEEDINGS VOLUME 60
PART 2 – FIELD TRIP GUIDEBOOK



INSTITUTE ON LAKE SUPERIOR GEOLOGY

60TH ANNUAL MEETING
May 14-17, 2014
Hibbing, Minnesota

Sponsored by

PRECAMBRIAN RESEARCH CENTER, UNIVERSITY OF MINNESOTA DULUTH,
and
MINNESOTA GEOLOGICAL SURVEY

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Co-Chairs

Proceedings Volume 60

Part 2 – Field Trip Guidebook

Edited by Mark Jirsa, Terrence Boerboom, and Amy Radakovich, Minnesota Geological Survey

Cover Photos

Historic photo of steam shovel was acquired from the archives of Minnesota Discovery Center Research Library; all other photos were taken by Jirsa. Photos vaguely represent the subject matter of some field trips described in this guidebook (*relevant trip numbers are shown in parentheses*). Photos in large “6” include conjugate faults (1) in oxidized Biwabik Iron Formation at Susquehanna Mine; glacial till in Albany Mine (6); taconite pellets (3); folded Soudan Iron Formation (2); gray-green stromatolites photographed from polished sample in collection of Dan England, Eveleth Fee Office (1); and jointed, oxidized Biwabik Iron Formation from Glenn Mine (E). Photos in large “0” include taconite-bearing drill core (A); red stromatolites from Dan England collection (1); pillowed metabasalt from near Gilbert (7); and historic photo of steam shovel, location and date unknown (B, C).

Generalized geologic map showing locations of field trips in this guidebook

Note that no Trip 4 is shown, as it was cancelled

Geology simplified from Minnesota Geological Survey Map S-21 statewide bedrock geology

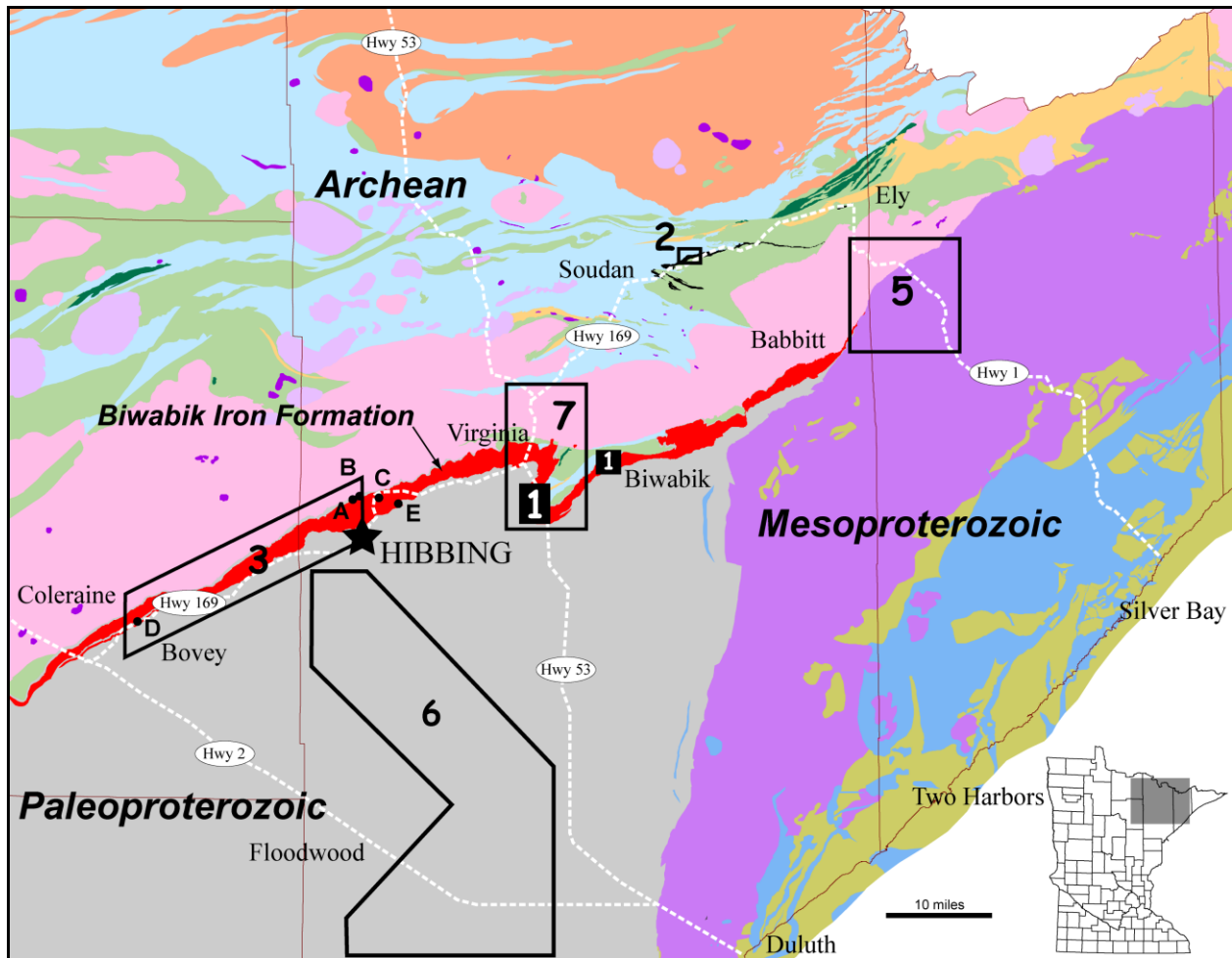


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The editors extend sincere thanks to all who contributed to this field trip guidebook. The time and effort expended to prepare field trip descriptions are greatly appreciated. Special thanks to **Minnesota Coaches Voyageur Bus Company** in Duluth for substantially discounting transport costs, and to **Greyhound Bus Museum** in Hibbing for providing a vintage bus for field trip B.

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

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Some figures in this field trip guidebook were submitted by authors in color, but are printed grayscale to conserve printing costs. Full color imagery will appear in the digital version of the guide when it is available on-line at <http://www.lakesuperiorgeology.org>.

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FIELD TRIP 1

Wednesday, May 14, 2014

STRATIGRAPHY, SEDIMENTOLOGY, STRUCTURE AND MINERALIZATION OF THE BIWABIK IRON FORMATION, CENTRAL MESABI IRON RANGE

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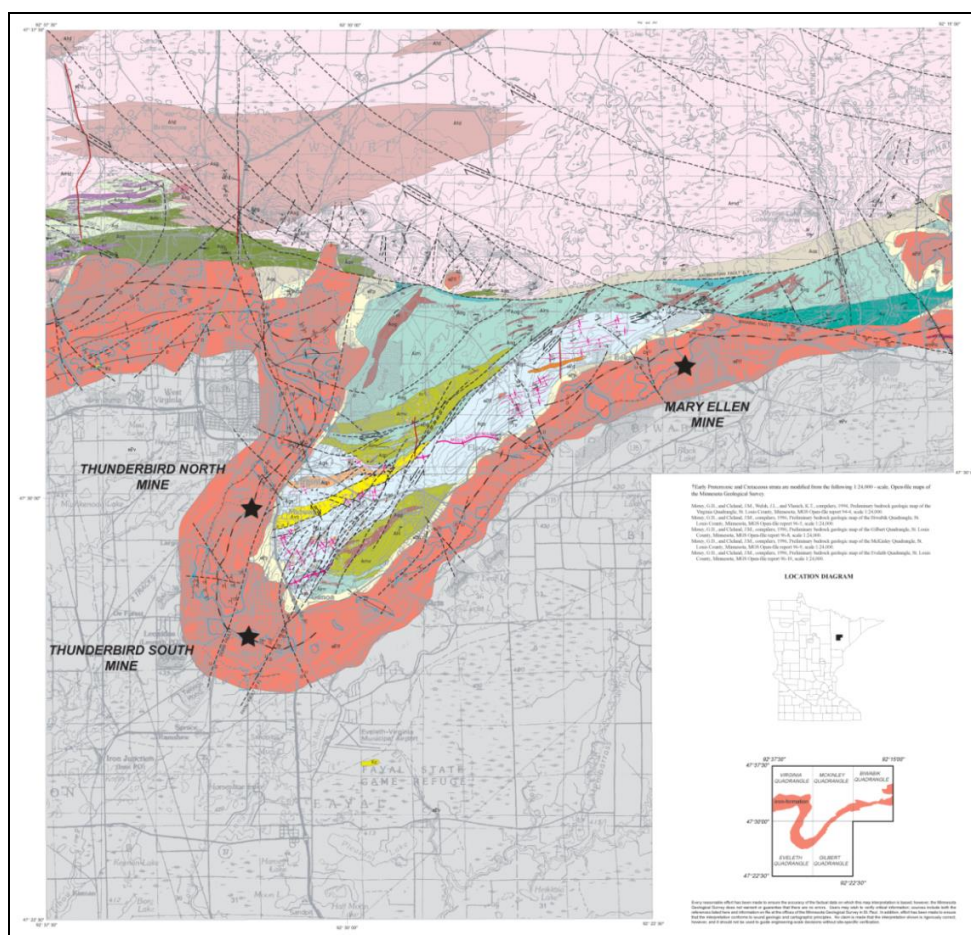


Figure 1. Bedrock geologic map of the central Mesabi Iron Range showing 3 main field trip localities. Stops 1-4 are located on United Taconite's Thunderbird North and South Mines near Eveleth; stop 5 is the Mary Ellen Mine near Biwabik. Archean metavolcanic, metasedimentary, and granitic rocks are shades of green, blue, and pink, respectively; Paleoproterozoic Pokegama Quartzite is yellow; Biwabik Iron Formation is red; and Virginia Formation is gray (modified from Jirsa and others, 1998).

INTRODUCTION

This field trip explores the geology of the Paleoproterozoic Biwabik Iron Formation (BIF) in the central Mesabi Iron Range of northeastern Minnesota. The formation hosts iron-ore deposits that have been mined continuously for nearly 125 years, constituting the most economically significant mining district in the United States. This trip will visit exposures in 3 localities (Fig. 1): the Cliffs Natural Resources - United Taconite LLC Thunderbird North Mine in Eveleth, an active magnetite taconite mine; the inactive satellite Thunderbird South Mine; and the Mary Ellen Mine near Biwabik, a closed direct-shipping (hematite) ore mine. Because this field trip guide was compiled by a large number of co-leaders having different experiences and perspectives, it may contain some content that is repetitive or has minor inconsistencies.

The iron-bearing strata of the Biwabik Iron Formation were first noted in 1866 by Henry Eames, on what was to become the eastern end of the Mesabi Iron Range. Sporadic exploration for iron ore deposits began soon after, notably by Peter Mitchell and the Ontonagon Syndicate. Lack of infrastructure hampered exploration and development until 1884, when the Duluth and Iron Range Railroad to the Soudan Mine on the Vermilion Range was completed. The resulting exploration boom initially focused on the well-exposed, metamorphosed iron-formation near the contact with the Duluth Complex. It was not until 1890 that the Merritt Brothers re-directed their exploration focus to an area farther SE along the Mesabi range, correctly surmising that iron ore float located south of the Giants Range had been transported some distance by glaciation. Their discovery of soft, friable, high-grade iron ore at the future Mountain Iron Mine on November 16, 1890 revolutionized the global iron ore and steel industries. The first shipment of 4,245 tons in 1892 came at the crest of a wave of exploration that was to quickly discover and develop billions of tons of iron ore along the 175 km strike-length of the Biwabik Iron Formation. The Mesabi Iron Range rapidly developed into the largest iron mining district in the United States, a status it continues to hold. For decades after its discovery, it was the largest iron ore district on earth, accounting for nearly half of global production in the late 1940s.

The Biwabik Iron Formation ranges from 0.5 to 5.0 km width (0.25 to 3 miles) along a strike length of 175 km (100 miles) (Fig. 2). The formation, as much as 220 meters (750 feet) thick, generally dips gently to the southeast at angles of about 7° to 15°. Unweathered, unaltered iron-formation, colloquially known as “taconite,” contains about 30 percent iron and 45 percent silica, with the balance (2-10%) composed principally of MgO, CaO, and MnO. In numerous places along the length of the range, typically along joints, fractures, folds, and other structurally prepared zones, silica and other elements were leached under tropical weathering conditions, locally enriching the iron content to as much as 64 percent. So-called “natural” or direct-shipping ores dominated production through the Second World War. Depletion of higher grade reserves, combined with increasingly stringent quality requirements, led to a rapid conversion of the industry during the 1950s. Perfection of fine grinding, magnetic separation, and pelletizing technology (the “taconite process”) allowed for economic exploitation of pristine iron-formation. Between 1956 and 1977, eight taconite facilities with a combined capacity in excess of 54 million tons per year (mtpy) were brought to production. Production of beneficiated iron ores (gravity and taconite concentrates) exceeded that of direct-shipping ores in 1958, and by 1967, taconite concentrates accounted for over half of production. At this writing, six taconite (40 mtpy capacity), and three tailings recovery (3 mtpy capacity) facilities are in production (Fig. 3).

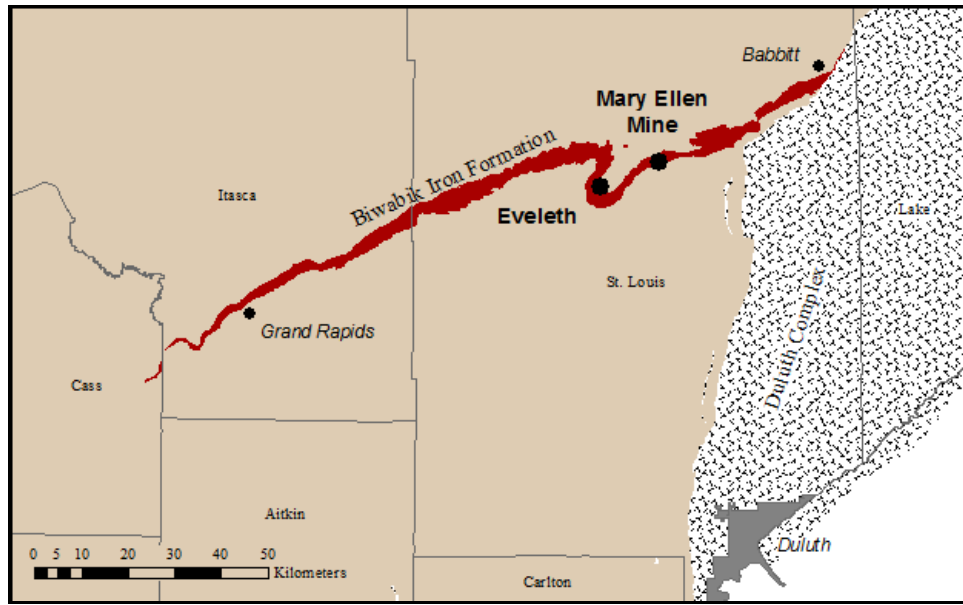


Figure 2. Location map of the Mesabi Iron Range (maroon). Note the Duluth Complex (Keweenawan, 1.1 Ga) on the east side).

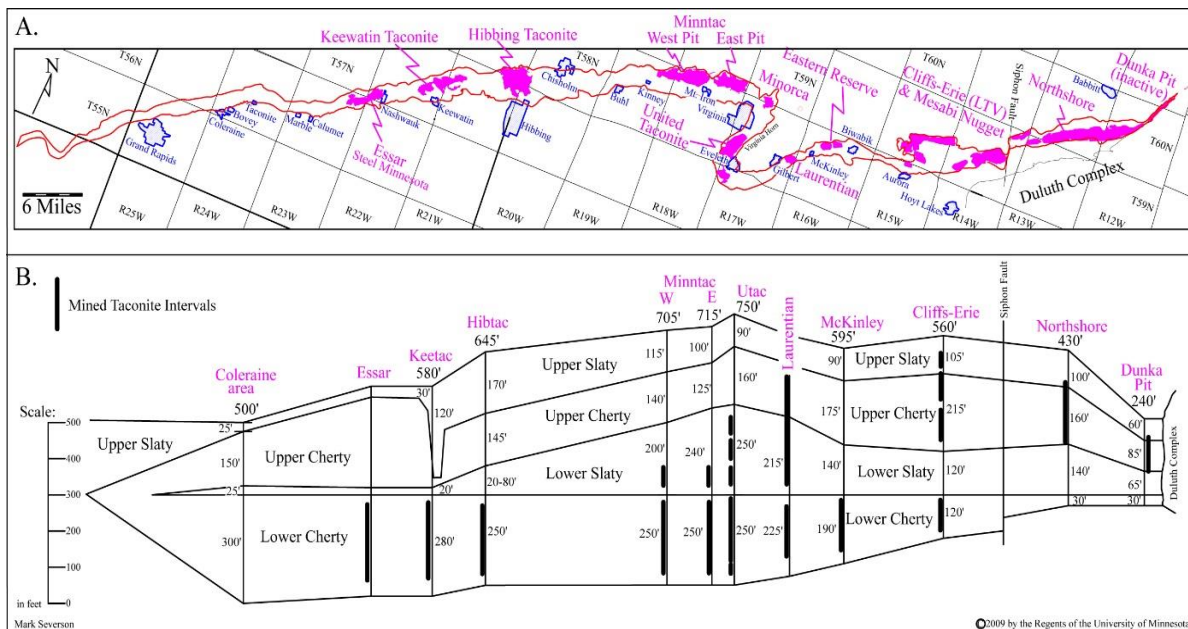


Figure 3. A) Aerial distribution of taconite pits and cities. B) A longitudinal section of the Biwabik Iron Formation showing: average thickness of the iron-formation at each taconite operation (along with the thickness of the various members at each operation), and mined taconite intervals (as black columns adjacent to the sections). From Severson and others (2009).

REGIONAL GEOLOGY

Basement rocks in the central Mesabi area consist of 2.7 Ga Archean granitic and greenstone terrains of the Superior craton. These were intruded by the 2.1 Ga Kenora-Kabetogama dike swarm, but were apparently eroded to a peneplained surface by about 1.9 Ga. The peneplained Superior craton formed the platform upon which a Paleoproterozoic continental shelf and margin assemblage was deposited. 1.88 Ga iron-formation and associated clastic sediments are preserved along 3000 km of strike length near the margin of the craton, and were once much more extensive. Available evidence indicates iron-formation accumulation occurred at all margins of the craton simultaneously, and is a reflection of global ocean chemical conditions rather than local basin geometry. Most of the iron-formation is preserved in mobile belts at the craton margin, and are thus variably deformed and metamorphosed (Trommald Iron Formation of east-central Minnesota, Gogebic Iron Formation of Wisconsin-Michigan, Negaunee Iron Formation of Michigan, Sokoman Iron Formation of Labrador and Quebec, Cape Smith belt of Quebec, Kipalu Iron Formation of Hudson Bay, Sutton Inlier iron-formation of Ontario). Relatively flat-lying, undeformed, and un- or weakly-metamorphosed iron-formation is locally preserved inboard of the craton margin (Temiscamie Iron Formation of Quebec, the Gunflint Iron Formation of Minnesota-Ontario, and the Biwabik Iron Formation). These were likely deposited on a clastic-starved platform, and indicate that an epeiric sea may have nearly or completely inundated the peneplained craton during the peak of iron-formation accumulation.

Paleoproterozoic sedimentation and iron-formation accumulation followed a general sequence of depositional events throughout the Superior craton. Nearshore, tidally influenced clastic sedimentation was succeeded by chemically-precipitated iron-formation. The transition from clastic to iron-formation is typically abrupt; however the presence of iron-rich, chert-cemented epiclastic strata indicate that iron and silica precipitation was occurring prior to significant accumulation of the epiclastic-poor iron-formation. Significant iron-formation accumulation was apparently triggered by a lack of epiclastic input, rather than abrupt onset of favorable iron-precipitating geochemical conditions. Iron-formation accumulation proceeded at very low rates until resumption of clastic deposition in the basin. Available evidence indicate that the Biwabik Iron Formation records accumulation over as many as 15 million years within a significantly larger time span (Larson, 2013). Iron-formation accumulation across the craton was likely diachronous within this time span, as significant internal unconformities are evident within individual iron-formations, and correlations between individual iron-formations are problematic. Iron-formation accumulation proceeded until resumption of clastic deposition in the basin; similar to the basal epiclastic strata, significant amounts strata containing iron-rich precipitates are found in the overlying epiclastic units.

In the Animikie Basin, the basal, near-shore, epiclastic sequence is represented by the Pokegama Formation, which was succeeded by deposition of the Biwabik Iron Formation. In excess of 200m of iron-formation accumulation was abruptly terminated at the contact with the overlying argillites, siltstones, and greywackes of the Virginia Formation. Within the regional Paleoproterozoic depositional system, this contact is also marked by chaotic (paleoseismic) deformation and deposition of ejecta related to the 1.85 Ga Sudbury meteorite impact event. (Jirsa and others, 2011). In the central Mesabi area, a conglomeratic bed containing angular argillite, chert, and carbonate clasts within the Upper Slaty member Dolomite/Limestone unit (submember US-2; see *Stratigraphy* section below) (Severson and others, 2009) has been correlated with the Sudbury meteorite impact event (Addison et al., 2005). Turbiditic sediment of the overlying Virginia Formation was ultimately sourced from 1.87-1.83 Ga volcanic rocks, and represents the collision of an island arc with the southern margin of the Superior craton during the Penokean orogen. Collision and sedimentation at the continental margin led to development of a foreland basin and the thick turbidite sequence.

Paleoproterozoic sedimentary rocks (including the Animikie Group) along the southern margin of the Superior craton were bisected by the 1.1 Ga Midcontinent Rift, a 2000km (1200 mile)-long rift system extending in an arcuate fashion from northeastern Kansas to southeastern Michigan. The rift consists predominantly of mafic flows and intrusions overlain by rift-fill sedimentary strata. In addition, numerous

mafic dikes, chonoliths, and other small intrusions were emplaced locally into rocks of the Animikie Group rocks at significant distances from the rift axis. These include mafic dikes cross-cutting the Biwabik Iron Formation near Keewatin, and a series of sills emplaced into iron-formation in the vicinity of Aurora. However, no such intrusions are known from the Central Mesabi area. Significant thermal metamorphism of iron-formation is limited to the area generally east of Aurora, within the aureole of the Duluth Complex.

By the end of the Cretaceous, peneplanation produced topography similar to that of modern day. The central Mesabi area lay close to the eastern extent of the Cretaceous Interior Seaway, and the Biwabik Iron Formation locally is overlain by near-shore shale and sandstone. A basal iron-ore-bearing conglomerate is locally present, indicating extensive formation of supergene-enriched, direct-shipping ores predated Cretaceous sedimentation. The roughly coeval formation of secondary iron oxides and hydroxides implies that supergene enrichment may have occurred under a tropical climate during the Cretaceous (Symons, 1966; Purucker, 1973). During this and subsequent weathering, it is likely that the iron-formation served as a geochemically resistant “cap” that protected the underlying Giants Range Batholith, and is indirectly responsible for formation of the *Missabe Wachu* (“Big Man Hills” from the classic David D. Owen’s 1852 Report of a Geological Survey of Wisconsin, Iowa, and Minnesota). The area is now known physiographically as the Giants Range.

Ice sheets advanced across the Biwabik Iron Formation multiple times during the course of the Pleistocene, primarily during the last 2 million years. Unconsolidated saprolite, including the supergene enriched direct-shipping ore (DSO), was preferentially eroded, leaving only remnants in deep, structure-hosted, trough-shaped bodies and stratiform layers variably protected by resistant cap rocks. Locally, large blocks of weathered iron-formation and even DSO were eroded by glaciotectonic activity. A variable thickness of till, outwash, and glaciofluvial sediment was deposited over the iron-formation during the final glacial cycle (See related Field Trip in this guidebook).

Archean Rocks

The Neoarchean bedrock of the central Mesabi Iron Range lies near the southern edge of the Wawa subprovince of the Superior Province, and constitutes the southwestern-most exposures of the terrane. The Archean supracrustal rocks on the Mesabi Range are separated from the well-known Vermilion district to the north by the Giants Range batholith, a large, composite body consisting of granitoid rocks of several generations and compositions. The Archean rocks are covered to the south, east, and west by Paleoproterozoic strata, including iron-formation of the Mesabi Iron Range. The Archean supracrustal rocks are subdivided into northern and southern panels on the basis of contrasting metamorphic grade and deformation style (Fig. 4). The northern panel, adjacent to the Giants Range batholith, contains intensely lineated, amphibolite-grade schist of volcanic, intrusive, and clastic protolith. The southern panel contains a similar stratigraphic sequence, but has minerals that indicate it underwent metamorphism to much lower grades, ranging from prehnite-pumpellyite to low greenschist facies. The two panels are separated by the east-trending, post-metamorphic, Laurentian fault. Amphibolite-grade rocks of the northern panel (north of the Laurentian fault) comprise the Minntac sequence that contains locally strongly layered schists having geochemical and outcrop-scale characteristics of volcanic, intrusive, and turbiditic protoliths. The lower grade strata within the southern panel are subdivided into the Mud Lake and Midway sequences. The Mud Lake sequence forms a broad, southwest plunging syncline (the Mud Lake syncline) defined by outer limbs of calc-alkalic and tholeiitic strata, and cored by graywacke, slate, and minor felsic tuff. The Mud Lake strata are unconformably overlain by, and locally lie in fault contact with, fluvial and alluvial conglomerate, subaerially deposited trachyandesitic flows, and pyroclastic rocks that comprise the Midway sequence.

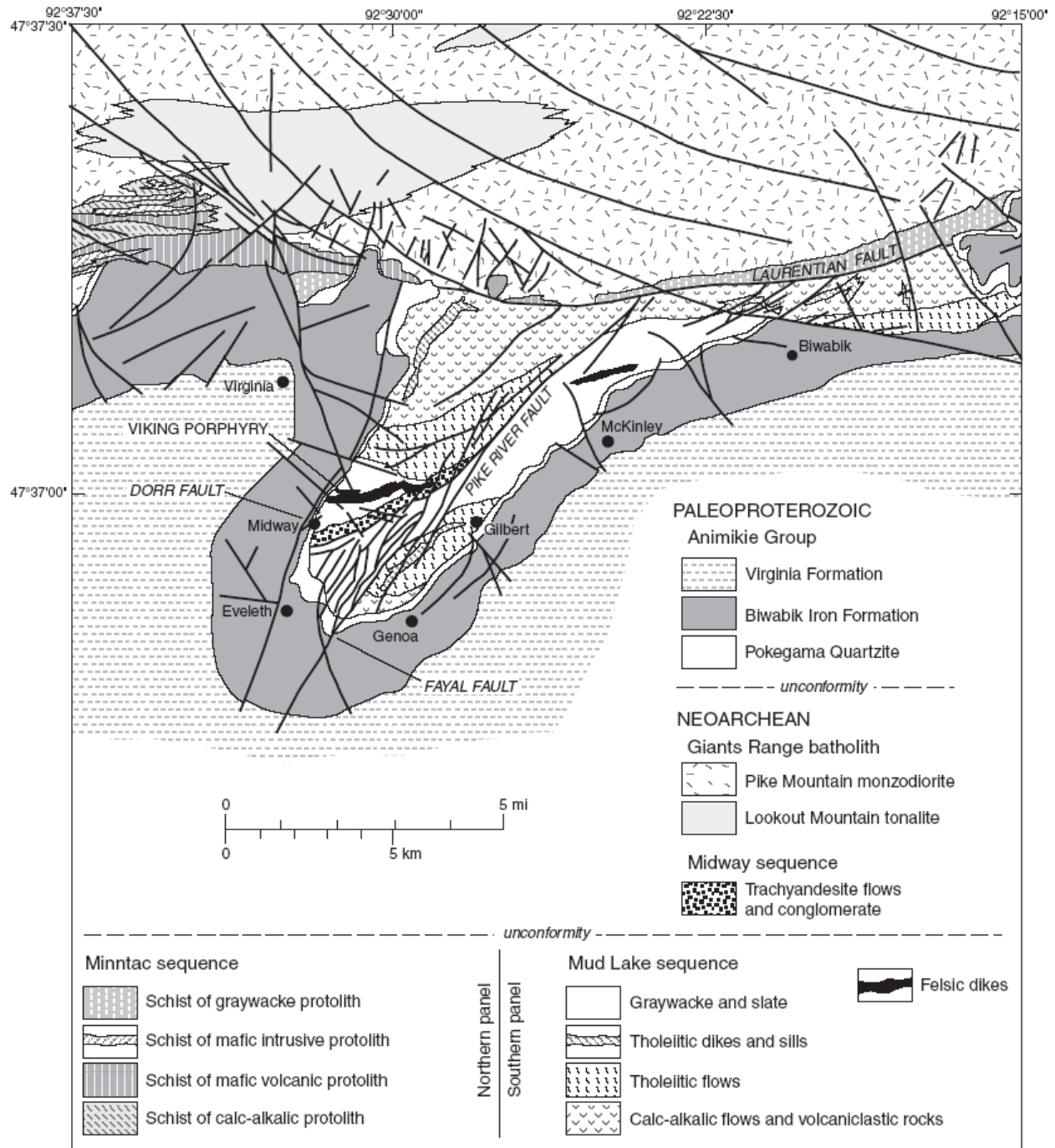


Figure 4. Geologic map of the central Mesabi range area, illustrating features of the Archean bedrock (From Jirsa and Boerboom, 2003). The Z-shaped fold/fault structure apparent in the strike of iron-formation is known locally as the “Virginia horn.”

Paleoproterozoic Animikie Group

The Paleoproterozoic Animikie Group unconformably overlies the Mille Lacs Group to the south in central Minnesota, and the Archean basement on the Mesabi Range to the north (Southwick and Morey, 1991). The Animikie Group consists of three major formations on both the Mesabi and equivalent Gunflint Iron Ranges. The respective units are the Pokegama Formation and the Kakabeka Quartzite (the lowest units), the Biwabik and Gunflint Iron Formations (the middle units) and the Virginia and Rove Formations (the upper units, composed of graywacke and shale). The Thomson Formation in the northern part of east-central Minnesota is inferred to be correlative with the Virginia and Rove Formations. The Biwabik and Gunflint Iron Formations are on strike with each other and were probably continuous prior to intrusion of the Duluth Complex at about 1.1 Ga.

Age

The age of the Animikie Group is relatively poorly constrained in the central Mesabi area due to a paucity of datable cross-cutting or intercalated units. A minimum age of deposition for the Pokegama Formation is $1,930 \pm 25$ Ma (Pb/Pb), which was obtained from quartz veins that cut the Pokegama Formation (Hemming and others, 1990). An age of $1,878 \pm 2$ Ma (U/Pb on euhedral zircons) has been obtained from an ash layer in the upper Gunflint Iron Formation (Fralick and others, 2002); this horizon may correlate with the Intermediate Slate of the Lower Slaty Member of the Biwabik Iron Formation (LS-1 submember). The ejecta layer correlated with the 1,850 Ma Sudbury impact event dates the stratigraphic top of iron-formation (Addison and others, 2005; Jirsa 2010). Zircon ages from an ash layer at the very base of the overlying Virginia Formation are dated at $1,832 \pm 3$ Ma (Addison and others, 2005). The latter sample was collected a few inches above the base of the Virginia Formation in drill hole VHD-00-1, located immediately to the west of the Thunderbird Mine. Vallini and others (2007) dated a metamorphic xenotime overgrowth on detrital zircon from the Pokegama Formation at 1763 ± 14 Ma, attributing this to a ~ 1786 Ma regional basin-wide, subtle thermal pulse.

Stratigraphy

In the central Mesabi area, the Animikie Group is composed of three formally defined formations: the Pokegama Formation, Biwabik Iron Formation, and Virginia Formation; and an informally named unit of breccia and ejecta related to the Sudbury meteorite impact event.

Pokegama Formation

The Pokegama Formation consists of up to 300' of shale, siltstone, and chert-cemented quartz arenite (quartzite). The formation consists mostly of siltstone and shale. Silica-cemented quartz arenite is confined to an interval a few meters thick beneath the overlying Biwabik Iron Formation. Ojakangas (1983) used gross stratigraphic relationships to subdivide the Pokegama Quartzite into three informal members: a basal member consisting dominantly of thinly bedded to laminated shale and lesser amounts of siltstone; a middle member consisting of shale and siltstone and scattered thin beds of quartz arenite; and an upper member consisting mostly of quartz arenite.

The formation was deposited on an irregular Archean bedrock surface. A basal conglomerate contains a poorly sorted array of clasts derived from the underlying bedrock set in a matrix of fine-grained sandstone to siltstone. Basal strata are marked by a second conglomerate that has angular to sub-rounded, pebble- and granule-size clasts of chert, jasper, algal fragments, and vein quartz. This implies that Pokegama-like clastic sedimentation and Biwabik-like chemical precipitation were, for a time, contemporaneous (Jirsa and Morey, 2003).

At most localities, the contact between the Pokegama Quartzite and the overlying Biwabik Iron Formation is conformable and gradational; the presence of beds of chert 6 to 12 meters beneath the Biwabik–Pokegama contact has been cited as evidence for a gradational sedimentary regime between the two formations (Ojakangas, 1983).

Biwabik Iron Formation

The Biwabik Iron Formation ranges from about 175-300 feet thick at the extreme eastern end of the Mesabi Iron Range (Dunka Pit) (Bonnichsen, 1968), to 730-780 feet thick in the Central Mesabi area near

Eveleth, decreasing to around 500 feet thick on the western Mesabi Iron Range near Coleraine, and eventually exhibits a “nebulous ending about 15 miles southwest of Grand Rapids” (Marsden and others, 1968). The formation is subdivided into four informal members (from bottom to top): Lower Cherty, Lower Slaty, Upper Cherty, and Upper Slaty (Wolff, 1917). Individual beds can be described as either sand-textured granular iron-formation (gif), composed predominantly of rounded oolitic grains and intraclasts, or mud-textured and laminated banded iron-formation (bif). Although interlayering of these two lithotypes occurs on all scales, the “cherty” members are composed predominantly of medium- to thick-bedded gif; the ‘slaty’ members are composed predominantly of thin-bedded bif. The terms “slaty” and “cherty” were originally used by miners, and are not indicative of metamorphism or slaty cleavage, or a predominance of silica as chert. The cherty gif members are largely composed of iron oxides and chert-cemented granules of iron silicates and carbonates. The slaty members are generally composed of laminated iron silicates and iron carbonates. The slaty bif members are envisioned to have been deposited below storm wave base. Granules comprising the cherty gif members formed in high-energy environments. Two models have been applied to explain formation of granules in the Paleoproterozoic iron-formations: direct precipitation, and reworking of intraclasts shoreward during storm events where they are reworked into granules in shallower water. Some granules appear to be the product of reworking of laminated bif material, however the self-similarity of granule sizes, lack of apparent intraclast source material, and geochemical dissimilarity between gif and bif material suggest the gif formed due to direct precipitation as oolites in shallow water.

A few diagnostic marker units within the formation allow basin-scale correlation. Two stromatolite-bearing intervals several feet thick are present, one at the base of the Lower Cherty member and the other in the middle of the Upper Cherty member (UC-6 submember). For submember terminology, see *Detailed Stratigraphy of Biwabik Iron Formation* discussion below. The black Intermediate Slate (LS-1 submember) at the base of the Lower Slaty member reportedly contains ash-fall tuff, with up to 5.5% Al_2O_3 (Morey, 1992). The top of the Upper Slaty member (US-2 submember) contains several feet of limestone and dolomite. Most of these marker units, which are prominent in the eastern and central parts of the range, pinch out in the vicinity of Hibbing, about 60 miles from the west end of the range (Severson and others, 2009). The Lower Slaty member is not present at the far western end of the range.

Sudbury Impact Layer

The contact between iron-formation and overlying slate of the Virginia Formation is marked by a thin layer of deformed substrate and overlying ejecta formed during the 1850 Ma Sudbury meteorite impact event. The horizon can only be seen in drill core on the Mesabi Iron Range; however, it is well exposed in the equivalent Gunflint Formation to the northeast. There, the deformed layer consists of $\leq 7\text{m}$ of chaotically folded and locally brecciated substrate iron-formation. At least some of the iron-formation beds were ductily deformed, implying they were not yet fully lithified at the time of deformation. The rheologic behavior of siliceous layers was more brittle, and they were dislodged and shattered. In the context of a major meteorite impact event, the deformation is inferred to have occurred by impact-induced seismicity (Jirsa and others, 2011). The deformed strata are draped by ejecta ($\leq 1\text{m}$ -thick) containing abundant petrographic evidence of impact origin, including the presence of zoned spherules and quartz fragments displaying multiple planar deformation features. The impact-related horizon, known as the Sudbury Impact Layer, is well exposed in several locations in the Lake Superior region (Fig. 5). It occurs in the Gunflint Lake area of northeast Minnesota (op. cit.), in the Thunder Bay area of adjacent Ontario (Addison and others, 2005), and in Michigan (Cannon and others, 2010; Pufahl and other, 2007).

At exposures near Thunder Bay and Gunflint Lake, the contact between the impact layer and the overlying Rove Formation (Virginia equivalent) is inferred to be a disconformity that may represent a significant depositional or erosional hiatus, perhaps as long as 15-40 million years (Jirsa and others, 2011). Whether this is also true for the contact on the Mesabi range is currently unclear. It is noteworthy that the depositional environment in which the Virginia and Rove Formations were deposited is nearly identical with that of the underlying iron-formation—the primary difference being a paucity of chemical precipitates (iron and silica) in the former.

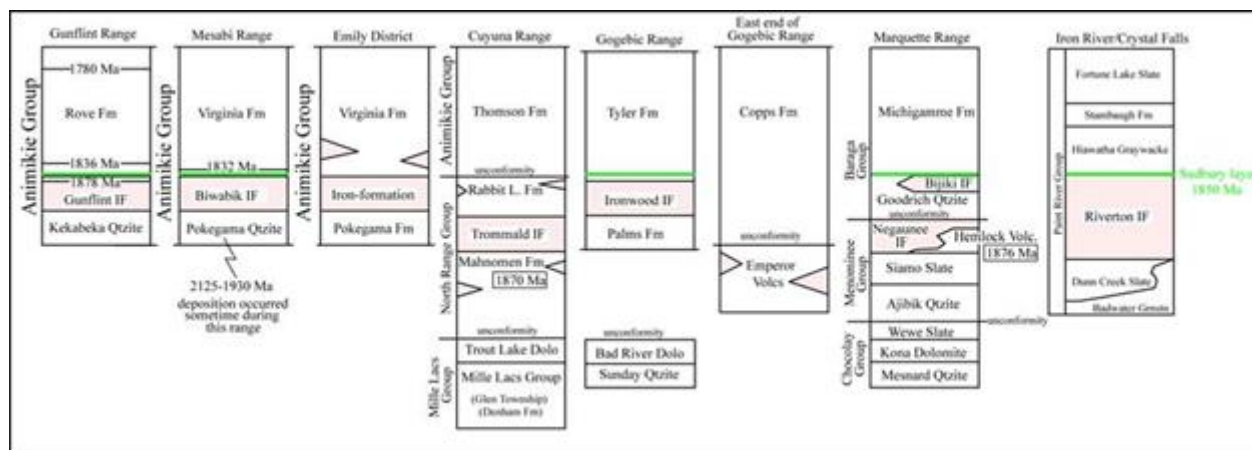


Figure 5. Generalized correlation diagram of Paleoproterozoic strata in the Lake Superior region. The position of the Sudbury Impact Layer is from Cannon and Addison (2007).

Virginia Formation

The Virginia Formation overlies the Biwabik Iron Formation and Sudbury Impact Layer on the south side of the Mesabi Iron Range, and is inferred to extend southward beneath glacial cover for an unknown distance. Presumably, it reappears in east-central Minnesota as a folded and metamorphosed sequence called the Thomson Formation. The Virginia Formation is a thick turbidite sequence composed of interbedded argillite, graywacke, and volcanoclastic rocks (Fig. 5). The 450 meters of Virginia Formation strata penetrated at a site south of Biwabik have been described in considerable detail in Lucente and Morey (1983). The lower part of the formation is composed almost entirely of alternating beds of dark-gray, silty mudstone and black carbonaceous shale. Quartz-rich siltstone and very fine- to fine-grained lithic graywacke become more abundant stratigraphically higher. The basal part of the Virginia Formation contains several beds of coarse-grained feldspathic graywacke and volcanoclastic rocks, as well as many lenses and irregular beds of limestone and dolomite. Dolomite-rich concretions of various sizes and shapes, also characterize the lower several hundred meters of the formation. Sandstone beds become coarser-grained and more abundant up-section, and are composed of angular quartz and feldspar grains in a matrix of muscovite and chlorite. Much of the matrix consists of diagenetically altered lithic fragments.

Hemming and others (1995) documented that shales of the Virginia Formation have Nd-depleted, mantle model ages of 2.35 to 2.14 Ga. Interbedded volcanoclastic sediments have younger model ages of 1.99 to 1.86 Ga. Craddock and others (2013) showed a dominant spectrum of Penokean orogeny ages (1.85-1.8 Ga) for detrital zircon populations from the correlative Thompson and Rove Formations. These data indicate Virginia Formation sediment was sourced from a comparatively young, differentiated volcanic arc, most likely from the Wisconsin Magmatic Terrane and equivalent rocks in Minnesota to the south.

Depositional Environments

The Animikie Group records a sedimentological transition from nearshore, tidally influenced, allochthonous clastic deposition, through shallow, autochthonous, chemical platform deposition, interrupted by impact-dominated deformation and deposition, and followed by deep water basinal turbidite sedimentation (Fig. 6).

The Pokegama Formation is interpreted to have been deposited in a tidally influenced, shallow marine setting near the shoreline, having received clastic detritus from the Archean basement to the north (Ojakangas, 1983; Craddock and others, 2013). In the central Mesabi area, the Pokegama Formation consists of a lower (argillaceous) member, a middle member of intercalated argillaceous and silty sedimentary strata, and an upper member of quartz sandstone. This succession is interpreted to represent a

transition from upper tidal flat to lower tidal flat/subtidal depositional environments (Ojakangas, 1983). Elsewhere along its strike length, the Pokegama contains pebble conglomerate that may represent a transgressive lag. The transition from near-shore, clastic-dominated sedimentation to autochthonous chemical sedimentation is recorded by an abrupt gradation into iron-formation. The abrupt decrease in clastic input is consistent with non-accretionary transgression across the peneplained craton, whereby relatively small rises in eustatic sea level translate into dramatic shifts in shoreline position.

In general, iron-formation can be geochemically divided into two components: an autochthonous chemically precipitated component, and an allochthonous clastic component. The elements comprising the autochthonous component (Fe, Si, Mg, Ca, Mn, P) are precipitated directly from seawater, while the allochthonous component (Al, Ti, K, Na) is derived from terrestrial sources (dust or suspended sediment). In general, the allochthonous component of the BIF averages 2.5%. The close correlation between lithofacies and mineralogy indicates that iron-formation mineralogy is a sensitive recorder of redox conditions at the sediment-water interface and, by extension, the depositional environment.

The Lower Cherty Member (LC) of the Biwabik Iron Formation is interpreted to have been deposited on a shallow marine shelf. The LC is dominated by granular iron-formation (gif), characterized by rounded oolitic grains, cross-bedding, and other features indicative of a high-energy environment. Accumulation was driven nearly entirely by autochthonous chemical precipitation, with very little epiclastic input. Ferric hydroxide apparently co-precipitated with carbonates, evidenced by an abundance of ankerite. The common presence of stylolites indicates significant volume loss of both iron- and carbonate minerals by chemical dissolution during compaction. The presence of herringbone cross-stratification in the lowermost LC indicates deposition in a tidally-influenced environment. The presence of thin laminae of slaty banded iron-formation interbedded with medium-bedded gif farther up-section within the LC suggests a transition to a deeper shelf environment. There, steady deposition of mud-textured banded iron-formation (bif) below the fair-weather wave base was periodically punctuated by rapid accumulation of gif, perhaps by shelf-to-basin transport during storm events.

The upper portion of the LC contains a thick-bedded, coarse-textured (with intraformational conglomerate), pink (oxidized) gif overlain by a medium-bedded, sand-textured, green (reduced) gif (LC-6/LC-7 contact) with abundant interbedded bif layers. This major lithostratigraphic break represents a significant disconformity within the overall BIF sequence. Trace element data show an abrupt transition from a high Al:Ti clastic source for LC strata below the break, to a low Al:Ti clastic source for all BIF strata above the break, indicating a change in sediment provenance.

The LC sequence above the LC6-LC-7 disconformity records a transition to deep water pelagic bif sedimentation, culminating in the Lower Slaty (LS) member Intermediate Slate (LS-1) submember. The Intermediate Slate/LS-1 is composed of iron silicates and sulfides in a laminated bif. The relatively large clastic content indicates a decrease in the rate of autochthonous chemical sediment accumulation, driven in part by redissolution of ferric hydroxide precipitates in anoxic bottom water.

The LS to Upper Cherty (UC) (LS-2 to UC-4) sequence is dominated by laminated bif, punctuated by lenticular or channel-shaped bodies of gif (Interbedded Cherts or IBCs). The bifs record deepwater sedimentation; the gifs consist of sand-textured granules that were exported from a high-energy environment and deposited in bypass channels on the shelf (channels) or deepwater fans (lenses).

The upper portion of the UC (submembers UC-5 to UC-7) is characterized by gifs deposited in a high-energy environment, as evidenced by abundant rounded oolitic grains, stromatolites, and cross bedding. Mineralogically, these submembers are characterized by the presence of significant amounts of primary hematite relative to the Biwabik Iron Formation as a whole, and the Algal Chert Horizon/UC-6 submember is significantly enriched in Mn, indicating deposition in oxic conditions. Combined, these features suggest deposition in shallow, tidally-influenced or subtidal environment. The high-energy environment, combined with a paucity of clastic sediment input, suggest this may have been analogous to a modern carbonate platform environment.

Chauvel and Dimroth (1974), noted similar features (chiefly oolitic and intraclastic sand) in the Sokoman Iron Formation, and their corresponding textural similarity to sediments in modern carbonate environments. They applied a carbonate facies model to the iron-formation, and attributed gif deposition

to a lagoonal platform depositional environment, characterized by rapidly shifting banks of oolitic and intraclastic sand, mud banks, and small, lagoonal basins ringed by oolite shoals. Sommers and others (2000) concluded that ooids and stromatolites from the Lower Algal Chert member of the Gunflint Formation (stratigraphic equivalent of the UC-6) were originally deposited as carbonates and later replaced by silica. Rankey and Reeders (2011) described how the interaction of wind, waves, and tides with channel morphology on platforms to create “Spin Cycle” currents which control formation, transport, grain size, sorting, and deposition of chemically precipitated oolites. Such currents are a viable mechanism for keeping precipitating granules frequently in suspension until they reached a critical size; this in turn provides an explanation for the remarkable self-similarity in granule size observed in gifs within the BIF (Patelke, in progress).

The sequence from UC to Upper Slaty (UC-8 to US-2) is dominated by laminated bifs, indicating a return to a deeper-water shelf environment. The US-2 is a limestone of enigmatic origin.

The BIF is overlain by argillite and graywacke of the Virginia Formation, deposited by northward-advancing basinal turbidite sedimentation. Some volcanic ash evidently settled into the basin, locally forming graded beds with a totally volcanic composition. The dominance of black, fissile shale suggests the "raining out" of clay and deposition in deep, anoxic water below the wave base. Minor, thin, sandstone lenses were deposited by bottom currents (Lucente and Morey, 1983).

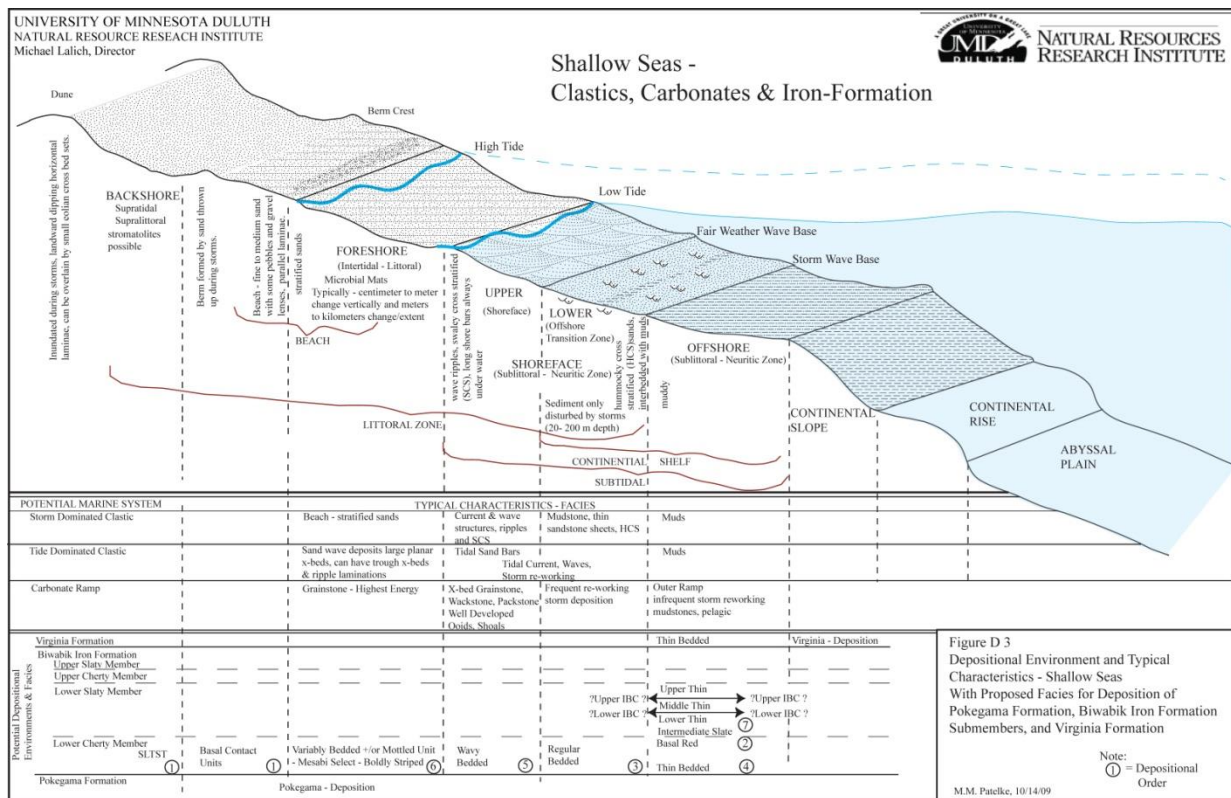


Figure 6: Possible siliciclastic environments for deposition of portions of the Biwabik Iron Formation in cross-sectional view (facies division in a shallow sea). Compiled, and drawn, by Marsha Patelke (in Severson and others, 2009) from sources including Boggs (2001), Leeder (1999), Nichols (1999), and Pratt and others (1992).

Mesoproterozoic

Despite being regionally extensive, no known Mesoproterozoic intrusive rocks are known in the Central Mesabi area. Similarly, the entire area of this field trip lies beyond the extent of thermal metamorphism of the Biwabik Iron Formation by the Duluth Complex.

Cretaceous

Cretaceous rocks are thought to be more or less continuous beneath glacial drift throughout the western half of the State, and form numerous outliers in the eastern half. On the Mesabi Range, Cretaceous rocks have been exposed by mining, and recognized in drill core and well cuttings (Fig. 7). The Cretaceous Coleraine Formation in the Mesabi district comprises iron-ore conglomerate, shale, and sandstone that form a thin irregular mantle over an uneven surface on the underlying bedrock (Sloan, 1964). The rocks are dominantly of marine origin in the west, but grade eastward in the central Mesabi to continental (fluvial and deltaic) sediments. In the western part of the district, fine conglomerate grades upward into a ferruginous grit and sandstone, including bluish-green shale in the vicinity of the Hill Annex mine near Calumet. The rocks grade eastward into continental sediments, including a widespread basal conglomerate composed of fragments of iron ore. The Cretaceous strata are virtually horizontal except locally, where they are interpreted to have slumped or to have compacted differentially over depressions in the underlying bedrock (Owens, 1956). Depressions and slumps were common topographic expressions of supergene-enriched natural iron ores, enabling the preservation of Cretaceous remnants post-glaciation.

In at least two notable central Mesabi locations, the Coleraine Formation conglomerates were mined as ore. At the Judson Mine near Buhl, eroded iron ore was transported short distances southward, and conglomerates were deposited in northwestward-trending channels cut in the underlying bedrock (Everett, 1956). Owens (1956) describes a 60-ft section of iron ore conglomerate, red and white shales, and lignite in the Enterprise Mine northeast of Virginia. The occurrence of Cretaceous iron-ore conglomerates have been confirmed just east of Gilbert (Sloan, 1964).



Figure 7. Panoramic view of the Enterprise Mine near Virginia; slump area in foreground and light-colored unit in background are Cretaceous sandstone (Owens, 1956).

Quaternary

The central Mesabi area was glaciated repeatedly during the course of the Pleistocene. Unconsolidated saprolite, including a significant amount of supergene enriched direct-shipping ore, was preferentially eroded, leaving only remnants in deep, structure-hosted trough-shaped bodies and in stratiform layers variably protected by resistant cap rock. Final ice retreat occurred about 13 C¹⁴ thousand years before present, when the margin of the Rainy Lobe retreated north of the Giant's Range, depositing a sandy-textured, boulder till in its wake. The St. Louis Sublobe advanced in a surge across the glacial lake to the south, reaching the toe of the Giant's Range, depositing a silty, boulder-poor till. A later glacial lake capped the low lying areas with silty glaciolacustrine sediment.

Glaciotectonism played a significant role in glacial erosion of direct-shipping ores. Large blocks of loose, porous oxidized and weathered iron-formation were frozen *en masse* onto the toe of the glacier, and thrust into the debris load. In the vicinity of the Fayal Mine in Eveleth, stripping operations in advance of a 'milling' mine encountered a block of direct-shipping ore entirely encased by glacial drift. A similar occurrence can be seen in the pit wall of the Glen Mine near Chisholm (*Field Trip E, this guide book*). The so-called Moose Track Mine produced in excess of 30,000 tons of ore, indicating the block contained over 10,000 m³ of material (Leith, 1903).

BIWABIK IRON FORMATION: STRATIGRAPHY, STRUCTURE, MINERALOGY, AND ORE DEPOSITS

Stratigraphy of the Biwabik Iron Formation

The four-fold stratigraphy of Lower Cherty, Lower Slaty, Upper Cherty, and Upper Slaty members (Wolff, 1917) is still used at each of the currently operating (and inactive) iron mines on the Mesabi Iron Range. However, each of the mining companies further subdivides the Biwabik Iron Formation into several submembers based on lithostratigraphic units and mineralogical assemblages observed in drill core and mine exposures (Fig. 8). District-wide correlation of individual mine stratigraphy is problematic because lithostratigraphic or mineralogically defined units important at the mine scale are not necessarily extensive at the district scale. Severson and others (2009) defined and correlated 25 laterally extensive submembers within the BIF based on examination of more than 380 drill holes along 75 miles strike length. Units were named for their characteristic bedding types. Definition of these 25 "Rosetta" units serve as a starting point for more detailed sequence stratigraphic studies and basin analysis (Fig. 9). There clearly are variations in some of the units and the four main iron-formation members along strike that are related to facies changes.

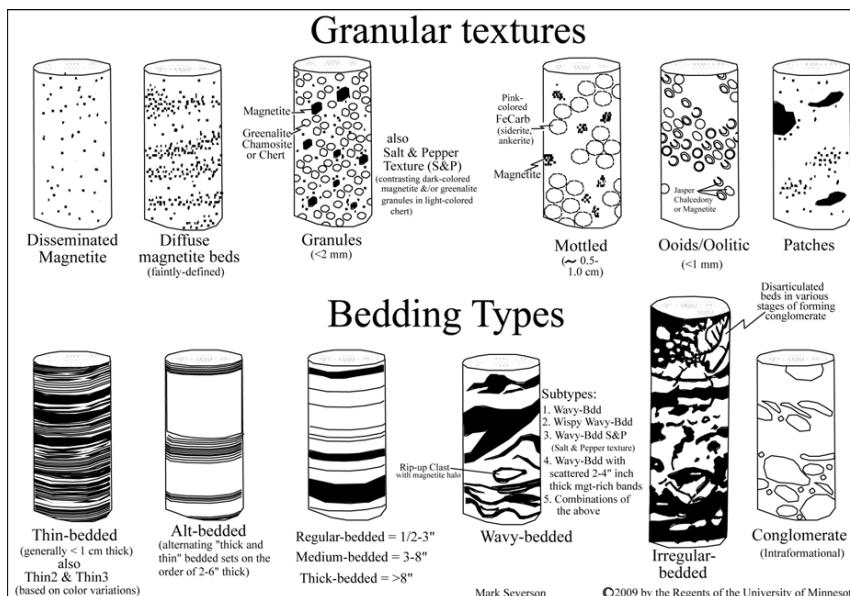


Figure 8. Textural characteristics of the Biwabik Iron Formation (from Severson and others, 2009 – modified from Pfleider and others, 1968). Dark bands represent magnetite-rich rock.

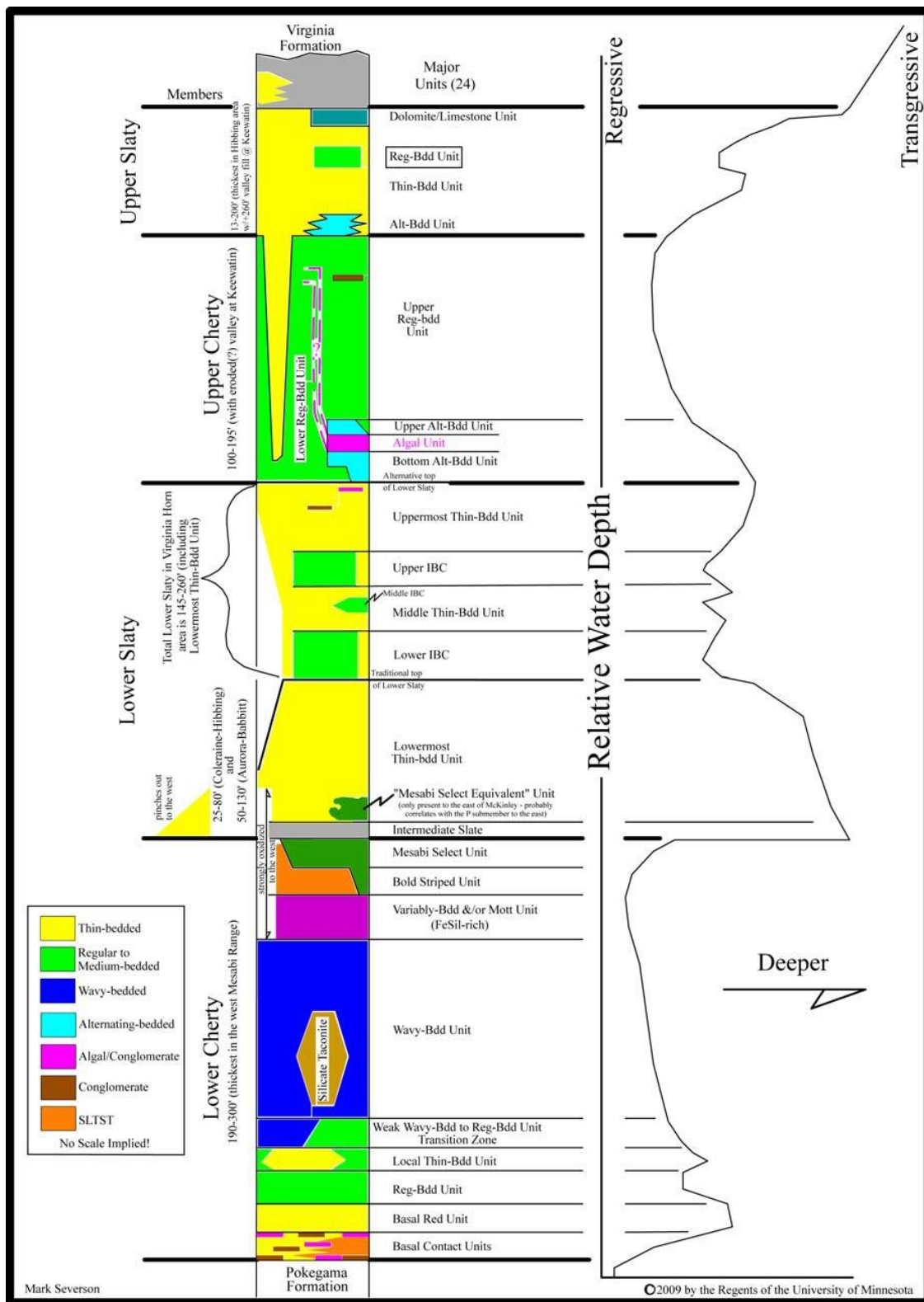


Figure 9. Graphic summary of the 25 major "Rosetta" units in the Biwabik Iron Formation that were identified and described in Severson and others (2009). Most of these units have corresponding submember designations at each of the taconite mines.

In addition to the member subdivisions recognized in the BIF throughout the Mesabi Iron Range, Thunderbird Mine geological staff further subdivide the iron-formation into 23 submembers based on lithologic, metallurgical, and mineralogical characteristics. These submembers form the basis for both resource estimation and grade control at the mine, and are described in detail below. For reference, the corresponding “Rosetta” unit name of Severson and others (2009) is listed in parentheses after the Thunderbird Mine submember name. The drilled thickness of the Biwabik Iron Formation in the vicinity of the Thunderbird North Deposit is approximately 686’ (Table 1).

Member	Thickness
Upper Slaty	51 feet (15.5 meters)
Upper Cherty	347 feet (121 meters)
Lower Slaty	52 feet (16 meters)
Lower Cherty	236 feet (72 meters)

Table 1. Average thicknesses of the four members, as recognized at Thunderbird North

Lower Cherty

The Lower Cherty member is approximately 236 feet thick in the Thunderbird North deposit. It has been subdivided into the following eight subunits:

LC-1 (Basal Red Unit)

The submember is a pink-green-gray heterogeneous unit comprised of interbedded thin-bedded slaty and thin-bedded cherty carbonate-silicate(minnesotaite-talc-stilpnomelane) iron-formation. LC-1 comprises the basal 64 feet of the iron-formation. It is defined as the footwall thickness of the iron formation, the magnetite grade of which is subeconomic. It is in general poorly described since the majority of exploration and development drilling terminates in the upper few feet of this unit.

LC-2 (Regular-Bedded Unit)

The submember is a gray thin-bedded cherty carbonate silicate(minnesotaite-talc) iron-formation. Magnetite occurs as disseminated and diffuse idiomorphic granules and as replacement of thin slaty laminae and early burial stylolites. Magnetite (slaty) laminae often have thin stringers of white talc. LC-2 averages 16 feet in thickness, but varies across the extent of the Thunderbird North Deposit, being thinner in the southwest extent of the deposit and thicker in the northeast extent. In the northeast portion of the deposit, the unit is of sufficient thickness and grade to warrant mining despite dilution by the overlying LC-3 waste unit.

LC-3 (Regular-Bedded Unit)

Rocks of the submember are characterized by interbedded greenish-gray thin-bedded cherty and green medium-laminated slaty iron-formation. The unit is weakly magnetic, with the cherty beds conspicuously low in magnetite. The unit averages 13 feet thick, but varies across the deposit. In the southwestern extent of the deposit, the unit is up to 30 feet thick and predominantly composed of slaty iron formation. In the northeastern extent of the deposit, the unit is consistently 10 feet thick and composed predominantly of thin bedded non-magnetic granular chert.

LC-4 (Wavy-Bedded Unit)

The submember is composed of gray medium-bedded cherty oxide-carbonate(ankerite)-silicate(minnesotaite-talc) iron-formation with minor thin irregular thin beds of slaty (magnetite) iron - formation. Magnetite occurs as disseminated idiomorphic granules, patchy haloes cored by coarse slaty intraclasts, and replacement of thin slaty laminae. LC-4 varies from 40-50 feet thick at Thunderbird North, thickening to the southwest. Notable features of LC-4 are magnetite haloes or reaction rims around

small intraclasts, and wispy laminae of magnetite, likely a later diagenetic overprint of early burial stylolites. The LC- 4 and its equivalents are widespread across the Mesabi District, and perhaps the most economically significant subunit, having a high weight recovery (36%) and being capable of producing a low silica concentrate (~2.0%).

LC-5 (Wavy-Bedded Unit)

The submember is composed of pink-gray medium- to thick-bedded cherty oxide-chert-carbonate(ankerite/kutnahorite) iron-formation. Magnetite occurs as disseminated grains and in mottles. The unit is notable for its high carbonate content, containing up to 3.0% CaO in ankerite. LC-5 varies from 40-50 feet thick at Thunderbird North, thickening to the southwest. LC-5 contains a small but variable amount of ‘primary’ (e.g. pre-supergene oxidation) hematite. LC-5 has appreciably more matrix chert than the underlying LC-4, and produces a significantly higher silica concentrate (~6.0%).

LC-6 (Variably-Bedded and/or Mottled Unit)

The submember averages seven feet thick, and consists of a pink massive- to thick-bedded cherty oxide-chert-carbonate(kutnahorite) iron-formation with conspicuous pink carbonate mottles. The unit is composed principally of coarse grained intraclasts, reflecting a relatively high energy depositional environment. The unit also contains an appreciable content of “primary” hematite, and has relatively low magnetite recovery. The unit is remarkably tough, and poses a challenge to mining in that it resists fragmentation during blasting and tends to produce large chunks.

LC-7 (Bold Striped Unit)

The submember is composed of interbedded thick irregular magnetite-carbonate-silicate slaty and green thin- to medium-bedded cherty carbonate(siderite)-silicate(greenalite) iron-formation. The unit averages 13 feet thick. The unit is remarkable in that magnetite occurs predominantly in the thick slaty laminae, resulting in a boldly-striped appearance in drill core. Green LC-7 sharply overlies the pink LC-6, and the contact is a highly visible stratigraphic marker throughout the Virginia Horn area. The transition upward from thick-bedded coarse-grained to thin-bedded fine-grained iron-formation, as well as the contrasting mineralogical assemblages at the LC-6/LC-7 contact, suggests an abrupt transition in the depositional environment. Very fine-grained magnetite (25-45 µm) and intimate association with very fine-grained chert and siderite contribute to grinding and processing difficulties with the LC-7.

LC-8 (Mesabi Select Unit)

The submember is visually similar to LC-7, consisting of a interbedded green medium- to thick-laminar massive slaty and greenish-gray thin-bedded granular cherty carbonate(siderite)-silicate(greenalite) iron-formation. However, the unit contains little or no magnetite and is a waste product that makes excellent aggregate material. The LC-8 averages 21 feet in thickness. LC-8 from the Thunderbird North Mine is the type material for “Mesabi Select” crushed aggregate currently being marketed regionally for road construction, noted for its high specific gravity and angular fragmentation.

Lower Slaty

The Lower Slaty member, as defined at Thunderbird North, averages 52 feet thick and is characterized by non-magnetic and thin-bedded waste rock between the Lower Cherty and Upper Cherty member ore horizons. Other interpretations (Severson and others, 2009) of the Lower Slaty in the Virginia Horn area extend it to up to the top of the UC-4 submember, and up to 309 feet thick.

LS-1 (Intermediate Slate)

The submember is composed of predominantly black massive- to thinly-laminated slaty carbonate(siderite)- silicate (stilpnomelane-minnesotaite)-sulfide iron formation. The LS-1 averages 17 feet in total thickness, and is divisible into a lower half composed of a composed of thick bedded massive intraformational debris flow breccias and an upper half composed of thinly-laminated planar- bedded slaty iron-formation. Locally, thin to medium bedded black flinty cherts are present in the lower portion; these flinty cherts typically occur in pod-like bodies extending a few 100s of feet along strike. The upper portion of LS-1 has undergone extensive bedding-parallel deformation; essentially the entire horizon served as a low-angle fault plane. Small-scale folds are common, as are bedding-parallel syntectonic

quartz-carbonate(ankerite-siderite) veins. The thinly-laminated planar-bedded slaty iron-formation in the upper portion is the so-called Intermediate Slate, a district-scale marker horizon. LS-1 is notable in that it contains a very high percentage of Al_2O_3 (1.86%) and other elements indicative of clastic input, suggesting the basin was experiencing either an influx of clastic detritus, or a sharp reduction in the rate of iron-formation deposition.

LS-2 (Lowermost Thin-Bedded Unit)

The submember is composed of a green to greenish-gray well-cemented very thinly-laminated slaty carbonate-silicate(minnesotaite) iron-formation. The unit averages 35 feet thick. The top of the LS-2 is defined by the appearance of significant magnetic slaty iron-formation. Commonly, this corresponds to the first appearance of thin-bedded intraclast breccias. These breccias commonly have a magnetite-rich matrix.

Upper Cherty

The Upper Cherty member at Thunderbird North contains all potential ore horizons situated above the Lower Slaty waste horizons. This comprises 347 feet in thickness and the remainder of the thickness of the iron-formation exposed in the present workings. The lowermost 257 feet of the Upper Cherty, as defined at Thunderbird North, consists of alternating horizons of dominantly slaty- and cherty-iron-formation; these horizons have been included in the Lower Slaty by other workers (Severson and others, 2009) including US Steel. The Upper Cherty has been subdivided into eleven subunits at Thunderbird North.

LUC-1 (Ore Zone of Lowermost Thin-Bedded Unit)

The submember is composed of gray laminar thin-bedded slaty chert-silicate(stilpnomelane) magnetite iron-formation. The unit averages 18 feet in thickness and is notable for producing a very high silica concentrate (up to ~10% SiO_2). This unit, in common with the other slaty iron-formation horizons in the Upper Cherty member, has a relatively high Al_2O_3 content (~0.56%).

LUC-2 (Lower IBC Unit)

The submember is a heterogeneous unit, composed variously of green-gray thin-bedded slaty iron-formation, interbedded green-gray thin-bedded slaty iron-formation, thin-bedded cherty iron-formation, and gray thick-bedded cherty iron-formation. The unit as a whole averages 46 feet thick. For the unit as a whole, thin-bedded granular cherty horizons predominate over thin- to medium-laminated shales. The abundance and frequency of cherty horizons generally increases up-section within the unit. Locally, pink to green-grey massive- to thick-bedded, coarse-grained, magnetite-bearing granular cherts, upwards of 20 feet thick, are present within the unit. These beds are characterized by significantly higher weight recovery, and significantly lower concentrate silica grades than the unit as a whole.

LUC-3 (Middle Thin-Bedded Unit)

The submember is composed of dark reddish-brown thin/planar-bedded slaty chert-silicate iron-formation. The unit averages 19 feet thick. Nodules and beds of chert are increasingly abundant up-section, culminating in the presence of a 1-2 foot thick horizon containing thin bedded flinty chert, an important marker horizon in the mine.

UC-1 (Middle IBC Unit)

The submember is composed of pinkish-gray thick-bedded cherty oxide-chert-silicate iron-formation. The unit averages 29 feet thick; however, it is not extensive through the deposit. The overall aspect of UC-1 is of a lenticular body on the order of several km in extent. UC-1 is notable in that it contains an appreciable content of 'primary' hematite; this hematite is intimately intergrown with magnetite, and thus, is recovered in the Fairlane concentrator circuit.

UC-2 (Middle Thin-Bedded Unit)

The submember is a dark reddish-brown thin-bedded slaty chert-silicate iron-formation, averaging 46 feet in thickness. UC-2 is generally characterized by low weight recovery and a high concentrate silica, and thus, is marginal ore at best.

UC-3 and UC-3A (Upper IBC Unit)

The submembers are gray thick-bedded cherty iron-formation. Combined, they average 91 feet in thickness. Similar to UC-1, these units are not laterally extensive, and have the overall aspect of lenticular bodies on the order of several km in extent. The two units comprise a single depositional package; however, the lower half (UC-3) is generally characterized by low weight recovery, while the upper half (UC-3A) is characterized by higher carbonate and magnetite content. UC-3A was historically one of the primary ore units at the Thunderbird North Mine; however, it is not now being mined and is only poorly exposed in the pit. UC-3A is notable for an abundance of coarse-grained jasper intraclasts; the vivid colors of these intraclasts have resulted in the name 'confetti ore' being attached to UC-3A.

UC-4 (Uppermost Thin-Bedded Unit)

The submember is a dark reddish-brown thin-bedded slaty silicate iron-formation, averaging 18 feet in thickness. In areas where UC-1, UC-3 and UC-3A are absent, UC-4 is the upward continuation of UC-2. The unit typically has a very low weight recovery. The uppermost 1-5 feet of the subunit is commonly a black, thin-bedded non-magnetic slaty silicate iron-formation. This has been recognized as an important marker horizon, and for some workers (Fig. 9), marks the top of the Lower Slaty member.

UC-5 (Alternating-Bedded Unit)

The submember consists of interbedded reddish-brown thin-bedded slaty silicate iron-formation and thin-bedded cherty iron-formation. The unit averages 15 feet thick. The thin cherty beds commonly contain abundant coarse-grained jasper intraclasts.

UC-6 (Algal/Conglomerate Unit)

The submember is very distinct in that it is composed of red medium- to thick-bedded coarse-grained intraclast conglomerates. Clasts in the conglomerate are composed predominantly of resedimented cherty algal stromatolites (oncolites). The conglomeratic matrix is commonly composed predominantly of manganiferous carbonates, including rhodochrosite (Zeilinski and others, 1994). UC-6 is notable in having the highest manganese content in the Biwabik Iron Formation, averaging ~6.0% Mn (7.8% MnO).

UC-7 (Regular/Medium-Bedded Unit)

The submember is composed of gray to red thick-bedded oolitic cherty oxide-chert-carbonate iron-formation. The unit averages 37 feet thick in the Thunderbird North Mine area, and is known mostly from oxidized drill hole intercepts. The unit appears to consist of a lower red hematitic oolitic cherty iron-formation and an upper gray magnetite-bearing oolitic chert- carbonate (ankerite) cherty iron-formation. The upper gray horizon contains abundant coarse poikiloblasts of ankerite; commonly these are weathered away, leaving vesicle-like vugs in the oolitic cherts.

UC-8 (Thin-bedded unit)

Similar to UC-5, the UC-8 consists of interbedded green-red thin-bedded slaty silicate iron-formation and thin-bedded cherty iron-formation. The unit averages 28 feet thick. UC-8 is known only from (commonly) oxidized drill hole intercepts. The thin cherty beds commonly contain abundant coarse-grained jasper intraclasts. The contact between UC-8 and the overlying US-1 is poorly defined.

Upper Slaty

The Upper Slaty member in the vicinity of the Thunderbird North Mine is only known from oxidized intercepts in a few drill holes, and is not exposed in outcrop.

US-1

The submember is comprised predominantly of reddish-brown thin-bedded slaty iron-formation, and is about 50 feet thick.

US-2

The submember is not exposed in outcrop or mine workings in the Thunderbird Mine, nor has it been intercepted in drilling. Regional drilling data indicates the unit is composed of grey thin-bedded micritic calcareous carbonate, and is about 19 feet thick.

Structure of the Biwabik Iron Formation

Recent studies of bedrock structure along the Mesabi Iron Range (Jirsa and others, 1998; 2002; 2005a; 2005b; Jirsa, 2006) reveal that a protracted history of deformation affected the Biwabik Iron Formation. Much of the formation forms a south-dipping homocline that contains little evidence of tectonic disruption, with the exception of locally well-developed deformation structures. A general sequence of deformation events can be inferred from those localized structures. The precise ages of events on the Mesabi range are unknown; however, a relative chronology for various structural elements can be established from cross-cutting relationships. Assigning deformation events to specific structures is speculative; nevertheless, the "D₀, D₁, D₂..." nomenclature is applied here to refer to suites of apparently related structures. The oldest are those presumably related to soft-sediment deformation (D₀), including slumps, sedimentary breccias, and structures that appear to be the result of differential compaction and localized faulting synchronous with deposition. The earliest "regional" deformation (D₁) is manifest in localized, small-scale rotational structures, bedding-parallel slickensides, and larger nappe and sheath folds. The structures commonly lie along boundaries between units having strong rheologic contrast, such as the contacts between packages dominated by mudstone vs. those composed of siliceous, intraclastic grainstone. Nearly all of these structures display a sense of asymmetry that indicates south-over-north tectonism. This northward vergence, and the apparent timing relative to later structures, is consistent with compressional deformation—potentially related to the Penokean orogen. One of the long-standing controversies in iron-ore genesis is the question of whether oxidation and leaching of iron-formation to form the high-grade hematite ores occurred by supergene or hypogene processes. Although not conclusive, the observation of several early-formed, south-dipping thrust faults with folded, mineralized wall rocks, and bedding-parallel slickensides that host abundant secondary iron and silica implies that at least some of the mineralization was coincident with compressional deformation, perhaps during Penokean orogenesis. This is consistent with the hypogene model proposed by Morey (1999) that attributes oxidation and leaching to groundwater flow driven northward from uplift in the Penokean fold and thrust belt. A second regional suite of structures (D₂) is largely extensional. These structures are monoclines and normal faults that are mutually transgressive; that is, faults that have sympathetically folded wall rocks, and folds that pass gradationally into faults along the trend of axial planes. These are some of the major structures along which oxidation and leaching has occurred, and the focus of most direct-shipping (hematite) ore mining. Veins, vugs, and other secondary mineralization features are abundant. D₂ structures likely formed as localized responses to regional tilting. The most recent deformation effects (D₃) are trough-like collapse structures, presumably related to post-leaching subsidence. The collapse, and associated oxidation and weathering, are best developed in the uppermost subcrop of iron-formation, implying supergene alteration played a significant role in their development. Thus, the answer to the supergene vs. hypogene debate appears to be that both processes were significant, perhaps at different times. Lacking finite ages, the structures can only be inferred to record components of Penokean (Geon 18), Yavapai (Geon 17), Mazatzal (Geon 16), and/or Keweenawan (Geon 11) deformation events.

The overall shallowly-dipping, northeast-striking homoclinal trend of Paleoproterozoic strata along the Mesabi Iron Range is interrupted near the city of Virginia, where strata are warped around an apparent anticline-syncline pair to form the structure known locally as the Virginia horn. In this area, dips as steep as 25° occur, and strikes trend N, NE, and NW. The origin of this structure has been variously ascribed to faulting and folding associated with uplift of Archean bedrock that now cores the anticlinal portion of the Z-shaped horn structure. Intuitively, the granitoid basement rocks were too competent to accommodate ductile compression, and it is therefore unlikely that the horn formed by simple flexural folding. The Alpena fault and several others are marked by differences in the thickness of internal units across them, indicating some deformation was synchronous with deposition (i.e., growth faults). Several faults are essentially continuous along strike with those in the Archean basement. Where displacement sense or magnitude differs significantly between these faults in the two ages of bedrock, the portions affecting Paleoproterozoic strata are inferred to have been reactivated along faults of Archean parentage. The conceptual model shown below (Fig. 10) depicts an interpretation of the structural development in the

horn that invokes some combination of faulting and folding, and addresses the reactivation of what were likely Archean faults reactivated during the Paleoproterozoic.

The development of direct shipping (hematite-goethite) ores appears to have been localized to varying degrees by fault and fold structures within iron-formation. Regionally, natural ore bodies extend from the bedrock surface to depths as great as 120 m. These ores formed by oxidation, hydration, and subsequent

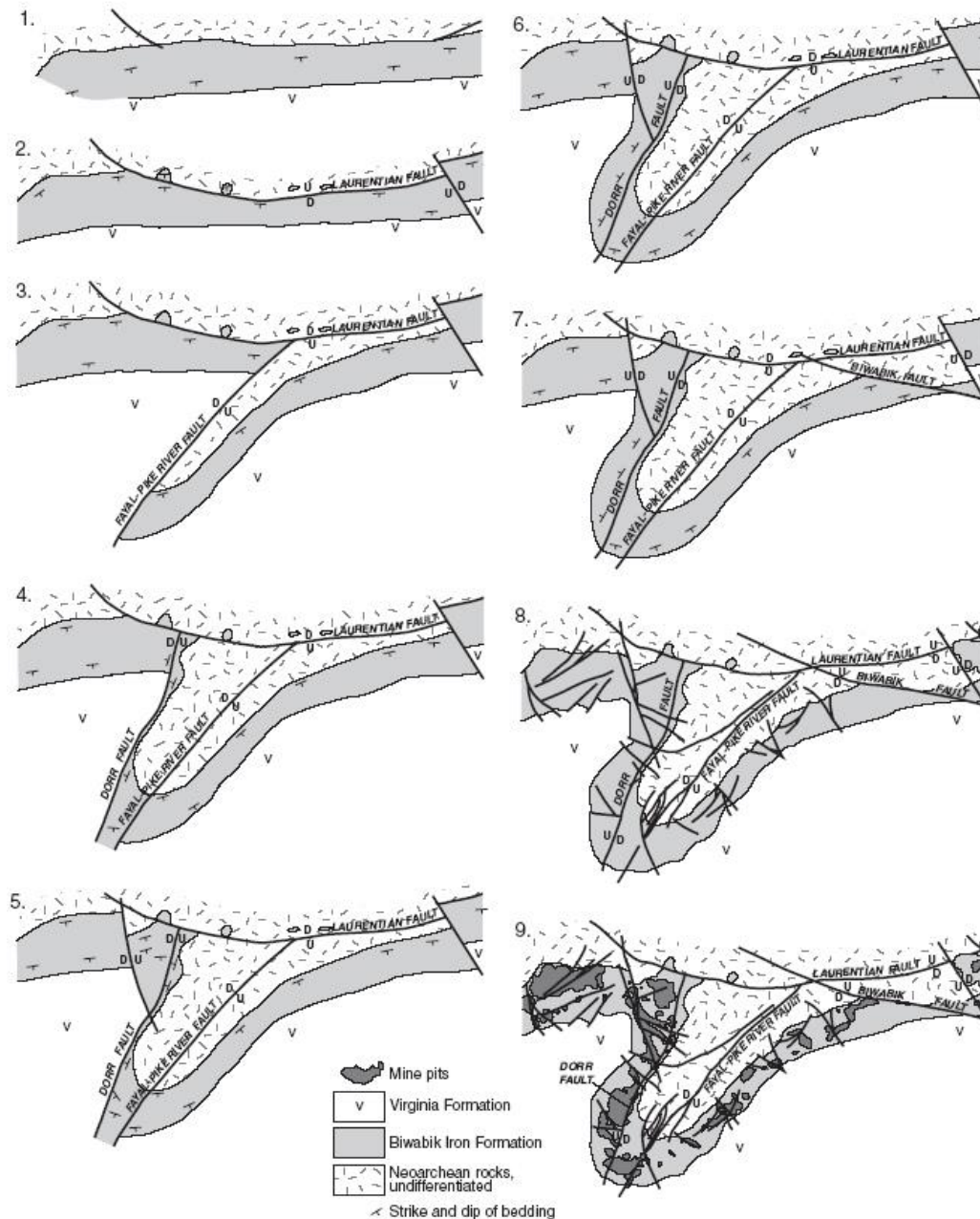


Figure 10. Schematic model showing structural evolution of the Virginia horn structure (from Morey, 2003).

leaching by through-flowing solutions after lithification. Bedding plane fractures, folds, and faults presumably acted as hydraulic conduits and traps for descending and/or ascending solutions that selectively altered certain lithologic units. The direction of fluid movement and the possibility of multiple episodes of alteration are currently unclear, but work to more fully understand these questions is underway (e.g., Losh and Rague, 2013; see *Diagenesis, Alteration, and Fluid Flow discussion below*).

Zones of oxidation along structures are apparent in derivative aeromagnetic imagery (Fig. 11). Linear zones of less magnetic, presumably oxidized iron-formation typically cross the strike of iron-formation at varied angles. Where it can be verified on the ground, many of these zones are coincident with mapped faults, axial planes of minor folds, and major joint networks. Some are also coincident with mined natural ore bodies; though to be clear, most magnetic lows depicted do not represent mineable deposits of natural ore. However, they do represent local zones having variably decreased overall magnetite content.

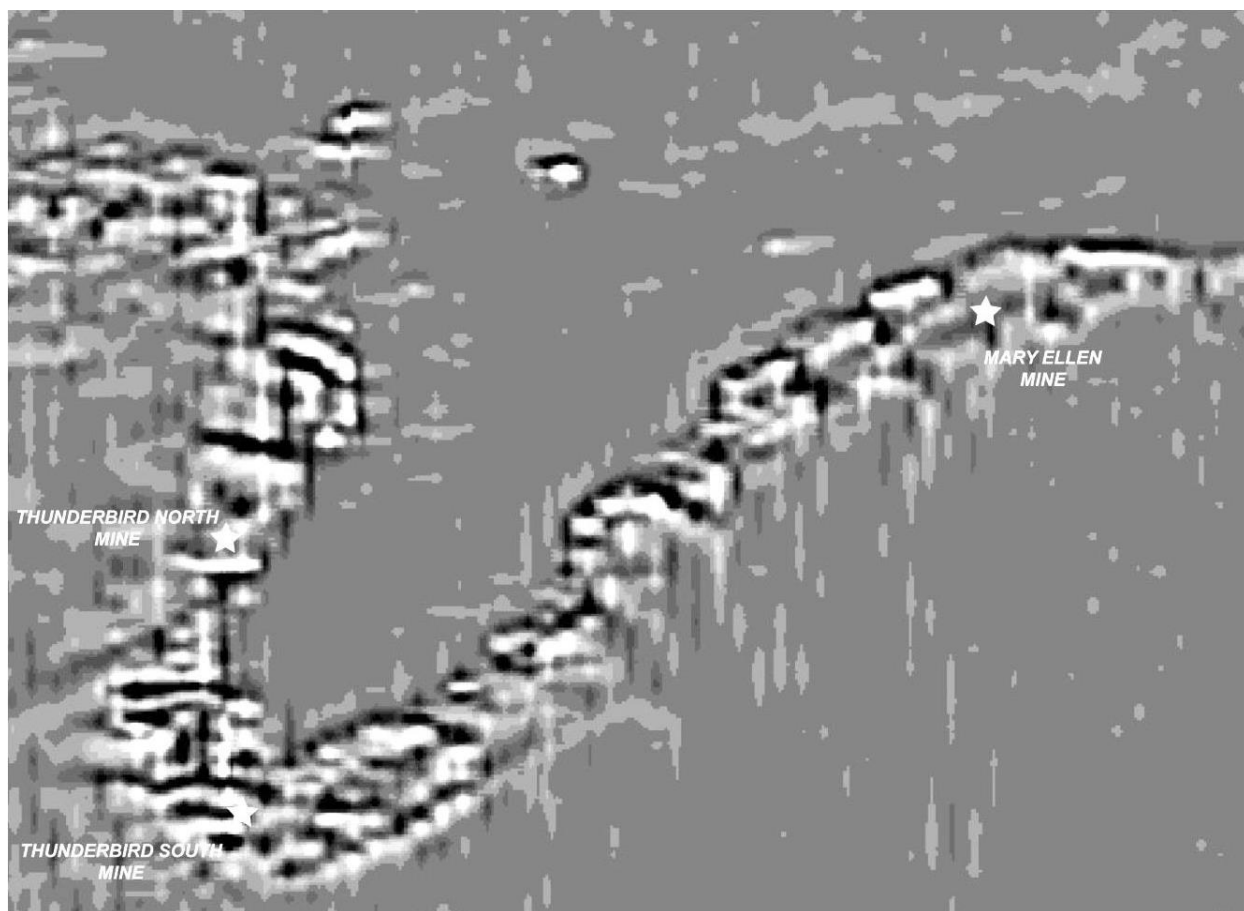


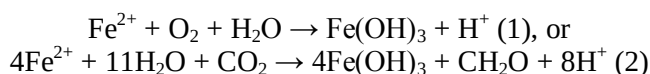
Figure 11. Derivative aeromagnetic map of the central Mesabi Range. Image was created from total field magnetic data, which was regridded from flight lines and band-pass filtered to remove broad wave-length (low frequency) anomalies and reveal contrasts in the short wave-length (near-bedrock surface) anomalies. Magnetic highs are light; lows are dark. In the north-trending limb of the Virginia horn structure, most of the linear magnetic lows that cross the strike of the overall high correspond with folds and faults that have been mapped within iron-formation (White, 1954; and field work by the authors). North-south striping is an artifact of gridding flight line data.

Origin of Iron-Formation

Iron-formation formed by chemical precipitation of dissolved ferrous (Fe^{2+}) iron as a solid phase, most likely a ferric (Fe^{3+}) bearing species. A reduced or low-oxygen atmosphere relative to modern conditions was necessary to allow accumulation of high concentrations of dissolved ferrous iron in seawater. Mineralogical and geochemical evidence indicates co-precipitation of variable amounts of Mg, Ca, Mn, P, Si, and CO_2 in addition to Fe. Silica precipitation may have occurred by adsorption onto ferric iron species settling through the water column (Fischer and Knoll, 2009), by diagenetic reaction with ferric iron precipitates at the sediment-water interface, or direct precipitation on the seafloor (Maliva and others, 2005). It is likely that all of these mechanisms may have played a role.

Geochemistry of Iron Deposition

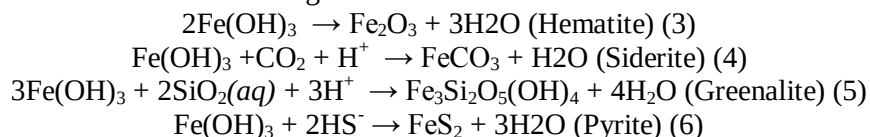
A number of mechanisms have been proposed to explain iron precipitation and deposition, including direct oxidation as a byproduct of oxygenic photosynthesis, anoxygenic photosynthesis utilizing iron as an electron acceptor, and abiotic photochemical oxidation. Regardless of mechanism, the reactions can be generalized as:



In each of these models, iron oxidation is placed close to the surface within the photic zone. The photosynthetic models intimately associate iron precipitation with biological activity, and presume that iron precipitates are raining out of the water column along with organic material. Each of these models also presumes a reservoir of dissolved iron in anoxic waters lying beneath a chemocline. In the case of shallow water iron precipitation, this implies a current- or tidal-driven flux of anoxic bottom waters into shallower water environs.

Fixation of Ferric Iron into Ferrous Iron species

Ferric iron hydroxide ($\text{Fe}(\text{OH})_3$) precipitates formed at or near the top of the water column and settled to the bottom. Hydrous ferric iron oxides have not been recognized as primary minerals in iron-formation. All iron-bearing minerals in iron-formation may have been produced by diagenetic reactions at or near the sediment-water interface. Basic iron-fixing reactions include:



With the exception of hematite, all these iron-bearing minerals contain ferrous (Fe^{2+}) iron, indicating diagenetic iron-fixation was accompanied by iron reduction. Iron reduction was driven by a combination of settling of precipitates through the chemocline into anoxic bottom water, or respiration of organic material at the sediment-water interface, or a combination of both processes. Absent carbonate, silica, or sulfur with which to react and form a stable mineral species, the transformation of insoluble ferric iron precipitate to highly soluble ferrous iron would return dissolved iron back into the water column.

Formation of geologically stable iron-formation is not a function of deposition of ferric iron precipitates, rather that of fixation of ferric iron into stable, dominantly ferrous iron mineral species. It is more appropriate to speak in terms of iron-formation accumulation and accumulation rates than iron-formation deposition and deposition rates. Viewed in this context, apparent decreases in “deposition” rate are actually decreases in accumulation rates, and may reflect a lack of suitable fixative at the sediment-water interface rather than a decrease in iron precipitation rates at the top of the water column.

Iron-Formation Mineralogy

The iron-bearing minerals in iron-formation consist of oxides, carbonates, silicates, and sulfides. James (1954) recognized that the iron mineralogy varied systematically, and reflected distinct lithostratigraphic facies, at least in part. His iron-formation facies concept (oxide, silicate, carbonate, and sulfide facies) continues to provide a compelling framework within which to interpret iron-formation sedimentology and mineralogy. The diversity of iron minerals found in iron-formation (Table 2) is a

direct reflection of the diversity of the sedimentological and geochemical environments in which the iron-formation formed.

The likelihood of extensive replacement of primary iron precipitates has resulted in significant controversy regarding the precise nature of the primary precipitate, and the precise reaction pathways responsible for formation of the observed mineral assemblages (Simonson, 2003). Eh and pH are major controls on the stability of the iron minerals in both the depositional and diagenetic environments (Ojakangas and others, 2005). Klein (2005) has suggested that the original precipitate materials were probably hydrous Fe-silicate gels of a greenalite-type composition; Na-, K- and Al-containing gels approximating stilpnomelane compositions; SiO₂ gels; Fe(OH)₂ and Fe(OH)₃ precipitates; and very fine-grained carbonate oozes. A variety of other primary chemical precipitates for iron-formation in general have also been postulated by an assortment of authors and include siderite, iron hydroxides, iron silicates (Konhauser and others, 2002; Rajan and others, 1996), and colloidal iron silicates (Lascelles, 2007).

“Clastic” components such as Al₂O₃, TiO₂, K₂O, and Na₂O were likely deposited as eolian dust, and reflect a far-travel clay mineral component eroded from exposed cratons. Nevertheless, within the iron-formation these elements are typically found in the iron silicate mineral stilpnomelane, suggesting that even the clastic component participated in diagenetic reactions.

Mineral	Formula
Oxides	
Magnetite	Fe ₃ O ₄
Hematite	Fe ₂ O ₃
Goethite	FeO(OH)
Silicates	
Chert	SiO ₂
Chalcedony	SiO ₂
Microcrystalline Quartz	SiO ₂
Stilpnomelane	K(Mg, Fe ⁺² , Fe ⁺³) ₈ (Si,Al) ₁₂ (O,OH) ₂₇
Minnesotaite	Fe ₃ Si ₄ O ₁₀ (OH) ₂
Talc	Mg ₃ Si ₄ O ₁₀ (OH) ₂
Greenalite	Fe ₃ Si ₂ O ₅ (OH) ₄
Chamosite (Al-rich Fe-chlorite)	Fe ₃ (Al,Si) ₂ O ₅ (OH) ₄
Carbonates	
Siderite	FeCO ₃
Ankerite	Ca(Fe,Mg)(CO ₃) ₂
Kutnohorite - Ferroan	(Ca,Mn)(CO ₃) ₂ - Ca(Mn,Mg,Fe)(CO ₃) ₂
Dolomite	CaMg(CO ₃) ₂
Calcite	CaCO ₃
Sulfides	
Pyrite	FeS ₂
Pyrrhotite	Fe _(1-x) S

Table 2. Common mineral names and formulas associated with the Biwabik Iron Formation (excluding the more highly metamorphosed eastern Mesabi Iron Range in proximity to the Duluth Complex).

Oxides

Hematite is the iron-bearing mineral most commonly associated with iron-formation. However, primary hematite is a relatively rare component of the BIF, occurring most prominently in oxide facies iron-formation of the UC member. Magnetite is common throughout the BIF sequence. The relatively coarse-grained idiomorphic magnetite characteristic of gifs are late diagenetic in origin (LaBerge, 1964; LaBerge and others, 1987; Zanko and others, 2003) and form by the replacement of pre-existing iron silicates and iron carbonates (French, 1973). Fine-grained magnetite in 'slaty' bif layers likewise formed by diagenetic reaction of iron silicates and carbonates.

Earlier work on the oxidized taconites of the western Mesabi Iron Range was accomplished by Bleifuss (1964). He showed that late hematite was developed by the oxidation and pseudomorphic replacement of magnetite octahedra, that layers of goethite were precipitated from solutions likely derived from the oxidation of siderite, and that some goethite formed by the oxidation of acicular iron silicate minerals.

Silicates

Greenalite is considered to most closely reflect the composition of an initial ferric hydroxide/silica gel precipitate in that it exhibits no detectable replacement of any pre-existing phase (French, 1973; LaBerge and others, 1987; Simonson, 1987; Klein, 2005). Within gifs, greenalite most often occurs as round-shaped granules that are <1 mm in diameter.

Stilpnomelane is a secondary mineral that commonly replaces early iron silicates (greenalite) (French, 1973). The presence of alumina and potassium suggests reaction with the detrital dust component found in the iron-formation. French (1973) suggests that stilpnomelane formed under conditions ranging from diagenesis to low-grade metamorphism.

Minnesotaite is a common component of throughout the BIF sequence, and the type locality for this mineral was located in the north end of the Thunderbird North mine. Stoichiometrically analogous to magnesian talc, but structurally dissimilar, it generally occurs as sheaves or needles replacing greenalite granules (French, 1973). True talc, including ferroan talc, locally comprises a significant amount of the silicate fraction within gifs. McSwiggen and Morey (2008) show that both chamosite and talc are common throughout portions of the Biwabik Iron Formation.

Carbonates

Ankerite and siderite are common early diagenetic minerals. Siderite commonly occurs within laminated bifs, while ankerite commonly occurs as idiomorphic replacement of primary granules within gifs. Locally, coarse-grained poikiloblastic aggregates of ankerite (mottles) are found in gifs. These mottles are clearly late diagenetic replacements. McSwiggen and Morey (2008) report manganese substitution for iron in the dolomite-ankerite series, leading to kutnohorite and ferroan kutnohorite composition in the Lower Cherty. Both Mg and Mn substitute for Fe in siderite in the Lower Cherty, with some samples containing as much as 20 to 25 mole percent MnCO_3 . The average composition of siderite from the Lower Cherty $[(\text{Fe}_{.60-.78}\text{Mg}_{.15-.19}\text{Ca}_{.02-.03}\text{Mn}_{.04-.22})\text{CO}_3]$ differs considerably from that of the remainder of the iron-formation, where siderite compositions averages $[(\text{Fe}_{.88-.84}\text{Mg}_{.03-.11}\text{Ca}_{.02-.03}\text{Mn}_{.02-.07})\text{CO}_3]$ (McSwiggen and Morey, 2008).

Sulfides

Sulfide minerals are ubiquitous throughout the BIF sequence. Pyrite and pyrrhotite are the most common, with minor amounts of arsenopyrite, cobaltite and galena (Theriault, 2011). Pyrite occurs ubiquitously as in trace amounts as idiomorphic cubic or dodecahedral crystals, framboids, and spheroids. Less commonly, sulfide occurs in larger blebs. The Intermediate Slate (LS-1) contains the greatest abundance of sulfide, apparently associated with elevated mercury and arsenic concentrations (Morey and Lively, 1999).

Alteration and Regional Fluid Flow

Common features within the Biwabik Iron Formation are quartz-carbonate±iron silicate veins that occupy vertical fractures and bedding-parallel slip planes. These veins postdate magnetite formation and diagenesis of the iron-formation, but display textures indicative of syntectonic growth, suggesting they may be related to far-field deformation, perhaps associated with the Penokean orogen. In the vicinity of direct-shipping ore deposits, these veins are commonly overprinted by: 1. complete or partial dissolution of carbonate minerals; 2. brecciation of the quartz, perhaps associated with volume loss collapse of the iron formation; and 3. recementation by secondary iron oxides and silica.

Recent fieldwork in the Hibbing Taconite, Thunderbird North, and Thunderbird South/Fayal Mines, combined with petrographic, SEM, fluid inclusion, and geochemical techniques, have elucidated oxidation by deep, saline, hydrothermal-diagenetic waters at relatively low water/rock ratios (e.g., Losh and Rague 2013). Fluid inclusions in fault breccia and low-angle and high-angle veins containing secondary minerals (quartz, calcite, minnesotaite, stilpnomelane, hematite) have average homogenization temperature of $155^{\circ}\text{C} \pm 17^{\circ}\text{C}$ ($n=278$), and salinity of 9.5 ± 5.3 wt% NaCl equivalent ($n=160$). Temperature correction due to pressure is on the order of 50°C . There is no significant difference in fluid inclusion homogenization temperature or salinity between fault breccias (including quartz cement) and veins of diagenetic affinity. These results agree well with oxygen isotopic temperatures of $150^{\circ}\text{--}200^{\circ}\text{C}$ for diagenesis determined by Perry and others (1973). The oxidizing fluids, a mixture of diagenetic and meteoric fluids, infiltrated along high-angle faults that contain vein quartz cemented by quartz ± iron oxides (typically goethite), and brought about oxidation of magnetite to hematite and Fe-silicates to goethite (the latter reaction also yielding silica as a reaction product), accompanied by quartz recrystallization. Silica liberated from this oxidation filled microfractures, typically only a few microns in width, and pits within altered magnetite grains. This contributes to the general observation of high silica in magnetite concentrates from oxidized ores. As silica was not dissolved but rather only remobilized during this oxidation event, ore was not significantly upgraded; in fact, the introduction of quartz into microfractures in magnetite locally diminished magnetic taconite ore quality. Quartz-filled microfractures in magnetite are also observed in unweathered ‘slaty’ iron formation near bedding-parallel faults, as in the Thunderbird North mine.

Ore Deposits

The Biwabik Iron Formation contains about 30% iron throughout its thickness, irrespective of lithofacies, mineralogy, and grain size. However, only a fraction of the formation hosts recoverable iron mineralization. Historically, two deposit types have proven economically feasible to mine: direct-shipping (natural) ores (DSO) and magnetic taconite.

Direct-shipping ores are composed of hematite and goethite enriched by supergene leaching of silica from the pristine iron-formation. Early in the development of the Mesabi Range, DSO were shipped directly from the mine. In later years, gravity concentration was used to upgrade the iron content of the ores. DSOs formed from leaching of any lithofacies ranging from thin-bedded slaty band iron-formation to thick-bedded, coarse-grained granular iron-formation. Because of this, DSO quality was quite variable in terms of deleterious elements, including phosphorus, alumina, manganese, and structural water (from goethite). This necessitated an elaborate and extensive system of ore grading and blending of railcar size shipments at rail yards and ore docks at the shipping ports to maintain consistent quality blast furnace feed.

Magnetic taconite ores are composed predominantly of coarser grained magnetite found in granular iron-formation. These rocks are capable of producing a high-grade magnetite concentrate after fine grinding by magnetic separation. The magnetic concentrates are agglomerated and fired into the 3/8”-1/2” taconite pellets familiar to would-be slingshot assassins throughout the Great Lakes region. The resulting pellets have a significantly higher iron grade, significantly lower deleterious element content, and superior smelting properties relative to the DSO production they have replaced. Furthermore, the product is easily transportable.

Direct-shipping Ores

The direct-shipping ores of the Mesabi Iron Range fall under the Soft Iron Ore category of Marsden and others (1968). They are generally porous masses of hematite, goethite, and minor magnetite and manganese oxides. Gangue minerals consist of quartz, clay, and minor carbonate. Fundamentally, soft ore formation is the product of preferential leaching of silicate and carbonate components from the iron-formation, and alteration of the primary iron oxide, silicate, and carbonate minerals to secondary hematite and hydrous iron oxides (Marsden and others, 1968). The iron content of the iron-formation is increased by loss of gangue material (primarily silica and carbonate), rather than enrichment or replacement by supergene iron minerals. On the Mesabi Iron Range, soft ore bodies occur in trough, fissure, and irregular ore bodies, reflecting variable degrees of ore formation along faults, folds, or zones of fracturing. They also occur as stratiform ore bodies, reflecting ore formation along favorable horizons. Generally, soft ore bodies extend from the bedrock surface to depths of 400 or 500 feet (Marsden and others, 1968). Ore formation was evidently a multi-stage process. Early desilicification of the iron-formation was accompanied by alteration of primary magnetite to hematite, and alteration of primary iron silicates and iron carbonates to goethite. Much direct-shipping ore exhibits textures indicative of a second stage of enrichment. Secondary porosity induced during desilicification is commonly filled by paragenetically late iron hydroxides and hydrous iron oxides. Dripstone textures indicate that at least some of these secondary iron hydroxides were precipitated in the vadose zone. Leith (1903) noted that hydrous minerals were more abundant in the shallower portions of the deposits, suggesting the presence of a supergene enrichment zone, perhaps coincident with a paleo-water table. Ore formation and desilicification were accompanied by mass loss (as much as 50% by weight) and, to a variable extent, volume loss. Unaltered iron-formation has a specific gravity of 3.3-3.4; Leith (1903) reported typical direct-shipping ore specific gravities in the range of 2.6-3.1, with some ores as low as 2.0-2.1. Mass loss was typically accompanied by structural collapse and formation of a synclinal structure in the ore body (D_3 deformation; see discussion above in *Structure of the Biwabik Iron Formation*). Commonly, the ore retained bedding and geochemical traits inherited from the precursor iron-formation, with the steepest dips adjacent to the margins of the deposits.

The nature and timing of ore formation has been the subject of much debate. The clear association of many deposits with fault and fracture zones, as well as the sharp wall contacts, has been cited as evidence in favor of a hydrothermal origin (Morey, 1999). In contrast, the clear association of the stratiform bodies with the paleosurface argues strongly in favor of a supergene origin. The complex paragenesis of the ores suggests that multiple events may ultimately have been responsible for development of the ores. Oxidation of iron-formation in the Mesabi Range has long been thought to have been solely the result of near-surface interaction with meteoric water, most intensely during saprolite formation during the Cretaceous (Leith, 1903; Sloan, 1964; Marsden and others, 1968), or during other time periods with a tropical climate. Gruner (1956) and Morey (1999) proposed that at least some if not all of the intense oxidation was of hydrothermal origin but did not characterize the effects, nature, or ultimate source of the fluids responsible for this alteration. Hydrothermal oxidation, accompanied by dissolution of non-oxide minerals, has been implicated in the upgrading of iron ores in Australia (Thorne and others, 2008) and Brazil (Rosiere and others, 2008). On the Mesabi Iron Range, oxidized magnetic taconite ore has been locally characterized by high Davis Tube concentrate silica values, particularly adjacent to faults. Hydrothermal oxidation may have taken place during the Paleoproterozoic, when the currently-exposed Biwabik Iron Formation was the most deeply buried and was undergoing diagenesis. Later lateritic weathering, perhaps during the Cretaceous, dissolved silica and all other non-iron oxides, resulting in the natural ores as found in the Fayal Mine. Geochemically, these ores are characterized by pronounced cerium anomalies, which can result from intense oxidation near the surface, consistent with a lateritic interpretation for these ores. The older, fault-related hydrothermal oxidation did not produce cerium anomalies, consistent with its deep-seated setting.

Magnetic Taconite Ores

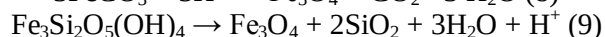
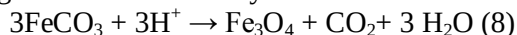
The taconite reserves of the Mesabi Range are comprised of magnetite-rich horizons in the Biwabik Iron Formation. Although the iron content of the Biwabik Iron Formation is relatively uniform, the proportion contained in magnetically recoverable magnetite is highly variable, ranging from less than 10% (typically in slaty banded iron-formation) to greater than 25% (typically in cherty granular iron-formation). Mineable horizons exist throughout the entire iron-formation thickness in the central Mesabi district. Principal ore units include the middle 100' of the Lower Cherty (UTAC LC-4, LC-5), and in variable-thickness Upper Cherty submembers (e.g. LUC-2, UC-3 & 3A).

Ores are classified on their concentrate weight recovery (typically >25%), crude magnetic iron (~17-25%), and concentrate Fe and SiO₂ grades (product averages 66% Fe, 4.5-5% SiO₂). Certain cherty ores (LC-4) can produce concentrate silica grades as low as 2% with standard grinding and separating techniques (75-80% -325 mesh, or P80 45 µm), with most cherty ores averaging ~5-6% concentrate SiO₂. Slaty magnetite taconite ores produce concentrates of higher concentrate SiO₂, reflecting finer magnetite grain size and textural intergrowth with gangue minerals. Minor contaminants (CaO, MgO, MnO from carbonates, K₂O and Al₂O₃ from silicates) are related to very specific stratigraphic horizons, allowing accurate mine-to-mill blending reconciliations. Three-position blending and maximizing the use of a single high-silica ore source are required for stable processing operations.

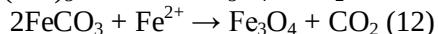
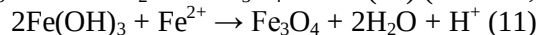
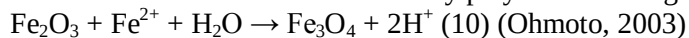
The magnetite in magnetite taconite ores formed as a result of low-temperature diagenetic recrystallization, likely from reaction of oxide and/or carbonate precursors:



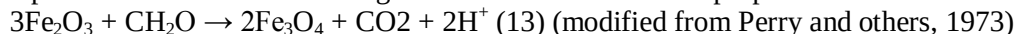
Textural relationships also suggest formation directly from a carbonate or silicate precursor:



Magnetite-rich iron-formation is typically enriched in iron relative to non-magnetite rich iron-formation (Fig. 12), suggesting reactions with dissolved ferrous iron may play a role in magnetite formation:



Reaction of precursor ferric oxide with organic material has also been proposed as a mechanism:



Overall, magnetite formation shows a clear affinity for horizons with sedimentological, mineralogical, and geochemical evidence for a significant carbonate component in the primary precipitate. The association of carbonate with subsequent magnetite formation suggests that a buffered pH was as significant a control as Eh. Textural relationships indicate that magnetite formation occurred after the onset of burial stylolitization and significant chemical compaction. However, it was apparently complete prior to post-Penokean deformation and fluid flow, as it is cross cut by hydrothermal quartz-carbonate veins associated with this event.

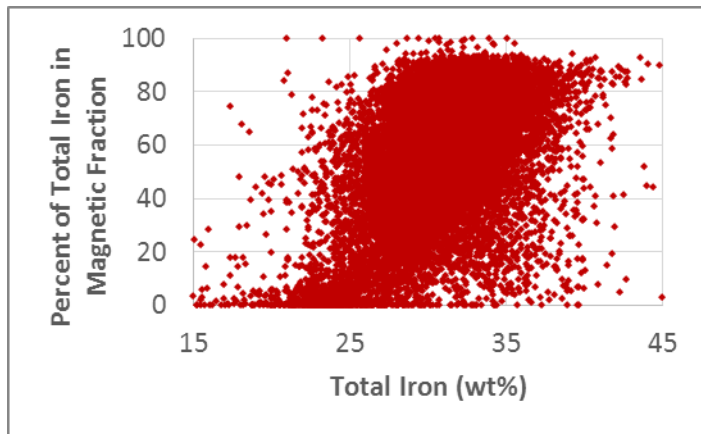


Figure 12. Total iron versus percent of iron in magnetic fraction of unweathered, unmetamorphosed Biwabik Iron Formation. Note that the highest total iron contents (>35%) are associated with the highest magnetic iron fraction; this suggests magnetite formation was accompanied by iron enrichment.

Production

Annual production of direct-shipped ore and taconite pellets produced on the Mesabi Iron Range are shown in Figure 13. Production of direct-shipping ore started in 1892 and rose steadily until 1942, when a record 54 million tons were produced. Gravity concentrate production rose steadily thereafter, until a record 77 million tons of direct-shipping and gravity concentrate ore was produced in 1953. Reserve Mining Company initiated the first large scale taconite operation in 1955, and by 1967 taconite production from six taconite facilities accounted for more than half of iron ore production. The Mesabi Range iron ore industry weathered the global resource recession of the 1980s largely intact, accounting for over 75% of US iron ore production by the end of the decade. The industry continues to evolve, with six taconite facilities (40 mtpy capacity), three tailings recovery facilities (3 mtpy capacity), and a value-added direct reduced iron facility (0.5 mtpy capacity) in production.

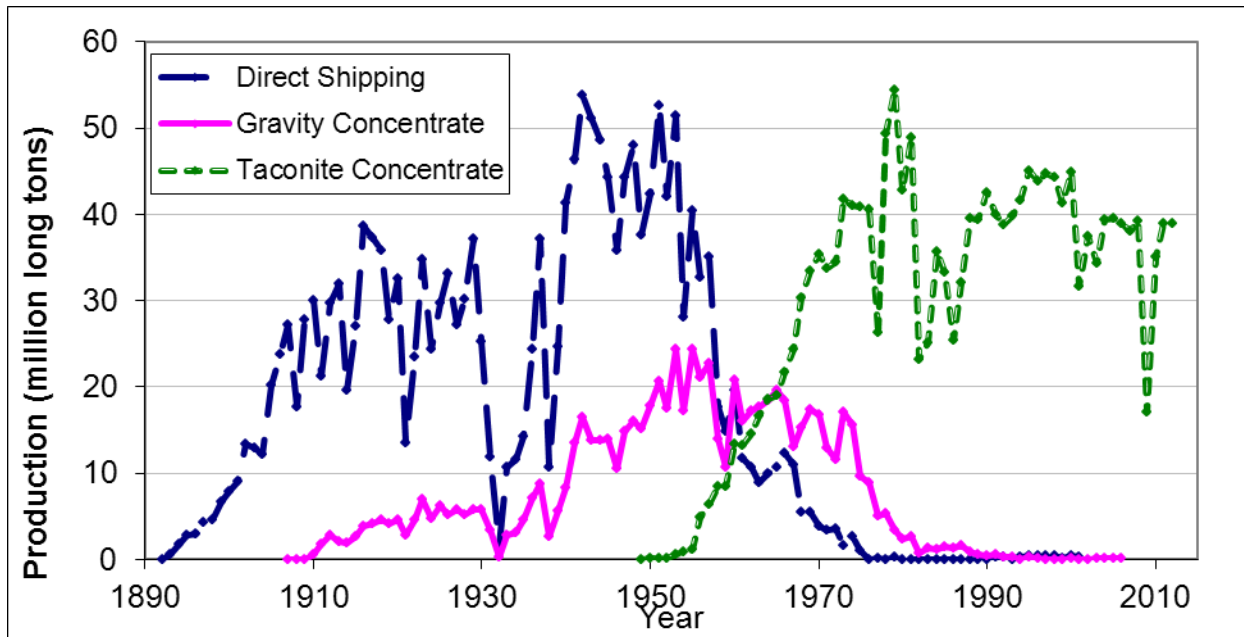


Figure 13. Annual production of direct-shipping ore, gravity concentrates, and taconite concentrates from the Mesabi Iron Range for the period 1892-2012.

DESCRIPTION OF FIELD STOPS

STOP 1—Stratigraphic section of the Biwabik Iron Formation, Thunderbird North Mine 535000E/5257910N (UTM Zone 15 coordinates, NAD83 datum)

Eveleth 7.5' USGS Quadrangle;

SWSW, Section 29, T58N, R17W

**NOTE: THIS SITE IS LOCATED ON AN ACTIVE MINE SITE. DO NOT ATTEMPT TO ENTER WITHOUT FIRST OBTAINING PERMISSION.*

Directions:

Beginning in Hibbing, proceed east on US Highway 169 to the interchange with Highway 53 in Virginia (22 miles). Turn south (right) to merge onto US Highway 53, and drive 4.1 miles to the stoplight intersection with Grant Avenue in Eveleth. Turn west (right) onto Grant Avenue, and drive south 0.5 miles to the Cliffs Natural Resources Thunderbird Mine entrance.

Historical Overview:

Material mined at the Thunderbird North Mine consists of taconite ore horizons from the Lower Cherty, Lower Slaty, and Upper Cherty members. Direct-shipping ore, also referred to as a natural ore, was originally mined in the immediate vicinity from the Auburn (1894-2002), Virginia (1910-1953), and Gross-Nelson (1944-1977) deposits. Exploration for magnetic taconite at this site began in earnest in 1960, after the opening of pioneering taconite operations at the Reserve Mining Company (now Northshore Mining) and Erie Mining Company (now Cliffs Erie site) in the mid-1950s. Drilling by Oglebay Norton Company identified a substantial magnetic taconite deposit in the area and the property was jointly developed with the Ford Motor Company – groundbreaking occurred in June, 1964.

The Thunderbird North mine and Fairlane plant began producing in November, 1965, with an initial rated capacity of 1.6 million tons of iron ore pellets per year. In 1977, addition of a second concentrating and pelletizing line, and the opening of the adjacent Thunderbird South mine, increased rated capacity to 6.0 million tons of pellets. The Thunderbird South mine closed in 1992, and in 1996, ownership of the operation was transferred to Eveleth Mines LLC. Eveleth Mines closed the concentrating and pelletizing Line 1 in May, 1999, reducing capacity to 4.2 million tons of pellets. The remaining operation was idled in May, 2003. The idled facility was purchased and re-opened by United Taconite LLC in December, 2003 (now owned 100% by Cliffs Natural Resources). They subsequently refurbished and reactivated Line 1 in December, 2004, which increased the annual rated capacity to 5.2 million tons of pellets.

Description:

Depending on access, one or more sites with mine exposures of the Lower Cherty, Lower Slaty, and Upper Cherty members will be visited. Refer to the detailed stratigraphy section for more information regarding the submembers visited.

STOP 2—Fault and Associated Quartz Veining and Alteration/Oxidation, Thunderbird South Mine

534100E/5255520N (UTM Zone 15 coordinates, NAD83 datum)

Eveleth 7.5' USGS Quadrangle

SWSW, Section 5, T57N, R17W

**NOTE: THIS SITE IS LOCATED ON AN ACTIVE MINE SITE. DO NOT ATTEMPT TO ENTER WITHOUT FIRST OBTAINING PERMISSION.*

Directions:

Proceed southward on company roads through the Thunderbird North Mine. Cross County Highway 101 (through two remote operated gates on either side of the highway) and continue southeast to the north side of Thunderbird South.

Description:

The site contains multiple exposures of Lower Slaty units on several benches, with the conspicuous fault/quartz vein area trending SW into the flooded pit. The complete Lower Cherty section remains as reserve in Thunderbird South

The exposure in the Thunderbird South pit contains a N45E-trending high-angle fault in the LS2/LUC1 units. The fault is approximately 30 cm wide, and contains quartz veins that have been brecciated and cemented by very fine-grained quartz intergrown with goethite. Fluid inclusions in the quartz cement, which is inferred to have precipitated during the hydrothermal oxidation event (hence its intergrowth with goethite), have average homogenization temperatures of 155° C (n=22) and salinity of 7.3 wt% NaCl equivalent, clearly indicating the involvement of saline hydrothermal fluids in goethite-forming oxidation, and furthermore implicating more widely-distributed diagenetic fluid in that alteration. Breccia clasts of quartz vein from this fault zone have essentially the same homogenization temperatures and salinities as the quartz cement. In terms of trace element geochemistry, the fault breccia is remarkably similar to iron-formation, implying it was largely buffered by iron -formation in a rock-dominated system. Notably, the fault breccia displays a distinctive positive Europium anomaly, as does the iron-formation (see also Planavsky and others, 2009). Similar oxidized high-angle faults are known throughout the Iron Range. Adjacent to these faults, iron-formation is oxidized, with iron silicates altered to goethite + quartz, chert textures are overprinted by recrystallized quartz, and magnetite is oxidized to martite. The hydrothermal oxidation is commonly overprinted by late red hematite (+/- goethite) that formed near Earth's surface: it coats fractures and is associated with silica dissolution.

STOP 3—Security Reserve/Fayal Complex Direct-shipping Ore

535070E/5255050N (UTM Zone 15 coordinates, NAD83 datum)

Eveleth 7.5' USGS Quadrangle

NWSE, Section 6, T57N, R17W

**NOTE: THIS SITE IS LOCATED ON AN ACTIVE MINE SITE. DO NOT ATTEMPT TO ENTER WITHOUT FIRST OBTAINING PERMISSION.*

Directions:

Drive around the western and southern sides of the Thunderbird South pit, past the crusher, and east to ramp into the Fayal Pit. The site is an approximately 200 meters (650 feet) long exposure of iron-formation and direct-shipping ore exposed along a northeast-trending access ramp into the flooded Fayal Mine complex.

Historical Background:

The Fayal Mine (1895-1965; total production 44.5 million tons) was discovered in November, 1893 by David T Adams of Duluth. The mine site was initially explored by the McInnis Mining Company and was sold to the Minnesota Iron Company (a component of the 1901 United States Steel merger), after which the mine was operated by the Oliver Iron Mining Company.

Production of direct-shipping ore began in 1895 and was initially extracted by shaft from underground operations. Open pit operations facilitated a rapid increase in production, reaching 1.9 million tons in 1902. Through the end of 1919, the complex had yielded an aggregate of 29.9 million tons – more than a million tons per year since 1895 (and two-thirds the ultimate production). The Fayal complex was closed in 1933, but was reopened on a smaller scale, as an open-pit truck operation, to recover lower grades of ore between 1944 and 1965. Final development plans included recovering approximately 794,000 tons of Lower Slaty- and lower Upper Cherty-hosted ores along the south side of the deposit (Security Reserve). However, by the time final mining was contemplated by Auburn Minerals LLC (ca. 2000), the sulfur content of the reserve was deemed unacceptably high.

Description:

Included in the Security Reserve is an access ramp to the flooded Fayal Mine. Along this ramp, direct-shipping ore is exposed in both the floor and wall of the ramp. This site is one of the few remaining locations on the Mesabi Iron Range to view in situ direct-shipping ore.

All direct-shipping ore in the Fayal deposit falls under the Soft Iron Ore Classification of Marsden (1968). The Fayal ore consists predominantly of hematite and goethite, with minor magnetite and manganese oxides, as is common with the other soft ore deposits of the Mesabi Iron Range. Silica and clay minerals are the predominant gangue minerals. In 1901, Fayal direct-shipping ore was reported to averaged 63.8% iron, 0.037% phosphorus, and 2.95% silica (dry basis; Leith, 1903). The direct-shipping ore visible in the Fayal ramp occurs along the margin of the deposit and likely has lower iron and higher silica content than the typical higher grade ore shipped from the Fayal deposit for most of its life.

Iron-formation exposed along the east side of the Fayal ramp parallels the contact of the direct-shipping ore deposit. The ore is formed from predominantly slaty proto-ore, and displays varying degrees of desilicification (leaching) and oxidation. Bedding in direct-shipping ore on the north end of the ramp clearly displays a relatively steep dip to the west and lies near the center of the trough.

The west side of the ramp parallels the northeast-trending Fayal fault, a high-angle, west-dipping normal fault, and one of the larger structures cross-cutting the Biwabik Iron Formation. The fault is occupied by a thick, brecciated, and re-cemented quartz $\pm$ carbonate vein. Visible immediately adjacent to the large vein is drag folding in the footwall iron-formation, indicating a hangingwall- (westside-) down sense of motion.

STOP 4—Drill Core Display

535000E, 5257910N (UTM Zone 15 coordinates, NAD83 datum)

Eveleth 7.5' USGS Quadrangle

SWSW, Section 29, T58N, R17W

**NOTE: THIS SITE IS LOCATED ON AN ACTIVE MINE SITE. DO NOT ATTEMPT TO ENTER WITHOUT FIRST OBTAINING PERMISSION.*

Directions:

Proceed back north along the same route to the core shack within the Thunderbird North Mine.

Description:

A drill hole cored through most of the Biwabik Iron Formation, and a portion of the upper Pokegama Formation, will be on display inside the core shack or, weather permitting, will be laid outside. Iron-formation submembers will be labeled according to the Thunderbird North classification scheme.

STOP 5—Algal/Conglomerate unit of the Upper Cherty member, Mary Ellen Mine

548260E, 5264380N (UTM Zone 15 coordinates, NAD83 datum)

Biwabik 7.5' USGS Quadrangle

NENW, Section 10, T58N, R16W

**NOTE: THIS SITE IS LOCATED ON AN ACTIVE MINE SITE. DO NOT ATTEMPT TO ENTER WITHOUT FIRST OBTAINING PERMISSION.*

Directions:

From the Thunderbird Mine entrance, turn left (north) on Grant Avenue. Proceed to the intersection with Highway 53 (0.5 miles), and turn left (north). Proceed 1.5 miles to the intersection with Highway 135, and turn right (east) to merge onto MN Highway 135. Drive east 10 miles to the intersection with County Road 715, located just outside the western outskirts of Biwabik. Turn left (north) on 715, and proceed 0.2 miles. The entrance to the Mary Ellen Mine will be on the south (left) side of the road.

Historical Background:

The Mary Ellen Mine was first opened in 1924 by Pioneer Mining (Stanley Mining, operator), and saw regular production of what was termed 'hard, bluish-red siliceous hematite' through 1928. Stanley Mining reopened the Mary Ellen in 1948, and it experienced sporadic production through to final topography in 1962 (last operated by Pittsburgh-Pacific). Total shipments in the period 1924-1962 were 4.6 million tons of gravity concentrates.

Description:

The Mary Ellen mine is perhaps most notable for its exposures of the algal submember of the Upper Cherty (equivalent to the UC-6 submember at the Thunderbird Mine, and the I submember of Gundersen and Schwartz (1962)). Here, stromatolites occur as mounds of fossilized algal colonies separated by intraformational jasper conglomerates. The algal and conglomeratic units exhibit a combined thickness of 2-20 feet. This horizon occurs only in the eastern half of the Mesabi Iron Range, pinching out in the vicinity of Hibbing. Planavsky and others (2009) attribute the stromatolites to Fe-oxidizing bacteria present in the Animikie Basin and in similar settings world-wide, where microbial communities proliferated at specific shallow-water redox boundaries in late Paleoproterozoic oceans (see Fig. 14, below). The Mary Ellen mine is noted for the abundance of colonies of finely-laminated, small (~1cm diameter) digitate, columnar stromatolites. They occur in mound-like aggregations that appear to have been buried in-situ on the seafloor, in contrast to the largely resedimented oncoliths comprising the algal chert submember farther to the west.

Discussion:

The presence of stromatolites and intraformational conglomerate at this stratigraphic horizon within the iron-formation is consistent with extremely shallow water, and perhaps even emergent (subaerial) conditions. This likely represents maximum marine regression during the transgressive-regressive-transgressive cycle that characterizes deposition of the Biwabik Iron Formation. The carbonate rocks that comprise the uppermost Upper Slaty member of the iron-formation (submember US-2), though enigmatic, may relate to development of a second regression.

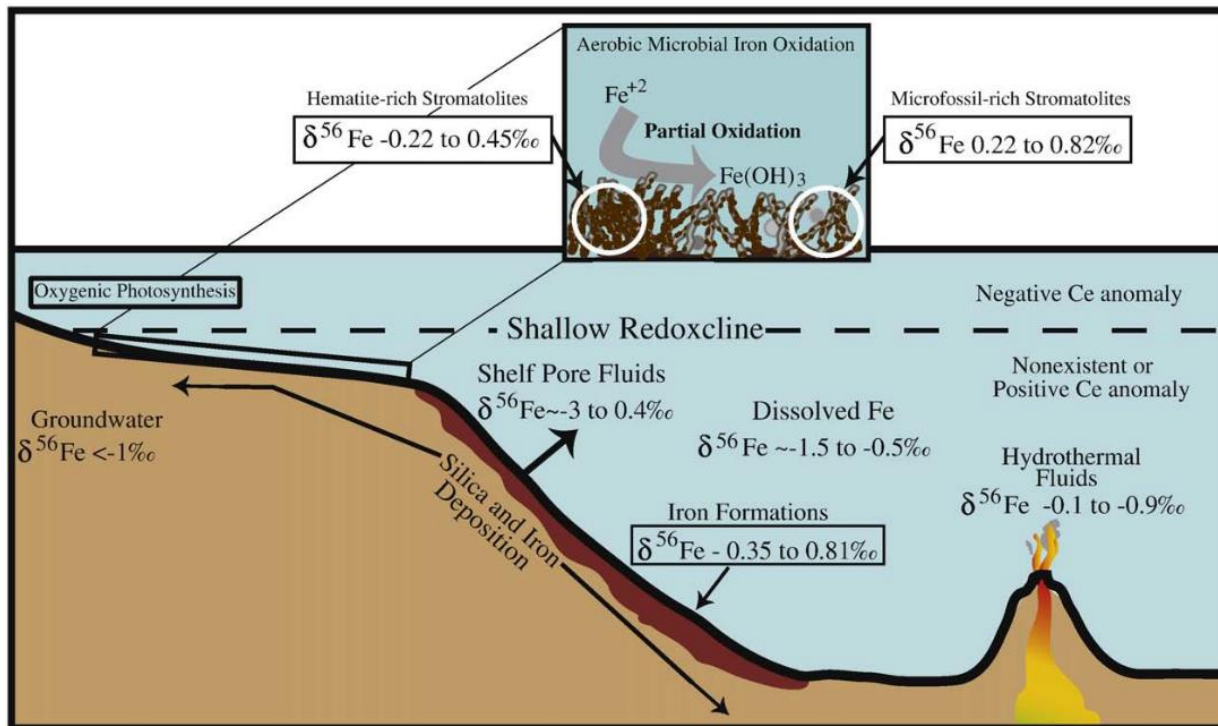


Figure 14. Model of stromatolite depositional environment (Planavsky and others, 2009).

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FIELD TRIP 2

Wednesday, May 14, 2014

A WALK IN THE PARK—NEOARCHEAN GEOLOGY OF LAKE VERMILION STATE PARK

LEADERS:

George J. Hudak (Natural Resources Research Institute – Duluth)

Amy Radakovich (Minnesota Geological Survey)

Geoff Pignotta and Kelly Schwierske (Geology Dept., University of Wisconsin - Eau Claire)

INTRODUCTION

The Vermilion District of northeastern Minnesota contains one of the classic greenstone belts in the United States. The district comprises the southwestern part of the Wawa Abitibi Terrane (Stott et al., 2007; Stott and Mueller, 2009) which encompasses Neoproterozoic metavolcanic, metasedimentary, and meta-intrusive rocks that extend northeastward through northwestern Ontario and Quebec. In Canada, this terrane hosts numerous volcanogenic massive sulfide deposits (e.g. Winston Lake, Geco, Noranda), gold-rich volcanogenic massive sulfide deposits (Horne (Noranda camp), Bousquet 2 – LaRonde 1, LaRonde-Penna; Mercier-Langevin et al., 2010), as well as a large number of lode (orogenic) gold deposits (for example, in the Hemlo, Timmins, and Kirkland Lake camps). The Vermilion District is known for its numerous, previously mined massive hematitic iron ore deposits (including the Pioneer Mine in Ely and the Soudan Mine in Soudan) which locally occur within regional extensive Algoma-type banded iron formations. To date, no volcanogenic massive sulfide, gold-rich volcanogenic massive sulfide, or lode gold deposits have been discovered in the Vermilion District, although several studies (Peterson and Jirsa, 1999; Peterson, 2001; Hudak et al., 2002a; Peterson and Patelke, 2003; Hoffman, 2007; Hudak et al., 2007; Hudak et al., 2012; Lodge et al., 2013) have indicated that evidence for volcanic, hydrothermal, and structural processes associated with these types of mineral deposits is present throughout the Vermilion District.

Since the late 1990's considerable geological research has been conducted in the region between Tower, MN (in the west) to Ely, MN (in the east) within the Vermilion District. Much of this research has been conducted to better understand the stratigraphy, structural geology, and economic geology of the belt, and the results of these studies have provided a solid foundation for the geological research that has, and currently is, taking place in Lake Vermilion State Park. Several recent Institute on Lake Superior Geology (ILSG) field trips (Hudak et al., 2004; Jirsa et al., 2004; Peterson and Patelke, 2004; Larson and Mooers, 2009; Peterson et al., 2009a; Jirsa and Hillman, 2009; Peterson et al., 2009b) describe these findings for specific areas in and around the Vermilion District.

The Vermilion District's iron ore mining heritage is currently preserved at two state parks located near Soudan, Minnesota. With the donation of land and infrastructure associated with the former Oliver Iron Mining Division's Soudan Mine by United States Steel to the State of Minnesota in 1965, Soudan Underground Mine State Park was established. This state park currently preserves the historical surface and underground workings from, as well as the wilderness adjacent to, Minnesota's oldest iron ore mine, the Soudan Mine. This mine operated from 1882 until December, 1962 and produced approximately 15.5 tons of hematitic iron ore. This popular tourist site continues to be the focus of a wide variety of research spanning geology, geochemistry, hydrogeology, biology, biochemistry and physics. Lake Vermilion State Park is Minnesota's newest state park, and comprises over 3,000 acres of land, including over five miles of undeveloped shoreline on Lake Vermilion (Bakst, 2013). In 2008, Minnesota State Legislature set aside \$20 million in bonding authority to buy, plan, and develop the park, which is located immediately east of Soudan Underground Mine State Park. The park was established in June 2010 after land was

purchased from U. S. Steel Corporation. At the present time, the park is undergoing considerable development, including establishment of trails, roads, and campsites. The park boasts a rich natural and human history, including a wide variety of ~2.7 billion year old rocks that were formed by a wide variety of genetic process, abundant wildlife, as well as archaeological evidence for human habitation dating back over 6,000 years. Additionally, considerable evidence for recent (within the past 140 years) mineral exploration efforts can be readily identified in the park.

In 2010 and 2011, students and faculty from the Precambrian Research Center at the University of Minnesota Duluth had the opportunity to conduct detailed (1:5000 scale) geological mapping in both Soudan Underground Mine State Park (Vallowe et al., 2010) and Lake Vermilion State Park (Radakovich et al., 2010; Heim et al., 2011). Twelve students (Nick Heim, Robert Kilduff, Chris Mahr, Charlie Parent, Molly Partridge, Rita Pierce, Amy Radakovich, Andrew Ritts, Christine Rahtz, Heather Scott, Andrew Vial, Spencer Young) and instructor George Hudak performed geological mapping in the northwestern (2010) and northeastern (2011) parts of Lake Vermilion State Park. Recently, Geoff Pignotta and Kelly Schwierske of the Geology Department at the University of Wisconsin Eau Claire compiled these geological maps and conducted lithogeochemical evaluations of several lithological units in the park (Schwierske et al., in press). This trip builds on these findings, and will be the first formal geology field trip in Lake Vermilion State Park. It will include a walk up-section through the stratigraphic sequence exposed along a single two-track trail that traverses the park. As well, two outcrops outside the state park boundary will be investigated, as they comprise classic outcrops that will add context to the geological story developed through observing rocks in the park.

REGIONAL GEOLOGIC SETTING

Supracrustal rocks in the Vermilion district consist of volcanic-dominated stratigraphic sequences of the Wawa Abitibi Terrane within the Superior Province of the Canadian Shield. Rocks of the Wawa Abitibi Terrane in northern Minnesota are divided on the basis of stratigraphic and structural setting into: (1) the Soudan belt, to the south, and (2) the Newton belt, to the north (Jirsa et al., 1992; Southwick et al., 1998). The boundary between these contrasting structural panels can be traced geophysically across the width of Minnesota, and was informally designated the Leech Lake structural discontinuity (Jirsa et al., 1992). In the region west and north of the Lake Vermilion State Park, the Leech Lake structural discontinuity occurs along the Mud Creek shear zone (Hudleston et al., 1988), small segments of the Vermilion and Wolf Lake faults (Sims and Southwick, 1985), and the Bear River fault (Jirsa et al., 1992). A simplified regional geological map of the Neo-Archean terranes of northeastern Minnesota and adjacent Ontario is presented in Figure 1.

The Soudan belt (Figure 2) contains large, broad folds involving calc-alkalic and tholeiitic volcanic strata overlain by, and locally interdigitated with, turbiditic rocks. In contrast, the Newton belt consists of elongate, northeast-trending, and mostly northward-younging volcanic and volcanoclastic sequences. Volcanic rocks of the Newton belt differ from those of the Soudan belt in containing locally abundant komatiitic flows and peridotitic sills. The two belts are fault-bounded, and the relationships between stratigraphic units within each belt are largely conformable (although faults obscure contacts locally). In its eastern extension, the Soudan belt is continuous with the Saganagons assemblage in Ontario and terminates against the Saganaga pluton and Northern Light Gneiss. The Newton belt extends discontinuously eastward into the Shebandowan District of Ontario to form the Greenwater and Burchell assemblages. Intrusive rocks in both belts vary from gabbroic and felsic porphyries demonstrably related to volcanism, to large plutons emplaced post-tectonically. Both districts contain unconformable, Timiskaming-type sequences composed of calc-alkalic volcanic rocks, conglomerates, and finer grained sedimentary rocks.

Lithostratigraphic units in the western Vermilion district (Table 1) include: (1) the Lower member, Soudan Iron-Formation member, and Upper member (Upper Ely) of the Ely Greenstone Formation, the Lake Vermilion Formation (including the informally named Britt and Gafvert Lake sequences), and the Knife Lake Group of the Soudan belt; (2) the Bass Lake sequence (Peterson and Jirsa, 1999) and the

Newton Lake Formation of the Newton belt; and, (3) syn- to post-tectonic granitoid intrusions of the Giants Range batholith, and a suite of post-tectonic alkalic stocks and plutons. Contacts between the different units are typically conformable, although considerable overlap in time and space is documented between volcanic and sedimentary sequences (Southwick, 1993).

Geochronological information for supracrustal and intrusive lithologies in the Vermilion District is relatively sparse (Figure 2). Peterson et al. (2001) obtained a U-Pb zircon age date of 2722 ± 0.9 Ma from a quartz-phyric rhyolite dome in the Fivemile Lake Sequence of the Lower Member of the Ely Greenstone Formation. Lodge et al. (2013) obtained a U-Pb zircon date of 2689.7 ± 0.8 Ma for a Gafvert Lake Sequence dacitic tuff breccia that occurs approximately 2m north of the contact with the Soudan Iron-Formation member of the Ely Greenstone Formation. As well, Lodge et al. (2013) obtained detrital zircon dates ranging from 2680-2690 Ma from greywackes that comprise the Lake Vermilion formation. This date confirms the source of the detritus in the Lake Vermilion Formation was derived locally from the volcanoclastic rocks comprising the Gafvert Lake Sequence. Jirsa et al. (2012) obtained a U-Pb age of 2690.7 ± 0.6 Ma for synvolcanic intrusions that cross-cut volcanoclastic rocks that comprise the Knife Lake Group. The upper part of the Knife Lake Group includes conglomerates which contain clasts derived from the Saganaga Tonalite, which has been dated by Driese et al. (2011) at 2690.83 ± 0.26 Ma. Peterson et al. (2001) also dated a non-foliated feldspar porphyry intruded into Newton Belt strata at $2683.1 \pm 1/-4$ Ma. This date provides a minimum age for the regional D2 deformation event that is described below.

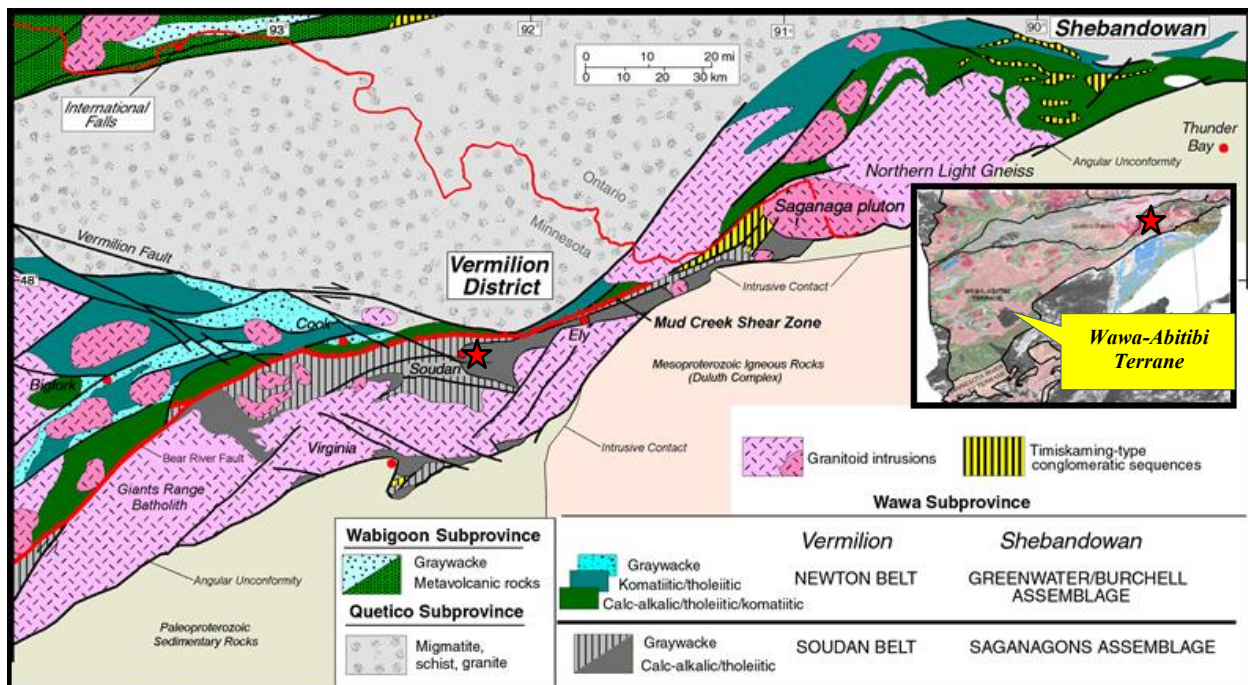


Figure 1. Simplified correlation map of Neoproterozoic assemblages in Minnesota and northwestern Ontario (after Peterson et al., 2001). Inset map illustrates location of the Wawa-Abitibi Terrane in Minnesota and northwestern Ontario (Stott et al., 2007). The Leach Lake structural discontinuity is illustrated in red. The red star symbols indicate location of Lake Vermilion State Park.

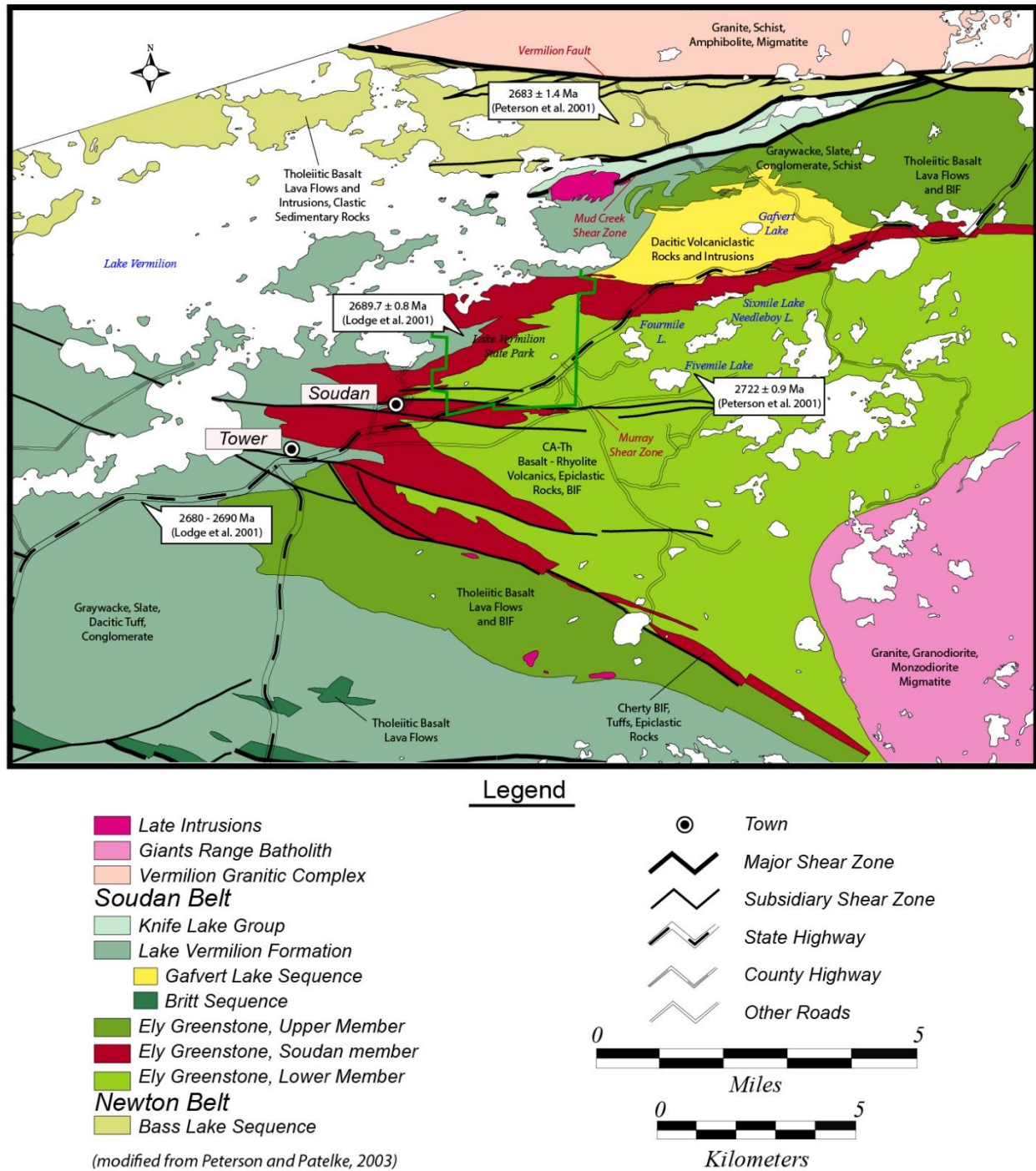


Figure 2. Generalized geology of the Vermilion District in the vicinity of the Tower-Soudan anticline (modified after Peterson, 2001). Locations, ages, and sources of U-Pb ages dates within the district are noted in the callout boxes. Generalized lithologies for each of the groups, formations or sequences are also noted.

Intrusive Rocks

Late Intrusions	Plutons and stocks of syenite, monzonite, diorite, and lamprophyre. A U-Pb zircon age date of a non-foliated feldspar porphyry intrusion in the Newton belt is 2683 ± 1.4 Ma (Peterson et al., 2001).
Vermilion Granitic Complex	Granite, schist, amphibolite, and schist-rich migmatite
Giants Range Batholith	Granite, granodiorite, monzodiorite, and schist-rich migmatite

Supracrustal Rocks**Newton Belt**

Newton Lake Formation	Tholeiitic and komatiitic basalt lava flows, intrusions, and clastic strata
Bass Lake Sequence	Tholeiitic basalt lava flows, iron-formation, and felsic porphyries

Soudan Belt

Knife Lake Group	Graywacke, slate, conglomerate, and sheared equivalents
Lake Vermilion Formation	Graywacke, slate, dacitic tuff, minor conglomerate. Detrital zircons from planar bedded, normal-graded resedimented volcanoclastic rocks have U-Pb age dates of 2680-2690 Ma (Lodge et al., 2013)
<i>Gafvert Lake Sequence</i>	Dacitic to rhyodacitic tuff, lapilli-tuff, tuff-breccia, and iron-formation. Basal dacite tuff-breccia deposits in Lake Vermilion State Park have U-Pb age date of 2689.7 ± 0.8 Ma (Lodge et al., 2013)
<i>Britt Sequence</i>	Tholeiitic basalt lava flows
Upper Member – Ely Greenstone	Tholeiitic basalt lava flows and iron-formation
Soudan Member – Ely Greenstone	Oxide-facies iron formation with intercollated basalt lava flows and felsic volcanoclastic rocks
Lower Member – Ely Greenstone	Calc-alkaline and tholeiitic basalt-rhyolite lava flows, tuffs, epiclastic rocks, and minor iron-formation
<i>Central Basalt Sequence</i>	Calc-alkaline to tholeiitic sparsely amygdaloidal basalt and minor basaltic andesite lava flows with MORB-like or back arc basin-like chemical affinities within 100-200 meters of the overlying Soudan Member iron-formation; FII- and FIIIa-type felsic volcanic and volcanoclastic rocks
<i>Fivemile Lake Sequence</i>	Calc-alkaline to transitional moderately to highly vesicular basalt and andesite lava flows and volcanoclastic rocks with arc-like chemical affinities: FI-, FII-, and FIV-type felsic volcanic and volcanoclastic rocks. Rhyolite dome at near Fivemile Lake has U-Pb age date of 2722.6 ± 0.9 Ma (Peterson et al., 2001). Epithermal-like zinc stringer mineralization is present near Fivemile Lake (Hudak et al., 2002a).

Table 1. Lithostratigraphic units within the western Vermilion District (modified after Peterson and Jirsa, 1999; Peterson et al., 2009; Hudak et al., 2012).

STRUCTURAL GEOLOGY

The structural geology of the Vermilion District has been well described by Peterson et al. (2009), and is reproduced below.

Periods of generally N-S directed compression resulted in three major regional deformation events in the Neoarchean terranes of northern Minnesota. The earliest deformation event (D1) produced broad, locally recumbent folds within the Soudan belt and major fault zones throughout the region. In the Newton belt, D1 was accommodated by thrust imbrication of large crustal blocks, resulting in mainly northward stratigraphic facing. Field relationships indicate that uplift, faulting, and the deposition of Timiskaming-type clastic sedimentary sequences in local fault-bounded basins occurred late in D1 deformation (Jirsa, 2000). A large, map-scale structure related to D1 deformation in the western Vermilion District is the Tower-Soudan Anticline, which is a west-plunging anticline within which the axis and plunge changes orientation along strike from nearly vertical in basalts to shallow NE plunging in the western sedimentary rocks. Axial-planar cleavage associated with this early fold typically is lacking, although Bauer (1985), Hooper and Ojakangas (1971), Hudleston (1976), and Jirsa et al. (1992) have described early cleavage (S1) locally.

A second deformation event (D2) associated with synchronous regional metamorphism resulted in foliation development and structures having largely dextral asymmetry. D2 is constrained in the Vermilion District to the time period 2674 to 2685 Ma (Boerboom and Zartman, 1993), and between about 2680 and 2685 Ma in the Shebandowan (Corfu and Stott, 1998). Because D2 deformation affected all of the supracrustal rocks in the area and is reasonably constrained by geochronology, the regional foliation (S2) can be used in the field to temporally relate other structural, intrusive, and deformation events. The relationship between S2 fabric and shear structures indicates that most shearing occurred relatively late in the D2 event. Major shearing that produced the Mud Creek and related shear zones is attributed to the late stages of D2 dextral transpression.

Structures related to the third deformation event (D3), which led to juxtaposition of the Wawa Abitibi and Quetico terranes (Peterson and Patelke, 2003), include abundant NE- and NW-trending faults that dissect the stratigraphic assemblages. Named structures related to D3 include the NE-trending Waasa and Camp Rivard faults east of the Soudan Mine area, and the WNW-trending, crustal-scale Vermilion and related faults that form the Wawa-Quetico Subprovince boundary.

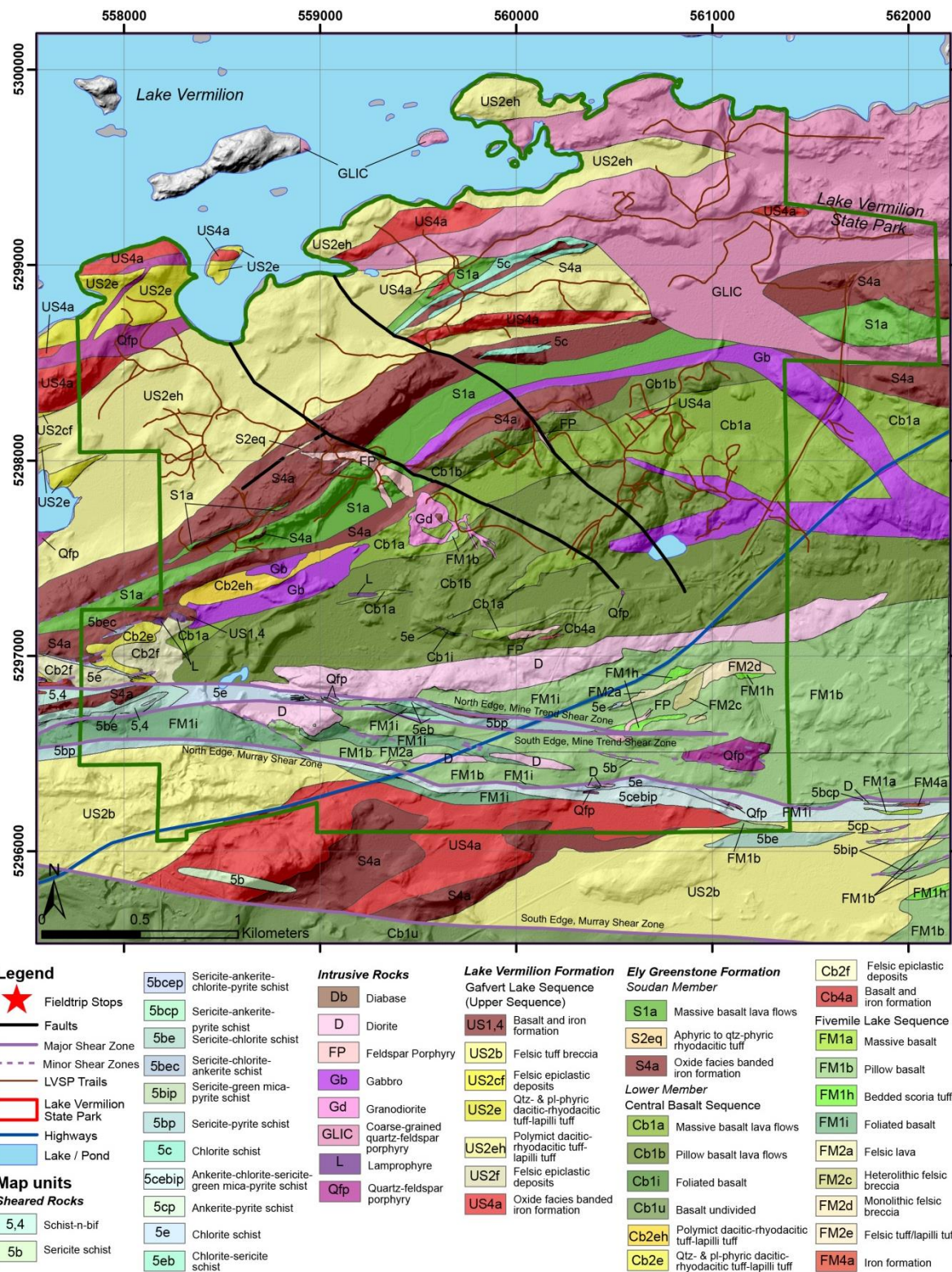


Figure 3. Geologic map of Lake Vermilion State Park (after Peterson and Patelke, 2003; Radakovich et al., 2010; Heim et al., 2011; Schwierske et al., in press).

GEOLOGY OF LAKE VERMILION STATE PARK

Lake Vermilion State Park contains a variety of supracrustal and intrusive lithological units. Supracrustal rocks that can be observed in the park (Figure 3) include the Lower Member of the Ely Greenstone Formation (both the Fivemile Lake and Central Basalt Sequences), the Soudan Member of the Ely Greenstone Formation, and the Gafvert Lake Sequence of the Lake

Vermilion Formation. As well, a wide variety of syn- and post-volcanic intrusive rocks crop out within the park, including diabase, gabbro, diorite, granodiorite, various types of quartz-feldspar porphyries, feldspar-porphyries, and lamprophyre (Peterson and Patelke, 2003; Radakovich et al., 2010; Heim et al., 2011; Schwierske et al., in press).

Two northwest-trending faults (which based on detailed mapping (Peterson and Patelke, 2003; Radakovich et al., 2010; Heim et al., 2011) possess higher concentrations of synvolcanic hydrothermal alteration mineral assemblages proximal to the structures) appear to be reactivated synvolcanic structures that offset stratigraphic units in the central and northwestern part of the park. As well, Peterson and Patelke (2003) have identified four major, more-or-less east-west trending shear zones that displace stratigraphy in the southern one-third of the park. The northernmost two shear zones represent the northern and southern limits of the Mine Trend Shear Zone, which extends westward into Soudan Underground Mine State Park, and appears to have played a key role in the development of hematite-rich iron orebodies that were historically mined there. The Mine Trend Shear Zone displaces lithological units higher in the stratigraphic sequence to the east. The southern two shear zones represent the northern and southern edges of the Murray Shear Zone (Peterson and Patelke, 2003). This fault system also displaces rocks higher in the stratigraphic sequence to the east. Rocks sandwiched between the southern edge of the Mine Trend Shear Zone and the northern edge of the Murray Shear Zone are in a structural domain known as the Linking Zone (Peterson and Patelke, 2003). According to Peterson and Patelke (2003), the net slip along the Mine Trend Shear Zone may have been as much as 7 km, whereas the net slip along the Murray Shear Zone may have been as much as 13.8 km (Table 2).

Shear Zone	Lineation Plunge	Calculated Displacement (Kilometers)		
		Strike Slip	Dip Slip	Net Slip
Mine Trend Shear Zone	72.6°	2.1	6.7	7.0
	72.6°	1.0	3.2	3.3
Murray Shear Zone	71.0°	4.5	13.1	13.8
	71.0°	3.0	8.7	9.2

Table 2. Calculated displacements among the Mine Trend and Murray Shear zones (Peterson and Patelke, 2003). Ranges of values were calculated geometrically by using the average plunges of lineations associated with the shear zones, and two measured lines of possible correlative stratigraphy offset by the bounding shear zones. See Peterson and Patelke (2003) for further details.

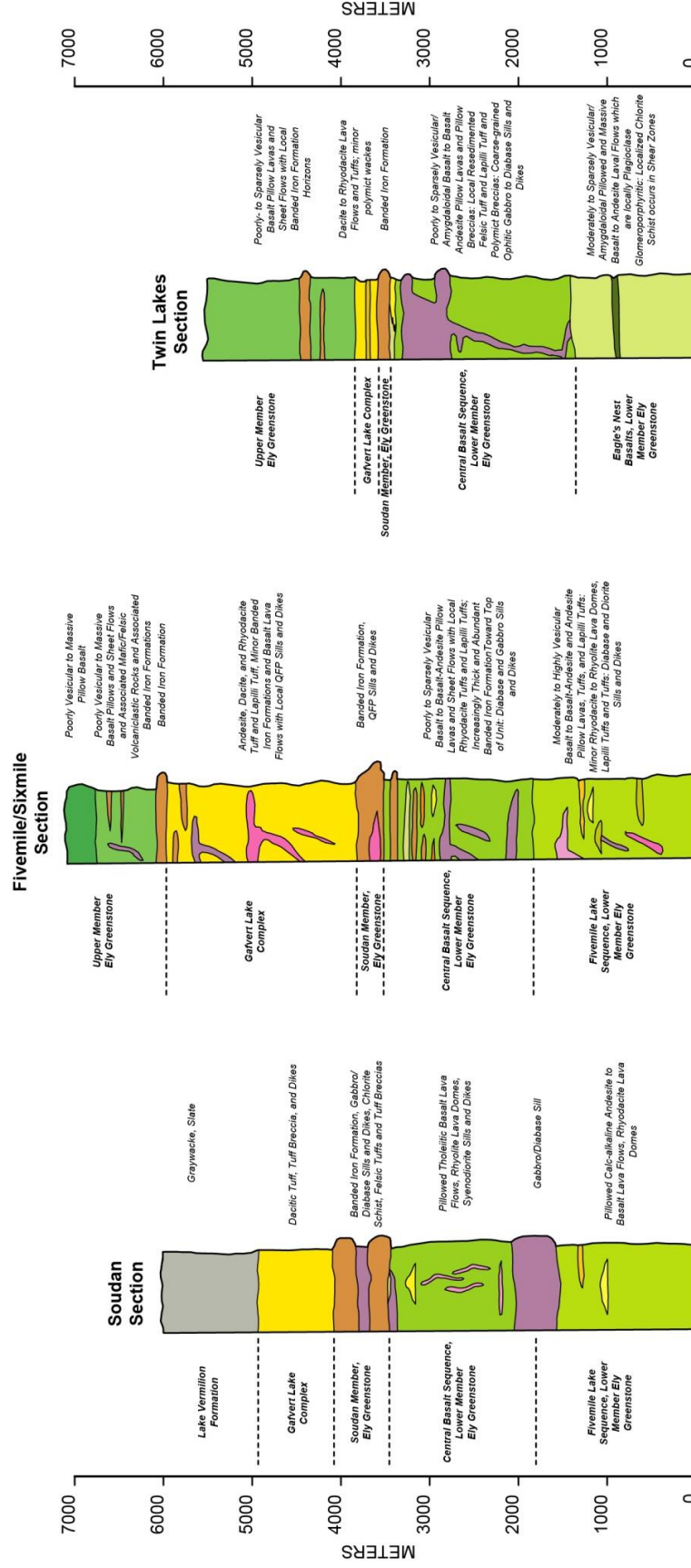


Figure 4. Regional stratigraphic correlations across the Vermilion District (after Hudak et al., 2012).

During our field trip, we will be observing exposures of various lithologic units that occur within the Vermilion District. Overall, lithological units observed in Lake Vermilion State Park correlate well with lithological units mapped regionally in the Vermilion District. A diagram (Hudak et al., 2012) illustrating stratigraphic columns from the Soudan Mine Area (in the west), the Fivemile Lake area, and the Twin Lakes area (in the east) is provided in Figure 4. Stratigraphic and intrusive units that occur within Lake Vermilion State Park are described below in order from oldest to youngest units.

The Fivemile Lake Sequence, Lower Member of the Ely Greenstone Formation

The Fivemile Lake Sequence comprises the lowermost mafic to intermediate and felsic volcanic and volcanoclastic lithologies associated with the Lower Member of the Ely Greenstone Formation. This generally east-west striking, north-topping sequence is dominated by well-vesiculated basaltic to andesitic pillow lavas (Hudak et al., 2007; Peterson et al., 2009; Hudak et al., 2012) that display bun, mattress, and lobe morphologies using the nomenclature of Dimroth et al. (1978). Locally, these pillow lavas display exceptional multiple selvages (Hudak et al., 2002b). Multiple pillow selvage morphologies have been interpreted as an indication of eruption in shallow water active volcanic settings (Kawachi and Pringle, 1988). Within Lake Vermilion State Park, horizons of Fivemile Lake pillow lavas are up to 1100m thick. Subordinate massive sheet lava flows associated with the Fivemile Sequence have been identified by Hoffman (2007) in one locale near Soudan. As well, numerous relatively thin horizons of massive to bedded basalt tuff, lapilli-tuff, and lapillistone deposits are present in the southwest ¼ of Section 25 in the south-central part of Lake Vermilion State Park. According to Hoffman (2007), these deposits vary from 50-150 meters thick, have a strike length of up to 350 meters, and comprise poorly-sorted and poorly-graded, matrix-supported, thickly to very-thickly bedded mafic pyroclastic deposits containing up to 65% lapilli- (64mm) to block-sized (>64mm) scoria fragments.

Hudak et al (2007, 2012) have evaluated the lithogeochemical characteristics of Fivemile Lake Sequence mafic to intermediate lava flows and have found them to be dominantly calc-alkaline to transitional basalt and andesite/basalt using the classification schemes of Ross and Bedard (2009) and Pearce (1996). These rocks also are characterized by significant negative niobium (Nb) anomalies when plotted on primitive mantle-normalized spider diagrams. This suggests derivation of the magmas associated with the Fivemile Lake sequence in an arc-like volcanic terrane.

Felsic volcanic and volcanoclastic rocks also occur within the Fivemile Lake Sequence, and crop out within the southern one-third of Lake Vermilion State Park based on mapping completed by Peterson and Patelke (2003). These include coherent and volcanoclastic facies dacitic to rhyolitic lithologies including lava flows, monolithic and heterolithic breccia deposits, and tuff and lapilli- tuff deposits. Lithogeochemically, Hudak (2007) and Hudak et al. (2012) have shown felsic rocks in the Fivemile Lake Sequence to be calc-alkaline to transitional andesites to rhyolites using the classification schemes of Pearce (1996) and Ross and Bedard (2009), respectively. As well, these felsic rocks have trace element characteristics of FI, FII, and FIV rhyolites based on classifications from Hart et al. (2004).

One relatively thin horizon (<20 meters thick) of oxide-facies iron-formation identified as being within the Fivemile Lake Sequence has been observed proximal to the northern margin of the Murray Shear Zone approximately 600 meters east of the Lake Vermilion State Park boundary. This iron formation typically occurs as localized, thin (<3m thick) horizons interbedded with Fivemile Lake Sequence pillowed lava flows (Peterson and Patelke, 2003). The various lithofacies comprising the Fivemile Lake Sequence of the Lower Member of the Ely Greenstone Formation are summarized in Table 3.

Lithofacies Associated with the Fivemile Lake Sequence, Lower Member, Ely Greenstone Formation	
<i>Unit Symbol (Figure 3)</i>	<i>Lithofacies</i>
FM1a	Massive Basalt - Andesite
FM1b	Pillow Basalt - Andesite
FM1h	Scoriaceous Basalt –Andesite Tuff and Lapilli-Tuff
FM1i	Foliated Basalt-Andesite
FM2a	Coherent Dacite – Rhyolite (lava flows and lava domes)
FM2c	Felsic Polymict Breccia
FM2d	Felsic Monolithic Breccia
FM2e	Dacite-Rhyolite Tuff
FM4a	Oxide-facies Banded Iron-Formation

Table 3. Lithofacies and map symbols associated with lithologies comprising the Fivemile Lake Sequence of the Lower Member of the Ely Greenstone Formation.

The Central Basalt Sequence, Lower Member of the Ely Greenstone Formation

The Central Basalt Sequence crops out in the east-central part of Lake Vermilion State Park, and is composed of massive and pillowed basalt lava flows, structurally deformed foliated basalt, and local thin (up to 3 meters thick) horizons of Algoma-type banded iron-formation. Within Lake Vermilion State Park, the Central Basalt Sequence varies from approximately 300-1000 meters in stratigraphic thickness. The Central Basalt Sequence mafic lava flows appear to be regionally correlative with basaltic lava flows comprising the Armstrong Lake Sequence in the northernmost parts of the Eagles Nest Quadrangle mapped by Jirsa et al., 2001 (Peterson and Patelke, 2003).

The Central Basalt Sequence mafic lava flows can be distinguished from the Fivemile Lake Sequence mafic-intermediate lava flows using the following criteria: 1) the Central Basalt Sequence mafic lava flows commonly comprise exceptionally well-preserved primary volcanic textures - such textures are generally not present in the Fivemile Lake Sequence mafic flows due to destruction of these textures from a combination of synvolcanic hydrothermal alteration combined with recrystallization during greenschist-facies regional metamorphism; 2) the Central Basalt Sequence mafic volcanic rocks tend to be dark green to green colored, whereas the Fivemile Lake Sequence mafic-intermediate volcanic rocks typically vary from gray green to blueish green in color; 3) the Central Basalt sequence mafic volcanic rocks tend to lack amygdules or be sparsely amygdaloidal, whereas the Fivemile Lake Sequence mafic-intermediate volcanic rocks tend to contain abundant amygdules; and 4) to date, multiple selvege pillow lavas have not been identified in the Central Basalt Sequence, whereas they are locally abundant within the Fivemile Lake Sequence.

Within Lake Vermilion State Park, the Central Basalt Sequence is composed primarily of pillowed basalt lava flows. These mafic volcanic rocks are medium green to dark green in color and tend to be sparsely amygdaloidal (<5% 2mm-1cm rounded to oval gray quartz-filled amygdules). Dark green, locally exceptionally well-preserved interpillow hyaloclastite deposits, ranging from 1-5cm wide, separate individual pillow structures. Locally these rocks are moderately- to strongly quartz- and epidote-altered. As well, in the southeastern part of the park, hydrothermally altered interpillow hyaloclastite deposits containing abundant andradite garnets have been identified. Massive basalt lava flows (interpreted as sheet flow facies lava flows) are also quite common, and comprise green to dark green, aphyric to sparsely-plagioclase phyric basalt. Foliated basalts locally occur in close proximity to shear zones present in the park.

Hudak et al (2007, 2012) have evaluated the lithogeochemical characteristics of mafic lava flows in the Central Basalt Sequence in the vicinities of Needleboy and Sixmile Lakes, which are located

approximately 4-5 kilometers east of the eastern boundary of Lake Vermilion State Park. These researchers have found them to range from calc-alkaline to tholeiitic basalt and andesite/basalt using the classification schemes of Ross and Bedard (2009) and Pearce (1996), respectively. Hudak et al. (2007) first observed that Central Basalt Sequence mafic flows could be divided into two types based on rare-earth element characteristics. The first of these types is characterized by calc-alkaline to transitional compositions with arc-like chondrite- and primitive-mantle-normalized rare earth element spider diagrams. The second type comprises tholeiitic basalt characterized by flat chondrite-normalized and primitive mantle-normalized rare earth spider diagrams that are characteristic of mafic volcanic rocks erupted within mid-ocean ridge (MORB) or back-arc basin (BABB) extensional tectonic environments. Detailed mapping indicates that these tholeiitic, MORB/BABB compositions only occur within 200 meters of the lower contact with the overlying Soudan Member Iron Formation. Hudak et al. (2007, 2012) have used both these lithogeochemical results, and results from detailed regional mapping at Lake Vermilion State Park, the Needleboy Lake-Sixmile Lake area, the Twin Lakes area (located approximately 14km east of the eastern boundary of Lake Vermilion State Park), and the Purvis Lake area (on the southern limb of the Tower-Soudan anticline approximately 5km south-southeast of the southern boundary of Lake Vermilion State Park) to indicate that the major hydrothermal event that led to the formation of the Soudan Member Algoma-type iron-formation occurred immediately following the opening of a nascent rift or back-arc basin environment during the youngest part of Central Basalt Sequence mafic volcanism.

Hudak et al. (2002b) and Hoffman (2007) have identified several localized occurrences of felsic coherent and volcanoclastic strata within the Central Basalt Sequence to the south and east of Lake Vermilion State Park. Hudak et al. (2007, 2012) have evaluated the lithogeochemical characteristics of these rocks, and have found them to be calc-alkaline andesite to rhyolite/dacite using the classification schemes of Ross and Bedard (2009) and Pearce (1996), respectively. As well, these felsic rocks have trace element characteristics of FII and FIIIa rhyolites based on classifications from Hart et al. (2004). The various lithofacies comprising the Central Basalt Sequence of the Lower Member of the Ely Greenstone Formation are summarized in Table 4.

Lithofacies Associated with the Central Basalt Sequence, Lower Member, Ely Greenstone Formation	
<i>Unit Symbol (Figure 3)</i>	<i>Lithofacies</i>
Cb1a	Massive Basalt
Cb1b	Pillow Basalt
Cb1i	Foliated Basalt
Cb1u	Undivided Basalt
Cb2eh	Polymict Dacite-Rhyodacite Tuff and Lapilli-Tuff
Cb2e	Dacitic-Rhyodacite Tuff and Lapilli-Tuff
Cb2f	Felsic Epiclastic Deposits
Cb4a	Interbedded Oxide-facies Banded Iron-Formation and Basalt

Table 4. Lithofacies and map symbols associated with lithologies comprising the Central Basalt Sequence of the Lower Member of the Ely Greenstone Formation.

The Soudan Member of the Ely Greenstone Formation

The Soudan Member of the Ely Greenstone formation is dominantly composed of laminated to thinly bedded Algoma-type oxide facies banded iron-formation, with subordinate, locally interstratified, sparsely amygdaloidal massive to pillowed basalt lava flows and resedimented felsic tuff deposits. Regionally, the stratigraphic thickness of the Soudan Member of the Ely Greenstone Formation varies

from 50-3,000 meters, with an average stratigraphic thickness of approximately 700 meters (Peterson et al., 2009). Within Lake Vermilion State Park, the Soudan Member ranges in stratigraphic thickness from approximately 300 – 680 meters in thickness. Individual horizons of oxide-facies iron formation range from approximately 70-345 meters thick, whereas the Soudan basalt lava flow units range from approximately 60-300 meters in thickness.

A gradational contact over several tens of meters to two hundred meters occurs between the underlying Central Basalt Sequence rocks and the overlying oxide facies iron-formations of the Soudan Member. This transitional zone is characterized by a decrease in abundance of basalt lava flows and associated volcanoclastic rocks, and an increase in the abundance and thickness of oxide-facies iron-formation horizons, as one moves toward the basal contact of the Soudan Member (Hudak et al., 2002b; Peterson and Patelke, 2003; Hudak et al., 2007; Hoffman, 2007; Hudak et al., 2012). Several characteristics suggest that the Soudan Member was deposited in relatively quiet water in a relative deep subaqueous environment (>200m, probably greater than 1400 m). This evidence includes: 1) a lack of primary mafic or felsic pyroclastic deposits within the stratigraphic sequence; 2) a lack of multiple-selvege pillow lavas in the stratigraphic sequence; 3) planar laminations and bedding combined with an absence of any wave-associated sedimentary bedforms within both the chemical and clastic sedimentary rocks within the sequence; and 4) lithological and geochemical evidence for the development of an extensional tectonic environment that resulted in deepening of the depositional environment in the uppermost sections of the stratigraphically underlying Central Basalt Sequence.

Within Lake Vermilion State Park, the Soudan Member oxide-facies banded iron-formation is planar laminated to medium-bedded, with black magnetite-rich horizons, light gray to black chert horizons, red to blueish-black hematite-rich horizons, and red jasper horizons defining the bedding. Locally, very tight, chaotically oriented folds, resulting from syn-depositional soft sediment deformation and subsequent tectonic deformation, are present. These iron formation deposits can be intimately interbedded with basalt lava flows such that mapping individual iron-formation and basalt horizons is impossible at 1:5000 scale. Where this occurs, these rocks have been mapped as a stratigraphic unit called “Basalt and Iron-Formation” by Peterson and Patelke (2003). Basalt lava flows associated with the Soudan Member of the Lower Ely Greenstone are characterized by a medium green to dark green color. They are aphyric to sparsely plagioclase ± pyroxene (now actinolite)-phyric. Plagioclase phenocrysts are present in abundances up to 3%, are typically less than or equal to 1mm in length, and vary from subhedral to euhedral tabular in morphology. Locally, 5-7% dark green actinolite pseudomorphs of pyroxene phenocrysts may be present. Where amygdaloidal, the unit contains up to 7% oval to round, light gray quartz-filled amygdules ranging from <1-4mm in diameter. The various lithofacies comprising the Soudan Member of the Ely Greenstone Formation are summarized in Table 5.

Lithofacies Associated with the Soudan Member, Ely Greenstone Formation	
<i>Unit Symbol (Figure 3)</i>	<i>Lithofacies</i>
S1a	Massive Basalt
S2eq	Aphyric- to Quartz-phyric Rhyodacite Tuff
S4a	Oxide Facies Banded Iron-Formation

Table 5. Lithofacies and map symbols associated with lithologies comprising the Soudan Member of the Ely Greenstone Formation.

The Gafvert Lake Sequence, Lake Vermilion Formation

The Gafvert Lake Sequence (mapped as the “Upper Sequence” by Peterson and Patelke, 2003; Radakovich et al., 2010; and Heim et al., 2011) comprises dacitic to rhyodacitic volcanoclastic and epiclastic rocks that are locally interbedded with Algoma-type banded iron-formation and chert deposits.

This sequence, which is part of the Lake Vermilion Formation, has been found to unconformably overlie the Soudan Member of the Ely Greenstone Formation in the north-central and northwestern parts of Lake Vermilion State Park based on recent mapping and geochronological evidence reported by Lodge et al. (2013). Within Lake Vermilion State Park, the overall stratigraphic thickness of the Gafvert Lake Sequence is up to approximately 1300 meters thick, with individual felsic volcanoclastic deposits having stratigraphic thicknesses ranging from approximately 75 – 400 meters thick, and individual Algoma-type oxide facies banded iron formations and associated massive- to bedded chert deposits ranging from 25-250 meters and up to 175 meters in stratigraphic thickness, respectively. To the west in Soudan Underground Mine State Park, the Gafvert Lake Sequence is locally interlayered with, and overlain by, greywacke deposits associated with the Lake Vermilion Formation.

Within Lake Vermilion State Park, several lithofacies comprise the Gafvert Lake Sequence. The basal member of this sequence comprises massive, very-thickly bedded, quartz- and plagioclase-phyric polymict dacitic to rhyodacitic tuff, lapilli-tuff, and tuff-breccia deposits. These light gray, non-sorted, non-graded, matrix-supported deposits contain 3-8% 1-2mm (rare 3mm) pale gray anhedral to subhedral quartz phenocrysts, 10-15% <1-2mm subhedral to euhedral tabular plagioclase phenocrysts, and a wide variety of lapilli- to block-sized clasts including: 1) 10-20% 1-10 cm quartz- and plagioclase-phyric coherent dacite to rhyodacite lapilli and blocks; 2) 5-7% $\leq$ 3cm diameter pale gray-green lens-shaped, locally quartz- and plagioclase-phyric pumice lapilli; 3) up to 1% dark gray to light gray angular chert lapilli ranging from 0.5-3cm in diameter; and 4) 1-3% 0.5-5cm dark gray to black to red magnetite-rich, hematite-rich, or jasper-rich banded iron formation lapilli. These deposits are overlain by, and interbedded with, light gray, matrix-supported, non-sorted and non-graded quartz- and plagioclase-phyric dacitic to rhyodacitic tuff deposits which contain 10-25% 1-3mm subhedral to euhedral tabular plagioclase phenocrysts, 1-3% 1-3mm subhedral to anhedral, commonly broken, quartz phenocrysts, as well as 10-15% subangular quartz- and plagioclase-phyric coherent dacite to rhyodacite lapilli and up to 5% locally quartz- and plagioclase-phyric pumice lapilli. Spectacular felsic epiclastic deposits comprising polymict volcanoclastic conglomerates and lithic sandstones are also present in the Gafvert Lake Sequence and crop out west of Lake Vermilion State Park in Stunz Bay (Radakovich et al., 2010).

In addition to felsic volcanoclastic and epiclastic rocks, two types of chemical sedimentary rocks have also been identified in the Gafvert Lake Sequence. These include laminated to medium-bedded Algoma-type banded iron formation that varies from red (hematite- and jasper-rich) to dark gray (magnetite-rich) to light gray (chert-rich) in color. Immediately west of the Lake Vermilion State Park boundary, light gray to black laminated to very thickly bedded black chert deposits are present. These chert deposits may represent the distal facies equivalent of Algoma-type banded iron-formation horizons that are present in the northeast part of Lake Vermilion State Park south of Cobble Bay.

A limited number of Gafvert Lake Succession felsic volcanoclastic rocks have been studied by lithogeochemical means by Geoff Pignotta and Kelly Schwierske at the University of Wisconsin Eau Claire (Figure 5). These researchers (Schwierske et al., in press) have found that the volcanoclastic and epiclastic deposits associated with the Gafvert Lake Sequence consistently plot as rhyodacite/dacite in composition when using the immobile element lithological classification scheme of Winchester and Floyd (1977). These compositions are very similar to the composition of a quartz- $\pm$ plagioclase-phyric rhyodacite sill that crops out in the northeastern part of Lake Vermilion State Park (see Field Trip Stop 9 below), although the sill has consistently higher Nb/Y ratios than the Gafvert Lake volcanoclastic and epiclastic rocks. This sill may represent a synvolcanic intrusion that is genetically related to the evolution of the Gafvert Lake sequence based on this lithogeochemical evidence, as well as field evidence from Peterson and Jirsa (1999), which indicates that this intrusion is most commonly observed within the Gafvert Lake Sequence where it is thickest, in proximity to Gafvert Lake west of the Mud Creek Road. Peterson (2001) has hypothesized that this intrusion represents feeder intrusions to a Gafvert Lake Sequence stratovolcano located in this area. The various lithofacies comprising the Gafvert Lake Member of the Lake Vermilion Formation are summarized in Table 6.

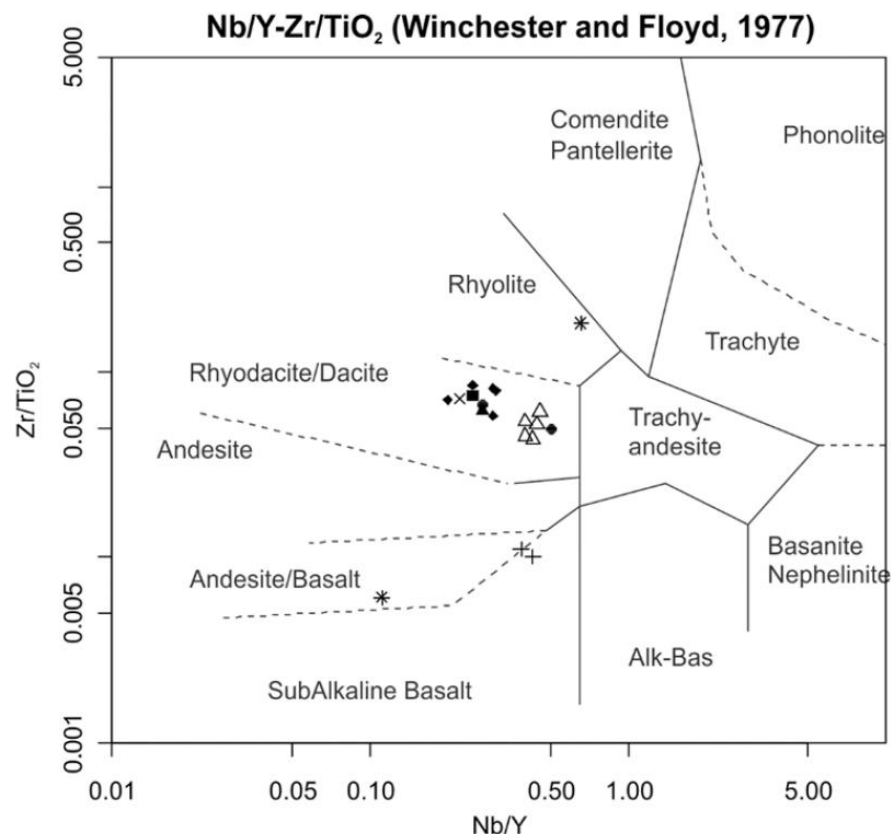


Figure 5. Chemical classification of various lithologies within Lake Vermilion State Park (Schwierske et al., in press) using the classification scheme of Winchester and Floyd (1977). Open triangles represent samples from a quartz- ± plagioclase-phyric rhyodacite/dacite sill in the northeastern part of Lake Vermilion State Park. The black squares, large black diamonds, and small black diamonds represent various Gafvert Lake Succession volcanoclastic and epiclastic rock units.

Lithofacies Associated with the Gafvert Lake Sequence, Lake Vermilion Formation	
<i>Unit Symbol (Figure 2-3)</i>	<i>Lithofacies</i>
US1,4	Interbedded Basalt and Oxide-facies Banded Iron-Formation
US2b	Dacite-Rhyodacite Tuff-breccia
US2cf/US2f	Dacite-Rhyodacite Epiclastic Deposits
US2e	Quartz- + Plagioclase-phyric Dacite-Rhyodacite Tuff/Lapilli-tuff
US2eh	Polymict Dacite-Rhyodacite Tuff/Lapilli-tuff
US4a	Oxide-facies Banded Iron-Formation

Table 6. Lithofacies and map symbols associated with lithologies comprising the Gafvert Lake Sequence of the Lake Vermilion Formation.

Intrusive Rocks

Eight types of intrusive bodies have been mapped within the boundaries of Lake Vermilion State Park (Peterson and Patelke, 2003; Hoffman, 2007; Radakovich et al., 2010; Heim et al., 2011; Figure 3). From oldest to youngest, these intrusions include:

- Gabbro (mapped by Peterson and Patelke, 2003; Hoffman, 2007; Radakovich et al., 2010; Heim et al., 2011, unit Gb) – Identified as sills throughout the central one-third of Lake Vermilion State Park, this unit is characterized by grayish-green to black, medium-grained equigranular gabbro that is locally highly magnetic and displays ophitic texture.
- Diabase (mapped by Peterson and Patelke, 2003; Hoffman, 2007, unit Db) – Identified in the southern one-third of Lake Vermilion State Park, this unit comprises black to dark green, fine-grained plagioclase-phyric diabase dikes and sills that have been interpreted to represent feeder dikes to mafic volcanic rocks located stratigraphically up-section.
- Coarsely porphyritic Quartz-Feldspar Porphyry (mapped by Heim et al., 2011, unit GLIC) – Identified in the northeastern part of Lake Vermilion State Park, this intrusion comprises light gray, massive, quartz $\pm$ plagioclase-phyric coherent rhyodacite. The light gray aphanitic groundmass contains 3-7% gray to light blue subhedral rounded to euhedral square quartz phenocrysts that range from 3-10mm in diameter, and 10% pale gray to tan, subhedral to euhedral tabular plagioclase phenocrysts ranging from 1-4mm in length. Similar intrusive rocks have been mapped in the vicinity of Needleboy and Sixmile Lakes by Hudak et al. (2002b), and near Gafvert Lake by Peterson and Jirsa (1999) and Peterson (2001).
- Diorite (mapped by Peterson and Patelke, 2003; Hoffman, 2007, unit D) – Occurs as a generally east-west striking sill in the southern one-third of Lake Vermilion State Park. Composed of gray to gray-green, fine- to medium-grained, equigranular diorite. This unit was informally named the “Sugar Mountain Diorite” by Peterson and Patelke (2003), and is notable for its massive, indurated nature and lack of prominent joints, veins, and alteration.
- Granodiorite (mapped by Peterson and Patelke, 2003; Hoffman, 2007, unit Gd) – Identified in the central part of Lake Vermilion State Park, and composed of whitish-pink to gray-green, fine- to medium-grained, commonly xenolith-rich granodiorite and locally hornblende granodiorite.
- Quartz-Feldspar Porphyry (mapped by Peterson and Patelke, 2003; Hoffman, 2007; Radakovich et al., 2010; Heim et al., 2011, unit Qfp) - Found locally throughout Lake Vermilion State Park, this intrusion comprises a light gray to pale green-gray groundmass that contains 20-25% 1-3mm (locally up to 5mm) subhedral to euhedral tabular plagioclase phenocrysts and 7-12% 1-3mm (locally up to 5mm) subhedral to euhedral gray-blue quartz phenocrysts
- Feldspar Porphyry (mapped by Peterson and Patelke, 2003; Hoffman, 2007; Radakovich et al., 2010; Heim et al., 2011, unit Fp) – Identified in the south and central parts of Lake Vermilion State Park, this intrusion is white to pink in color, and contains subhedral rounded to euhedral tabular 4mm feldspar phenocrysts and locally, subhedral to euhedral prismatic to tabular actinolite pseudomorphs of hornblende phenocrysts.
- Lamprophyre (mapped by Peterson and Patelke, 2003, unit L) – Located in the southwestern part of Lake Vermilion State Park, this intrusion is characterized by black, fine-grained, massive hornblende-feldspar rock that contains 10-15% fine hornblende needles in a gray-black to red matrix, as well as large (>25cm) rounded granite and supracrustal rock xenoliths.

TERMINOLOGY OF VOLCANICLASTIC ROCKS

It is important to note the terminology utilized in this field trip guide for: 1) volcaniclastic rocks; and 2) bedding characteristics. Use of consistent terminology is required in order to accurately describe these geological features.

Volcaniclastic rocks contain abundant volcanic material irrespective of their origin or depositional environment. Such rocks can be formed directly from volcanic eruptions (whether subaerial or subaqueous), result from resedimentation of non-lithified volcanic deposits (for example, resedimentation of pyroclasts prior to lithification), or result from weathering and resedimentation of pre-existing lithified volcanic rocks.

Primary (juvenile) volcaniclastic particles result directly from eruptive processes, and are of three types:

- *Pyroclasts*, which form by explosive fragmentation of magma into particles (including ash, highly vesiculated glass (pumice, scoria), crystals and crystal fragments, and lithic fragments);
- *Hydroclasts*, which form by explosive interaction with external water (via phreatic (steam only) and/or phreatomagmatic (steam and magma) explosions) or by non-explosive quenching and granulation of lava (for example, the formation of hyaloclastite fragments on the margins of submarine lava flows or intrusions into wet sediments); and
- *Autoclasts*, which form by frictional breakage of moving viscous lava flows (for example, to form carapace breccias on the margins of subaerial lava flows).

Based on these different types of fragmentation, four types of primary volcaniclastic deposits have been identified by White and Houghton (2006):

- *Pyroclastic deposits*, which are generated from volcanic plumes and jets or pyroclastic density currents as particles first come to rest. Deposition mechanisms associated with these processes include suspension settling, traction, or *en masse* freezing;
- *Autoclastic deposits*, which are generated during effusive volcanism when lava cools and fragments as a result of thermal processes, or recently cooled lava breaks during flow. Deposition for these types of rocks is under the influence of continued lava flowage;
- *Hyaloclastite deposits*, which are generated during effusive volcanism when magma or flowing lava is chilled and fragmented as a result of contact with water. Deposition of such deposits is under the influence of the continued emplacement of the lava in the presence of water; and
- *Peperite deposits*, which are generated when magma intrudes into unconsolidated clastic material and mingles with (generally wet) debris to form a volcaniclastic deposit. Deposition of peperite deposits takes place essentially *in-situ*.

Secondary volcaniclastic particles are known as epiclasts:

- *Epiclasts* are lithic clasts and/or crystals derived from physical weathering and erosion of pre-existing rocks. Epiclasts are volcaniclasts when the pre-existing rocks are volcanic.

In recent years, the terminology for volcaniclastic rocks has become increasingly confusing because different classification schemes (for example Fisher, 1961; Fisher 1966; Schmid, 1981; Cas and Wright, 1987; McPhie et al., 1993; White and Houghton, 2006) are preferentially used in different parts of the world, and terminology relating to volcaniclastic rocks is commonly misused. Four classification schemes have been used most commonly in the recent geological literature:

- Fisher (1961, 1966) – Classification based on particle size, particle formation, or particle fragmentation mechanism;
- Schmid (1981) – Particle type within the deposit;
- Cas and Wright (1987) – Mode of fragmentation and deposition; and
- McPhie et al. (1993) – Transport and deposition mechanisms.

According to R. V. Fisher (1998), the difficulties with volcaniclastic rock classification can be understood because “volcaniclastic rocks are essentially igneous on the way up and sedimentary on the way down”.

In fact, Fisher's thesis advisor, when observing the volcanoclastic rocks that were the focus of his thesis studies, indicated that they were "the ugliest and most undistinguished rocks I've seen in my 30 years of petrology!" As well, classification is especially difficult in ancient volcanoclastic rocks because key aspects of classification can be obscured by subsequent metamorphism and/or structural deformation (e.g. particle type, particle size) or because genetic processes cannot be ascertained unambiguously (e.g. transport and deposition mechanism, fragmentation mechanisms).

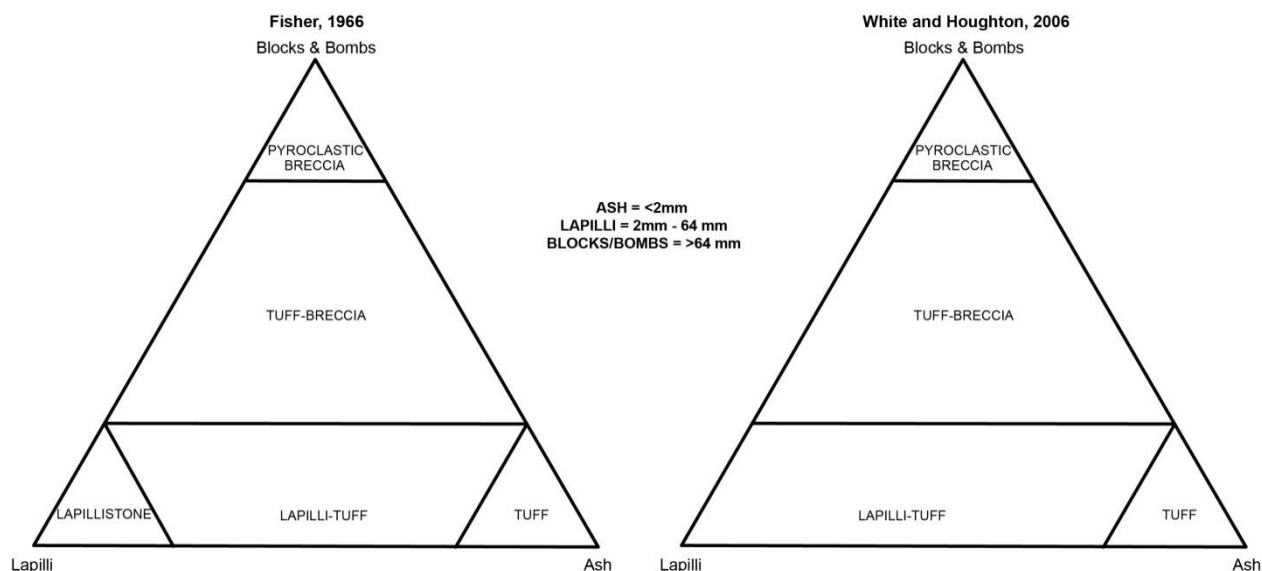


Figure 6. Volcanoclastic rock classification schemes of Fisher (1966) and White and Houghton (2006). This field trip guidebook will classify volcanoclastic rocks using Fisher's (1966) classification scheme.

For this field trip guidebook, we will utilize Fisher's (1966) classification (Figure 6) for volcanoclastic rocks. This classification scheme is based on the relative proportions of ash-sized material (< 2mm), lapilli-sized material (64mm), and blocks/bomb sized material (>64mm) in the rock. Both Gibson et al. (1999) and Mueller and White (2004) suggest that this classification be used for field-based rock classification (mapping, diamond drill core logging, petrography) of ancient volcanoclastic deposits for the following reasons:

- The classification scheme is "field-user friendly" because it accommodates both the historically important pyroclastic rock names and enables comparison at both the hand sample and thin section scale (Mueller and White, 2004);
- It is a Wentworth-based scale, and thus enables comparison of volcanoclastic deposits to sedimentary deposits; and
- Rock classification does not require knowledge of the specific transport mechanism or depositional processes involved with the genesis of the deposit.

More recently, White and Houghton have developed a modified version of Fisher's (1966) volcanoclastic classification scheme (Figure 6). The scheme is essentially equivalent to the Fisher (1966) scheme, with the exception that the lapilli-tuff field in the White and Houghton (2006) classification comprises the lapilli-tuff and lapillistone fields of Fisher's (1966).

Specific terms for bedding thicknesses are also used in this guidebook. The terms used, and their bedding thickness characteristics, have been adopted from McPhie et al. (1993) and include:

- Laminated <1 centimeters thick
- Very thinly bedded 1-3 centimeters thick
- Thinly bedded 3-10 centimeters thick
- Medium bedded 10-30 centimeters thick
- Thickly bedded 30-100 centimeters thick
- Very thickly bedded >100 centimeters thick

ROAD LOG AND FIELD TRIP STOPS

All stop locations for this field trip are given in Universal Transverse Mercator (UTM) coordinates, Zone 15N, using the North American Datum of 1983 (NAD83). Section subdivisions read from smallest to largest quarter; e.g., “NW, SE” should be read “NW quarter of the SE quarter.” The small topographic map insets illustrating field trip stop locations have been taken from the Tower and Soudan USGS 7.5-minute quadrangle maps. A selected number of field trip stops will take place outside Lake Vermilion State Park, with the majority of the stops taking place along a 4.5 mile traverse through the state park. More detailed geological maps with stop locations are given in Figures 7 and 8 later in this guidebook.

From Hibbing, our field trip route will proceed north and east from the Hibbing Park Hotel along Minnesota Highway 169 North. We will make one stop at a spectacular outcrop displaying various flow facies of Central Basalt flows that are located south of Lake Vermilion State Park on the south side of Highway 169 North (Figure 7). We will then proceed into Lake Vermilion State Park (Figure 8) where we will observe an outcrop of Neoarchean gabbro just southeast of the Old Ely Road. After a coffee break, we will strap on our hiking boots and make several field trip stops at outcrops along an approximately 4.5 mile traverse through Lake Vermilion State Park. All major stratigraphic units in the park will be observed during this traverse. We will then leave Lake Vermilion State Park, and start our return to Hibbing via Highway 169 South. We will make one final field trip stop on the south side of Highway 169 South just west of Tower to investigate a recent road cut comprising Gafvert Lake Sequence rhyodacite tuffs, lapilli tuffs, and tuff breccias (Figure 7) prior to returning to the Hibbing Park Hotel via Highway 169 South. Mileage for this roadlog starts at the intersection of East Howard Street and Highway 73/169 North. Roadlog (vehicle) mileage will be denoted in bold italic text. Mileage for the traverse through the park will be denoted in *italic* text.

Bus Log

0.0 miles Turn north on Highways 73/169 North and proceed to Virginia, Minnesota.

22.0 miles Turn left on to Highways 53/169 North.

26.6 miles Veer right at the exit for Ely, Minnesota on Highways 1/169 North. You will pass the “Y” store at approximately 44.2 miles, and the intersection for Highway 135 on the west side of Tower at approximately 48.3 miles. Continue on Highway 1/169 North through Tower. At approximately 50.2 miles you will see the intersection of Main Street, Soudan, Minnesota, and a sign for Soudan Underground Mine State Park. Continue on Highway 1/169 past Soudan. At approximately 53.4 miles you encounter the intersection of Highway 1/169 North and the Murray Forestry Road (on the south side of Highway 1/169). Get prepared to turn south off of Highway 1/169 on to a dirt road in approximately 0.6 - 0.7 miles.

54.1 miles Turn right (south) on to the dirt road and immediately park in the open area at the base of the hill. Hike 0.15 miles (approximately 240 meters) up the hill on the dirt road to Field Trip Stop 1.

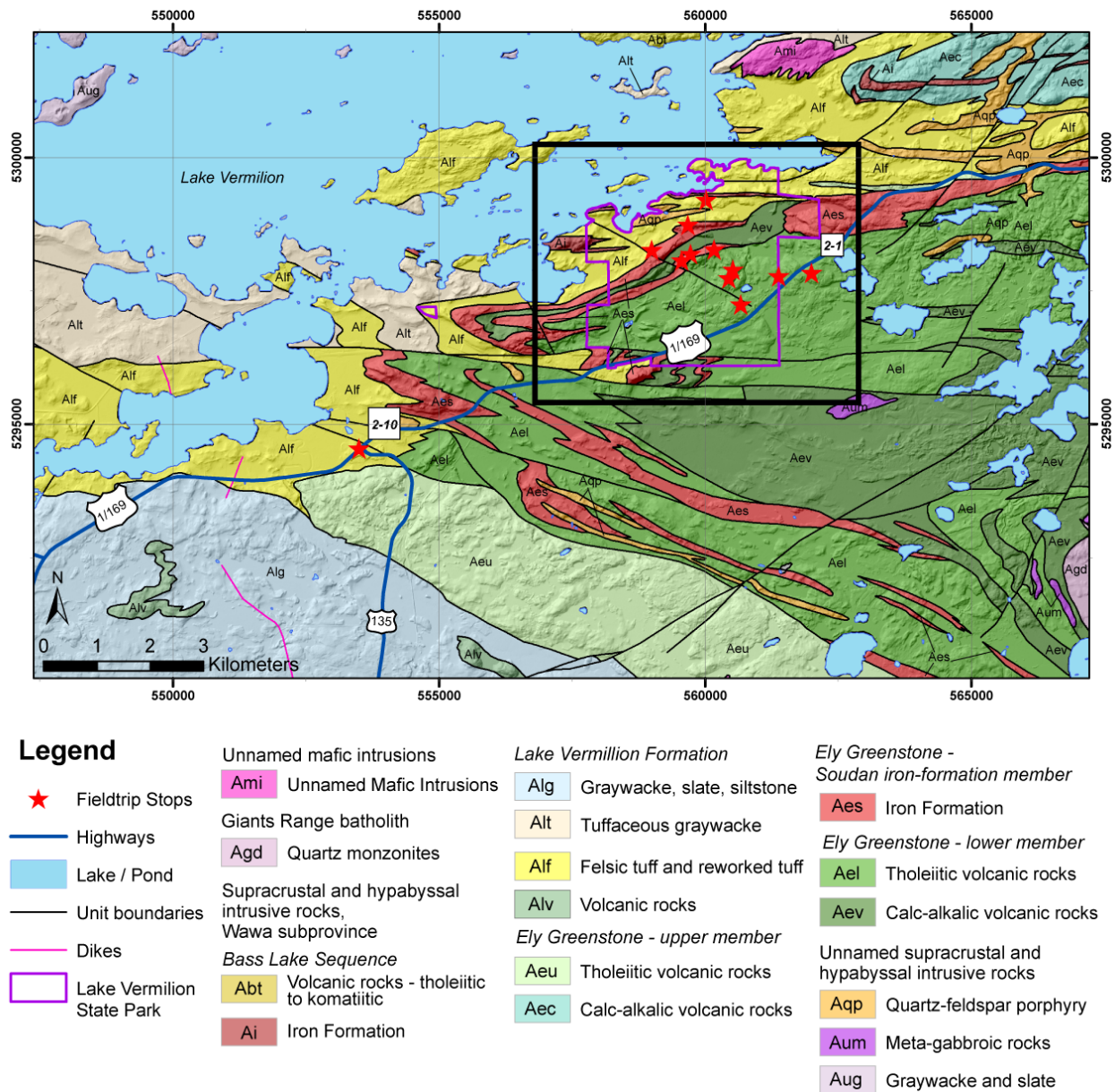


Figure 7. Map illustrating regional geology in the vicinity of Lake Vermilion State Park. Field trip stops 1 and 2 are outside the state park boundaries and are illustrated. The location of Figure 8, a more detailed map illustrating field trip stops in the state park, is illustrated by the bold black box.

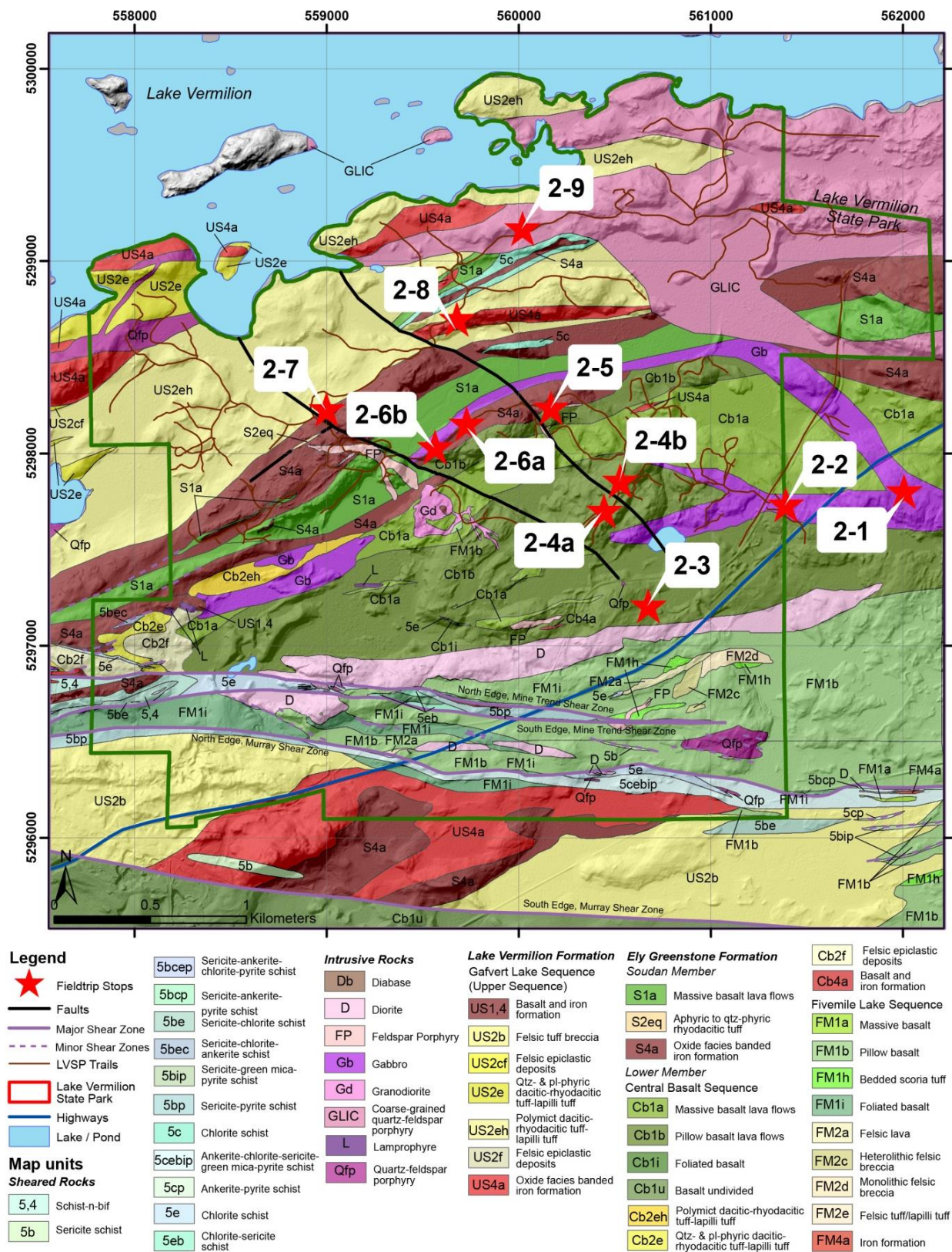
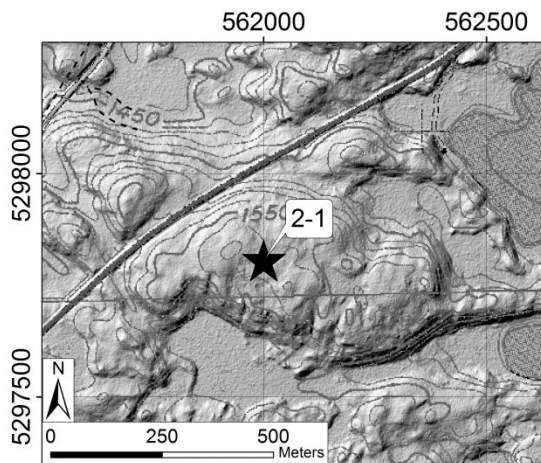


Figure 8. Detailed geologic map of Lake Vermilion State Park (after Peterson and Patelke (2003), Radakovich et al. (2010) and Heim et al. (2011)). Field trip stops within the park are labeled.



Stop 1: Central Basalt Sequence Sheet Flows, Pillow Lavas, and Perlitic Hyaloclastite

Location: T. 62N, R.14W, sec. 19, SE, SW,
Soudan 7.5-minute quadrangle
UTM: 562,000E / 5,297,805N

This classic outcrop has been visited during field trips associated with both the 2004 and 2009 ILSG conferences (Hudak et al., 2004; Peterson et al., 2009). *This is a no-hammer outcrop, as the preservation of the delicate textures here rivals those observed in other classic Neoarchean camps in the Superior Province containing well-preserved volcanic textures such as Noranda, Quebec and Timmins, Ontario.* The description and figure below has been taken from Peterson et al.

(2009).

The Central Basalt sequence (Peterson and Patelke, 2003) comprises a steeply north-dipping (75°-vertical), north-facing sequence of sparsely amygdaloidal pillowed and massive lava flows of basalt andesite to basalt composition that are believed to be correlative with the tholeiitic Armstrong Lake volcanic sequence mapped in the Eagles Nest quadrangle (Jirsa et al., 2001), approximately 11km to the east. Hudak et al. (2007), Jansen et al. (2007), and Hudak et al. (2012) have shown that the lowermost sections of the Central Basalt Sequence are composed of submarine basaltic andesite to basalt lava flows that have rare earth element lithogeochemical patterns similar to mafic rocks in oceanic volcanic arcs. However, locally, submarine basalt lava flows that occur within 50-200m stratigraphically below the contact between the Central Basalt Sequence and the overlying Soudan Member of the Ely Greenstone Formation illustrate MORB-like or back-arc basin-like lithogeochemical patterns. This change in rare earth element characteristics may be interpreted to indicate a change from an oceanic arc to back-arc environment immediately prior to the deposition of the Soudan Member. Relative to massive and pillowed basalt and andesite flows in the Fivemile Lake sequence, Central Basalt sequence lava flows are notably less amygdaloidal, and lack multiple pillow rind structures. In addition, the Central Basalt sequence lacks the thick sequences of scoriaceous basalt-andesite lapilli tuffs that are commonly interstratified with lava flows in the Fivemile Lake sequence. These characteristics of the Central Basalt sequence indicate eruption and deposition in a deeper submarine environment than the stratigraphically older Fivemile Lake sequence, and suggest overall increasing water depth during the temporal development of the Lower Ely. Deepening of the water column could be accommodated by extensional tectonics and normal faulting associated with the development of the proposed back-arc environment.

The outcrop comprises two east-southeast striking massive basalt flows, ranging from at least five to nine meters in thickness, that are separated by a ten meter thick flow unit comprising pillows and pillow lobes (Fig. 9). All three lava flows at this vicinity illustrate tholeiitic, MORB-like lithogeochemistries (Hudak et al., 2007).

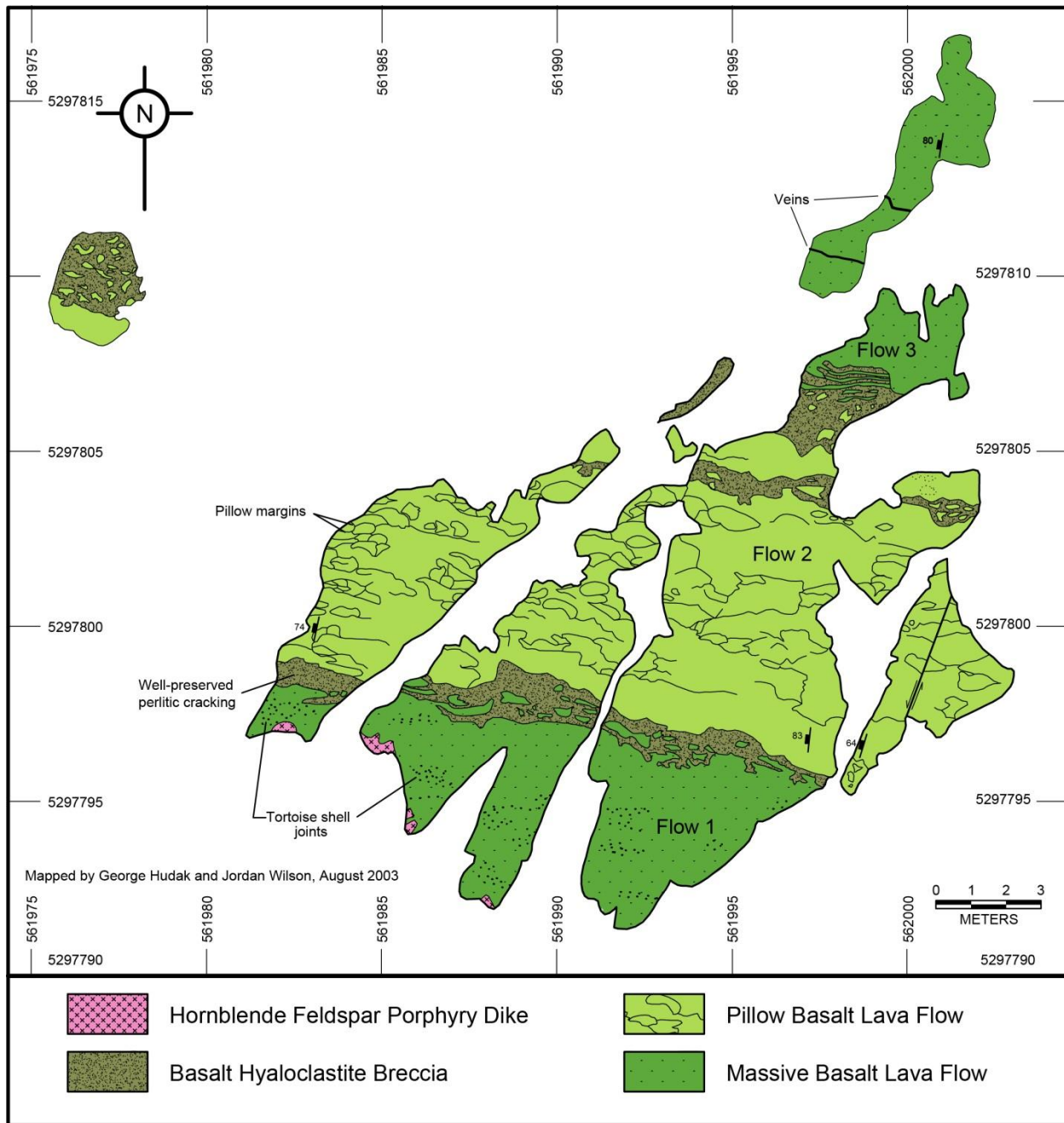


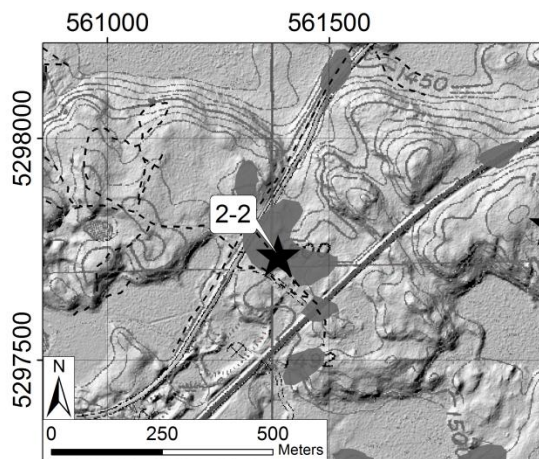
Figure 9. Detailed geological map of sheet flows, pillow lavas, and associated hyaloclastite deposits at field trip Stop 1.

Flow 1, at the southern part of the outcrop, is composed of a pale- to dark green, faintly feldspar-phyric (~10% 0.5-1 mm laths), sparsely amygdaloidal, basalt sheet flow that locally exhibits tortoise-shell jointing formed in response to contraction during cooling. The uppermost 10-40 cm of the coherent part of Flow 1 is generally silicified and epidotized. Petrographic observations indicate that this section of the flow also contains up to 70% <0.1 cm round spherulites. An irregular contact occurs between the coherent basalt flow and an overlying one- to two meter thick unit of dark green, exceptionally well-preserved perlitic in-situ hyaloclastite and associated self-peperite (c.f. Batiza and White, 2000). The hyaloclastite formed from non-explosive fracturing of the basalt glass developed on the flow top due to quenching by

water, whereas the perlite formed following deposition by hydration of volcanic glass. An irregular contact occurs between the hyaloclastite and Flow 2, which is composed of north-facing mattress- to bun-shaped pillow lavas and pillow lobes with numerous “neck and knob” structures. Individual Pillow structures have well developed perlitic hyaloclastite margins that range from 1-4 cm in width. Pillow buds indicate propagation from east to west, suggesting the volcanic vent was located east of this location. The coherent pillows and lobes are overlain by up to 2.5 meters of hyaloclastite breccia that contains 20-40% subround to subangular pale gray green basalt lapilli in a jigsaw puzzle-fit dark green perlitic hyaloclastite matrix. The upper contact of Flow 2 and the overlying basalt sheet flow (Flow 3) is irregular, and is marked by thin (1-8 cm thick), sheet- like basalt fragments that are up to 1.6 meters in length. These fragments locally appear to be isoclinally folded about an east-west-trending fold hinge. Although the genesis of this structure is currently not well understood, it may be due to syneruptive deformation of either thin slabs of hot, basal flow margin crust from the overlying flow, or thin injections of basalt magma into the hyaloclastite from either the underlying pillows or the overlying sheet flow. Flow 3 comprises an at least ten-meter thick pale green-gray, slightly feldspar-phyric, sparsely amygdaloidal sheet flow. Steep, NNE-trending west dipping D3 joints are well developed in this unit, as are lens-shaped pseudo-pillows that are up to 50 cm in diameter.

Return to the bus by walking back down the hill.

- 54.1 miles** *Turn left and follow Highway 1/169 South approximately 0.6-0.7 miles to the west. You will see the Murray Forestry Road on your left (south side of road)*
- 54.7 miles** *Turn right (north) on the dirt road immediately north of the intersection with Highway 1/169 South and the Murray Forestry Road. Proceed north approximately 0.1 miles. Park the vehicle on the Old Ely Road immediately outside the gate to Lake Vermilion State Park. Walk approximately 0.3 miles northeast on the Old Ely Road, then approximately 0.04 miles southwest along the trail to Field Trip Stop 2.*
- 54.8 miles** *Drop off field trip participants on Old Ely Road immediately east of the gate to Lake Vermilion State Park.*
 After dropping off field trip participants, the bus will drive northeast up the Old Ely Road for 1.32 miles.
- 56.1 miles** *Make sharp left turn on dirt road farthest to the south. Continue west on dirt road approximately 0.28 miles to the gate at Lake Vermilion State Park.*
- 56.4 miles** *Enter Lake Vermilion State Park through gate. Proceed 0.48 miles west, where you will see a two-track trail on the south of the road immediately before the well-maintained road turns right sharply to the north. Park the bus in the grass on the south side of the road at the intersection of the well-maintained road and the two-track trail.*
- 56.9 miles** *Park Bus.*



Stop 2: Neoproterozoic Gabbro Sill

Location: T. 62N, R. 14W, sec. 19, SW, SW,
Soudan 7.5-minute quadrangle
UTM: 561,385E / 5,297,735N

Detailed geologic mapping in the Vermilion District (Peterson, 2001; Hudak et al., 2002a, Hudak et al., 2002b; Hudak et al., 2006) has indicated the presence of several gabbro/diabase dikes and sills within both the Fivemile Lake and Central Basalt sequences of the Lower Member of the Ely Greenstone Formation. These intrusive rocks vary from fine-grained diabase with well-developed trachytic textures, to medium- to coarse-grained gabbro and quartz gabbro that locally display well developed ophitic and sub-ophitic textures.

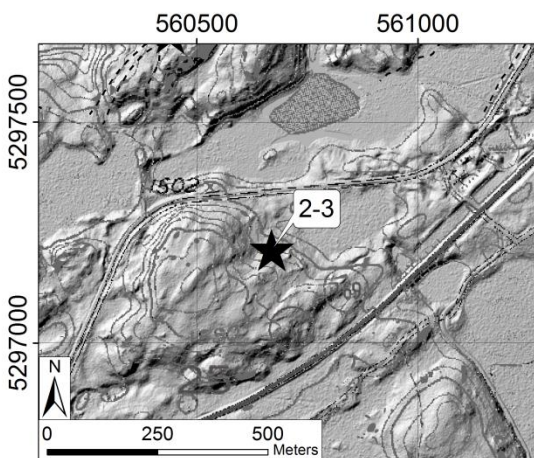
Petrographic observations indicate the presence of relatively unaltered (minor sericite $\pm$ carbonate alteration) euhedral to subhedral plagioclase and subhedral to anhedral actinolite pseudomorphs of original clinopyroxene. At this location, we will observe dark green to dark greenish-black medium- to coarse-grained gabbro that locally displays exceptional sub-ophitic and ophitic textures.

Return to the bus for our morning coffee break and a brief explanation of our traverse through the park

We will now begin our traverse through Lake Vermilion State Park. Make sure to have proper field gear (hat, rain gear, etc), your lunch, and drinks along with you, as we will be out on the trails in the park for nearly the remainder of the trip. Per state park rules, no hammering on the outcrops, or taking of samples, will be allowed while we are on the traverse.

Traverse Log

- 0.0 miles *Leave the bus and go through the gate at the entrance to Lake Vermilion State Park for 0.26 miles. Turn south along the dirt road/trail and proceed approximately 0.9 miles along the trail to Stop 3.*
- 0.35 miles *Stop at outcrop on ridge of hillside.*



Stop 3: Garnet-altered Central Basalt Sequence Pillow Lavas

Location: T. 62N, R. 15W, sec. 25, SW, NE,
Soudan 7.5-minute quadrangle
UTM: 560,670E / 5,297,210N

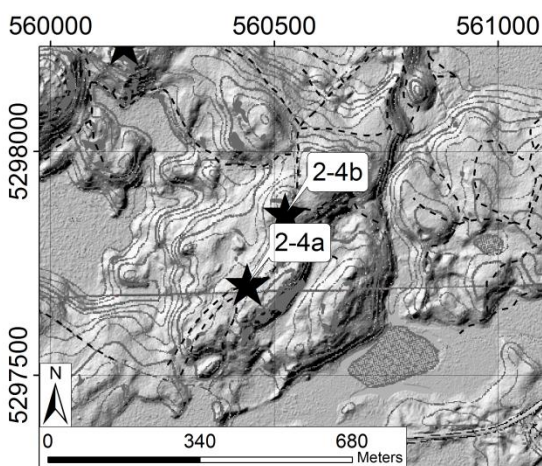
In several locations in the vicinities of Sixmile Lake (Hudak et al., 2006) and Twin Lakes (Moosavi et al., 2007), mafic volcanic and volcanoclastic rocks in the Central Basalt Sequence have been intensely altered to form mineral assemblages comprising quartz, epidote (both pistacite and zoisite/clinozoisite), actinolite, sericite, and/or chlorite (both Mg-rich and Fe-rich compositions). Locally, these altered rocks also

contain minor to moderate abundances of subhedral to euhedral, dark reddish-brown garnets that have been identified as andradite via x-ray diffraction analysis (andradite chemical formula is $\text{Ca}_3\text{Fe}_2\text{Si}_3\text{O}_{12}$, a member of the ugrandite garnet series (Phillips and Griffen, 1981, p. 117)).

At this location, you will observe hydrothermally altered Central Basalt Sequence pillow lavas that contain an abundance of andradite garnet in hydrothermally-altered interpillow hyaloclastite deposits. Recognition of mineral phases that are not consistent with greenschist-facies metamorphism of original basalt composition protoliths is essential to identifying hydrothermal alteration zones. Detailed mapping of such alteration mineral assemblages provides important data regarding the processes and timing of hydrothermal alteration, which in turn, provide essential clues to economic geologists in their quests to find mineralization.

Return to Old Ely Road along the same path used to access Stop 3.

- 0.45 miles *At the intersection of the trail leading to Stop 3 and Old Ely Road, turn left (west) and walk approximately 0.2 miles to the intersection of a two-track road that leads to the north.*
- 0.65 miles *Turn north on the two-track road and proceed north-northeast. Walk approximately 0.27 miles to the first major curve in the road. It is likely to be a bit wet here, so plan to continue walking north-northeast along the edge of the trail. Continue for another approximately 0.27 miles northeast along the two-track road.*
- 1.10 miles *You will encounter a series of five outcrops along the two-track road that will extend for a distance of approximately 0.1 miles. This sequence of outcrops is Stop 4.*

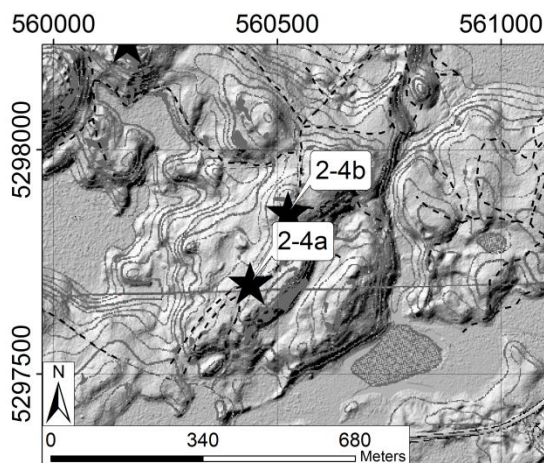


Stop 4a: Central Basalt Sequence Pillow Lavas

Location: T. 62N, R. 15W, sec. 24, SE, SW,
Soudan 7.5-minute quadrangle
UTM: 560,440E / 5,297,700N

At this location, a series of six small outcrops occurs as one traverses approximately 125 meters up a small hill from southwest to northeast. Here you will observe well-preserved, commonly muffin-shaped, sparsely vesicular variably altered Central Basalt Sequence pillow lavas. In several locations, dark reddish-brown sulfide burn can be observed where sulfide minerals (pyrite, locally minor chalcopyrite) have been oxidized. Near the central part of the outcrop exposure (outcrop number four as one proceeds from south to north), locally strong silicification and actinolite alteration may be observed.

- 1.2 miles *From the northernmost outcrop associated with Stop 4, proceed north approximately 0.1 miles.*
- 1.3 miles *You will see a series of small outcrops that extend for approximately 0.1 miles along the east side of the two-track road.*

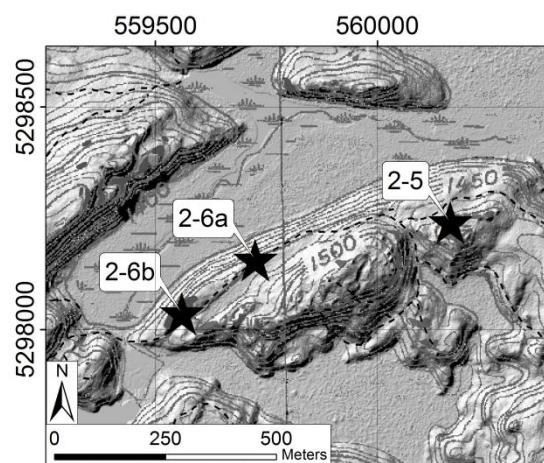


Stop 4b: Hydrothermally Altered Central Basalt Sequence Pillow Lava

Location: T. 62N, R. 15W, sec. 24, SE, SW,
Soudan 7.5-minute quadrangle
UTM: 560,525E / 5,297,860N

This stop comprises a series of east-northeast striking, north-topping, locally silicified and actinolite-altered Central Basalt Sequence basalt to basaltic-andesite pillow lavas that extend for approximately 0.1 miles as one traverses toward the north. Note the brown stains within both the interpillow hyaloclastite deposits and the cores of the pillow lavas that result from weathering of minor amounts of pyrite in the rock.

- 1.4 miles *From the northernmost outcrop exposure, continue walking northwest along the two-track road.*
- 1.75 miles *Gather on the two-track road, and follow the field trip leader approximately 0.02 miles (~30 meters) through the bush.*
- 1.77 miles *Gather on the north slope of the north-south trending outcrop. This is Stop 5.*



Stop 5: Contact Between Soudan Member Banded Iron Formation and Soudan Basalt

Location: T. 62N, R. 15W, sec. 24, NW, SW,
Soudan 7.5-minute quadrangle
UTM: 560,165E / 5,298,240N

Mapping by Jirsa et al. (2001), Peterson and Patelke (2003), Radakovich et al. (2010) and Heim et al. (2011) has shown that the Soudan Member of the Ely Greenstone Formation is composed of Algoma-type oxide-facies banded iron formation horizons interbedded with massive and pillowed basalt lava flows, aphyric- to quartz-phyric rhyolite tuffs, and locally, polymict quartz- and plagioclase-phyric dacitic to rhyodacitic lapilli tuff deposits. To the south and

east, as well as within the 2700 drift of the Soudan Underground Mine, shearing of the interbedded iron formation and basalt horizons has resulted in a rock comprising chaotically intermingled chlorite schist and banded iron formation which Peterson and Patelke (2003) most appropriately termed “Schist ‘n’ BIF”.

The Algoma-type iron formations of the Soudan Member comprise laminated- to medium-bedded iron formation containing dark gray to black magnetite-rich bands, bluish-gray to red hematite-rich bands, red jasper bands, and light gray to black chert bands. Planar bedding is most common, with tight, commonly chaotic folds present that have been, in part, interpreted to be the result of soft sediment deformation. The unit is typically strongly magnetic, but is locally moderately to weakly magnetic where dominated by hematite-rich horizons or chert horizons.

The Soudan Basalts comprise medium-green to dark green, aphyric to sparsely plagioclase- ± pyroxene-phyric massive to amygdaloidal basalt. Typically, the recrystallized matrix (now chlorite-epidote-actinolite) contains up to 3% <1mm subhedral to euhedral tabular plagioclase phenocrysts and locally, 5-

7% <1mm dark green actinolite pseudomorphs of pyroxene phenocrysts. Locally, amygdaloidal basalt flows contain 5-7% oval to round, light gray to white, quartz- ± epidote- ± chlorite-filled amygdules ranging from <1-4mm in diameter. Locally, brownish-tan colored ankerite alteration and dark green chlorite alteration are present.

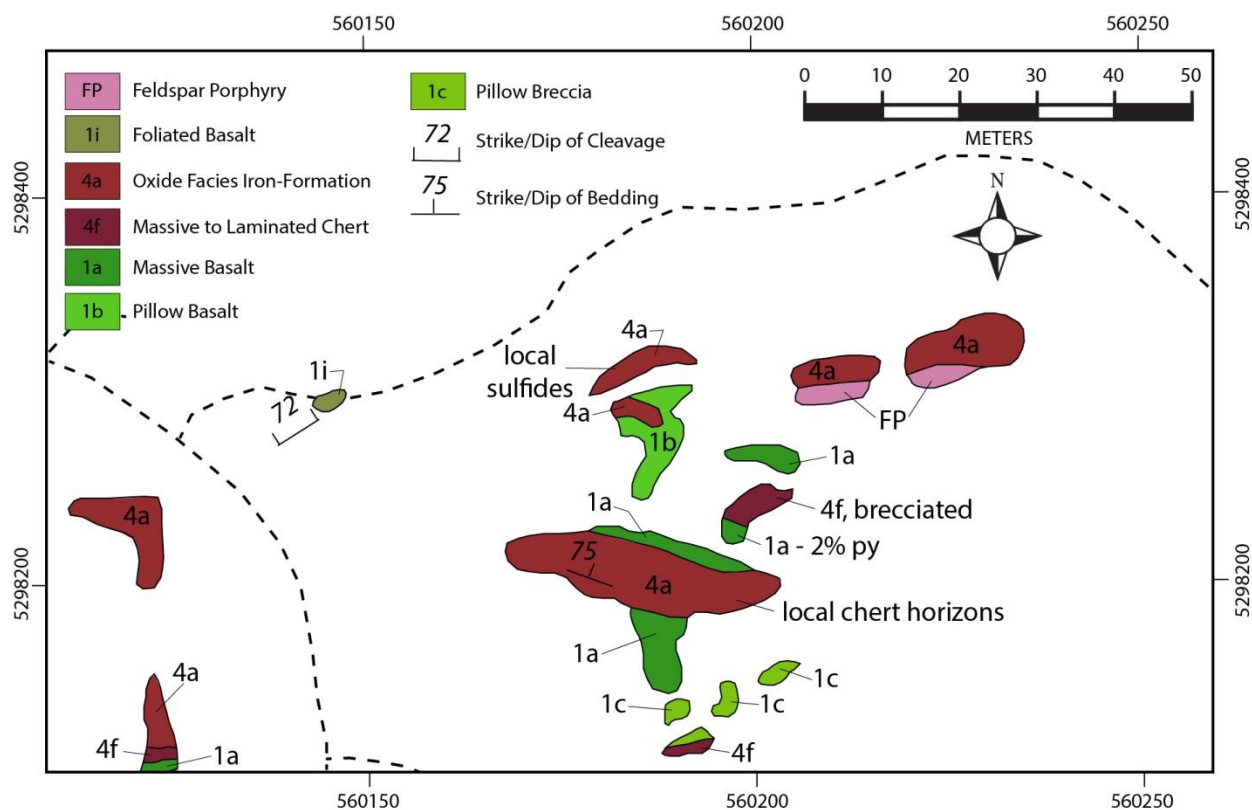
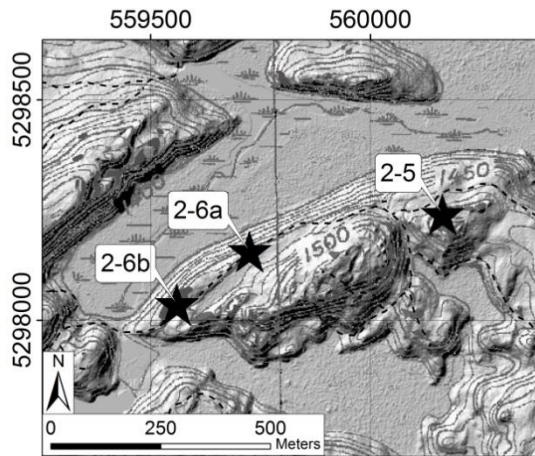


Figure 10. Detailed (1:5000 scale) map illustrating the complex contact relationships between Soudan Member oxide facies iron formation and basalt units in the vicinity of Stop 5.

Figure 10 is a reproduction of a detailed field map (originally mapped at 1:5000 scale) in the general vicinity of Stop 5. The map illustrates the complex contact relationships between Soudan Member banded iron formation and basalt units at this location. Here we will see the nature of the contact between one of the banded iron formation units and an adjacent massive basalt lava flow.

- 1.77 miles *Walk north-northwest through the bush back to the two-track road.*
- 1.79 Miles *Proceed west-southwest along the two-track road. At the fork in the road, proceed to the northwest along the two-track road. Continue walking along the two-track road for approximately 0.35 miles.*
- 2.14 miles *At this point you will encounter a series of outcrops within, and along the south and north edges, of the two-track road. This will be Field Trip Stop 6a.*



Stop 6a: Folded Soudan Iron-Formation Member Banded Iron Formation

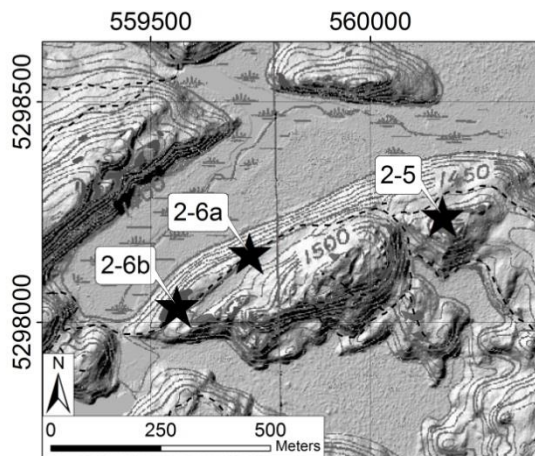
Location: T. 62N, R. 15W, sec. 23, NE, SE,
Soudan 7.5-minute quadrangle
UTM: 559,725E / 5,298,155N

As is well exemplified at the “classic” outcrop of Soudan Member banded iron formation located in the NE ¼ NE ¼ Sec. 27, T.62N, R. 15W (see stop 7-10 in Peterson et al., 2009), the Soudan Member banded iron formation commonly displays multiple generations of tight folds which can result in complex interference patterns (Figure 11). At this location, and at several other small outcrops along the north side of the two-track trail, we can observe highly folded, moderately- to strongly magnetic, magnetite-rich Soudan Member oxide facies iron formation.

- 2.14 miles *Continue walking southwestward along two-track road for approximately 0.02 miles.*
- 2.16 miles *You will see several outcrops between 0.01 and 0.02 miles into the bush on the northwest side of the two-track road. Proceed to these outcrops.*
- 2.17 miles *This will be Stop 6b.*



Figure 11.
Tightly
folded
Soudan
Member
Algoma-type
banded iron
formation



Stop 6b: Contact between Folded Soudan Iron-Formation Member Banded Iron Formation and Diabase/Gabbro Intrusion

Location: T. 62N, R. 15W, sec. 23, SE, SE,
Soudan 7.5-minute quadrangle
UTM: 559,560E / 5,298,035N

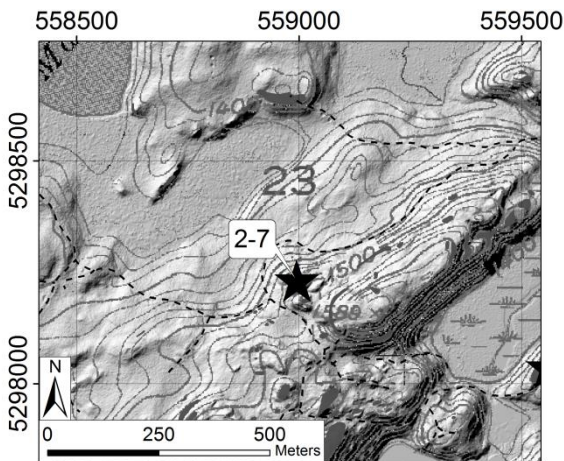
Note: Be extremely careful on this outcrop, especially if it is wet. The glacially polished surface combined with wet moss makes for very slippery conditions.

This outcrop is once again composed primarily of laminated to thinly-bedded Soudan Member oxide-facies banded iron formation. On the far western side of the outcrop, as well as in several small outcrops to the

northeast, we can observe a massive, fine- to medium-grained, dark green to grayish-green rock which has been interpreted to represent a diabase sill. This sill appears to intrude the contact between Soudan Member oxide-facies iron formation (to the south) and Soudan Member massive basalt lava flows (to the north). Based on the outcrop distribution in the park, this unit appears to get coarser grained to the east, where it represents dark green to blackish-green medium-grained gabbro.

- 2.17 miles *Return to the two-track road.*
- 2.19 miles *Walk southwest, then west, down the hill along the two-track road.*
- 2.30 miles *Cross bridge over small creek between unnamed pond (to the south) and creek/swamp (to the north). **Be extremely careful crossing this bridge as it may be wet and slippery!** Continue west, then southwest along the two-track road and begin to climb a moderately steep hill.*
- 2.49 miles *Continue walking up the hill. The outcrop ridge on both sides of the two-track road comprise poorly exposed interbedded Soudan Member banded iron formation and Soudan Member basalt lava flows exhibiting both sheet flow facies and associated flow breccia facies. Continue walking another 0.11 miles up to the top of the hill.*
- 2.60 miles *We are now at the top of the hill. We will reassemble here before moving on to the remainder of the outcrops along our traverse. To the west, the prominent hill and low lying outcrops along the trail comprise magnetite-rich Soudan Member banded iron formation.*

After reassembling the group, we will proceed north, then northeast, along the two-track trail for 0.09 miles.
- 2.69 miles *Take the left fork and proceed to the northwest along the two-track road.*
- 2.80 miles *We will once again reassemble the group at this location. Once reassembled, we will walk north-northeast approximately 0.01 miles up a hill through the bush to Field Trip Stop 2.7.*
- 2.81 miles *Field Trip Stop 2.7*



Stop 7: “Contact” Between Soudan Member Banded Iron Formation and Rhyodacite Polymict Lapilli Tuff/Tuff Breccia of the Gafvert Lake Sequence

Location: T. 62N, R. 15W, sec. 23, NW, SE,
Soudan 7.5-minute quadrangle
UTM: 558,995E / 5,298,230N

Here we will see one of the few places where the nature of the contact between the Soudan Iron-Formation Member oxide facies iron-formation and the overlying dacitic to rhyodacitic volcanoclastic rocks associated with the informally named Gafvert Lake Sequence (which is part of the Lake Vermilion Formation) can be observed. Based on regional

mapping, Sims and Southwick (1980), Southwick (1993), and Southwick et al. (1998) have indicated that the contact between the underlying Soudan Iron-Formation Member of the Ely Greenstone Formation and the overlying Lake Vermilion Formation is locally an unconformity.

Geochronological work in the Vermilion District (Peterson et al., 2001; Lodge et al., 2013), combined with detailed field mapping in the limited number of locations where the contact between the Soudan Iron-Formation Member and the Lake Vermilion Formation occurs, bears out this interpretation. Peterson et al. (2001) obtained a U-Pb zircon age of 2722 ± 0.9 Ma from a quartz-phyric rhyolite dome within the Fivemile Lake Sequence at the Fivemile Lake prospect, located approximately 850 meters stratigraphically below the base of the overlying Soudan Member Iron-Formation unit. Regionally extensive detailed mapping in the stratigraphic units that occur between the Fivemile Lake Sequence rhyolite dome and the base of the Soudan Member Iron-Formation has been completed by a number of researchers (Peterson and Jirsa, 1999; Peterson, 2001; Hudak et al., 2002a; Hudak et al., 2002b; Peterson and Patelke, 2003; Hoffman, 2007; Radakovich et al., 2010; Heim et al., 2011). Based on this detailed mapping, there are no indications of any unconformities within the Fivemile Lake Sequence or the Central Basalt Sequence that comprise the footwall to the Soudan Iron-Formation Member. As well, unconformities at the contacts between the Fivemile Lake Sequence and the Central Basalt Sequence, and the Central Basalt Sequence and the overlying Soudan Iron-Formation do not appear to be present. Hudak et al. (2007; 2012) have noted that the contact between the Central Basalt Sequence and the overlying Soudan Iron-Formation Member is transitional over several hundred meters, with the presence of iron-formation horizons near the top of the Central Basalt Sequence increasing in abundance, and the abundance of basalt lava flows and associated volcanoclastic rocks decreasing in abundance, as one approaches the base of the Soudan Iron-Formation Member. Therefore, it appears that the Fivemile Lake Sequence is stratigraphically overlain by the Central Basalt Sequence, which in turn is stratigraphically overlain by the Soudan Iron-Formation Member. Furthermore, there appears to be a major period of volcanism and associated hydrothermal activity between 2722 and 2718 Ma in the western part of the Wawa Abitibi Terrane in Ontario that produced both the volcanic rocks and volcanogenic massive sulfide orebodies that occur at the Winston Lake and Geco deposits (Lodge et al., 2013). Based on both the geochronological work and detailed mapping in the Vermilion District, as well as the regional volcanic and hydrothermal events in the western Wawa-Abitibi belt, we currently believe that the Soudan Iron-Formation Member was deposited between 2722 and 2718 Ma. Further geochronological studies within the stratigraphic package that comprises the Soudan Iron-Formation Member of the Ely Greenstone Formation will need to be completed in order to verify our current interpretation.

Based on field relationships recognized by Radakovich et al. (2010), Lodge et al. (2013) collected a sample of the basal part of the Gafvert Lake polymict dacite- to rhyodacite lapilli-tuff / tuff-breccia deposits that occur at this outcrop in order to determine the age of volcanism of the Gafvert Lake

Sequence relative to the ages of the Lower and Soudan Iron-Formation members of the Ely Greenstone Formation. Zircons from the sample of polymict rhyodacite tuff-breccia from this outcrop approximately 2m north of the contact with the Soudan Iron-Formation Member produced a high precision U-Pb age of 2689.7 ± 0.8 Ma using thermal ionization mass spectrometry (Lodge et al., 2013). Given that the basal Gafvert Lake Sequence deposits contain angular intraclasts of chert and banded iron formation, and that there appears to be no intense structural fabric in either the Soudan Iron-Formation Member or the Gafvert Lake volcaniclastic rocks, Lodge et al. (2013) interpreted the contact here to represent a disconformity, a type of unconformity characterized by strata that are essentially parallel on either side of the erosional or non-depositional surface.

Several outcrops occur at this location, but at 1:5000 scale mapping they have been combined into a single east-northeast trending outcrop (Figure 12). The majority of the outcrop, which extends east up the hill, is composed of laminated to medium bedded Soudan Iron-Formation Member. Alternating magnetite-rich horizons, chert horizons, and jasper horizons display planar bedding and are locally folded. Moving toward the northwest part of the outcrop, we observe a small break in the outcrop exposure. This break occurs directly above the contact between the Soudan Member Iron-Formation and the Gafvert Lake volcaniclastic rocks. In this area, note the lack of deformation in both lithological units. The lack of structural deformation at this contact, as well as geochronological data obtained from the Gafvert Lake volcaniclastic rocks near this contact (Lodge et al., 2013), supports the interpretation of a disconformity.

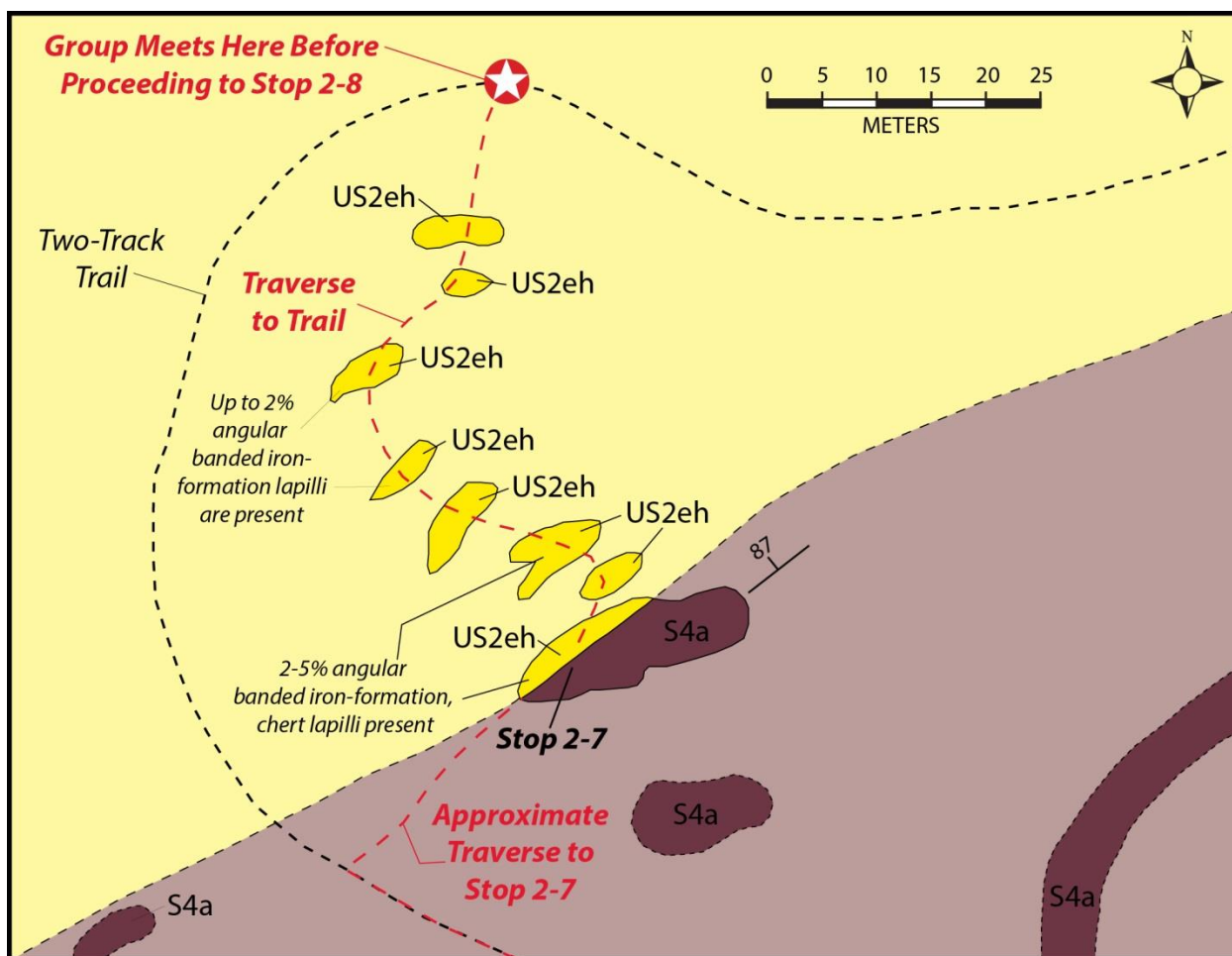


Figure 12. Detailed (1:5000 scale) map illustrating the disconformable contact between the Soudan Member Algoma-type banded iron-formation (unit S4a) and the Gafvert Lake Sequence quartz- and plagioclase-phyric polymict dacite-rhyodacite tuff-breccia / lapilli tuff deposits (unit US2eh). We will

start our investigation where Stop 7 is indicated, and traverse along the path indicated by the red dashed line over a series of outcrops. We will assemble on the two-track trail where indicated by the star symbol before proceeding to Stop 8.

Moving to the northwest, we observe the basal several meters of the Gafvert Lake Succession volcanoclastic rocks. Here, the rock is composed of a very thickly bedded quartz- and plagioclase-phyric polymict dacite-rhyodacite tuff-breccia / lapilli tuff. The rock is characterized by up to 5% 1-3mm diameter subhedral to euhedral gray to blue-gray quartz phenocrysts and locally, 5-10% subhedral to euhedral light gray to tan tabular plagioclase phenocrysts set in a fine-grained quartzo-felspathic matrix that is locally sericite altered. Accidental fragments comprising lapilli-sized light gray to grayish black angular to subangular chert, gray to dark gray subangular to angular banded iron formation (Figure 13), and rare angular to subangular reddish brown jasper fragments are present. As well, juvenile fragments comprising lapilli- to locally block-sized pumice are present. Lapilli- to block-sized accessory fragments of quartz- and plagioclase-phyric coherent dacite and rhyodacite are also present, in abundances up to 5%.

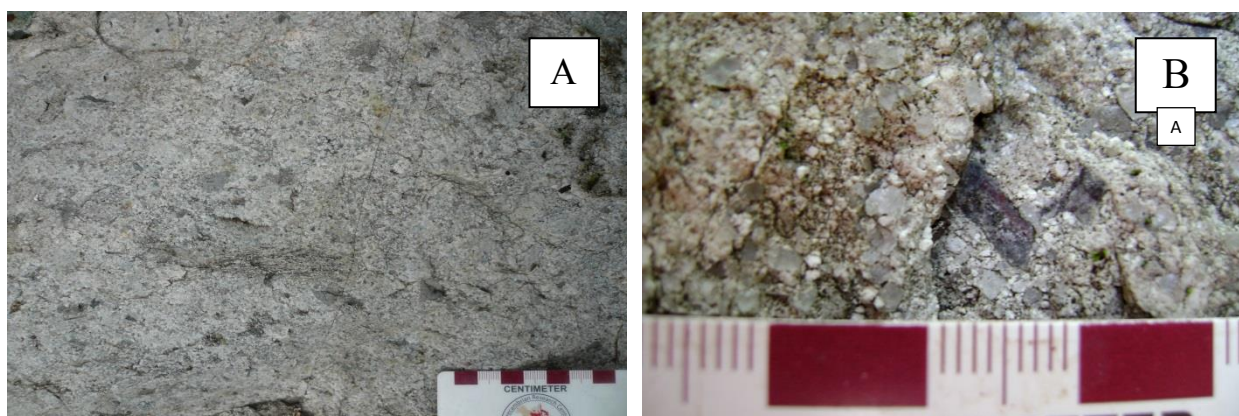
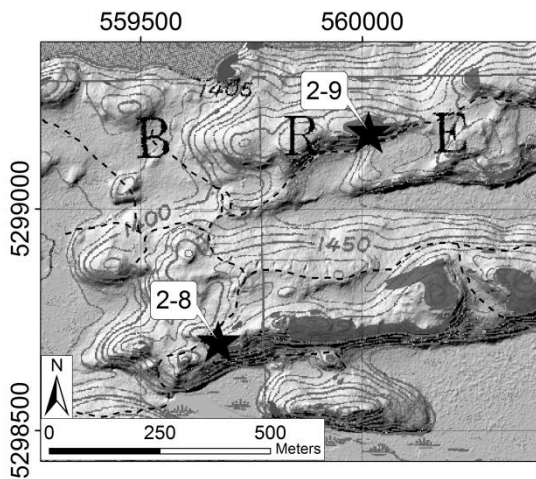


Figure 13. Gafvert Lake Sequence quartz- and plagioclase-phyric polymict dacite-rhyodacite tuff-breccia / lapilli tuff from the Gafvert Lake Sequence. A. Typical appearance of very thickly bedded quartz- and plagioclase-phyric polymict dacite-rhyodacite lapilli tuff. B. Close-up of unit illustrating tannish-white subhedral to euhedral tabular plagioclase phenocrysts, gray to gray-blue anhedral quartz phenocrysts, and 1cm diameter angular accidental fragment composed of jasper-rich banded iron formation.

- | | |
|------------|--|
| 2.81 miles | <i>We will proceed north down the north-sloping hillside for about 0.09 miles over a series of outcrops comprising Gafvert Lake Succession rhyodacite tuffs, lapilli tuffs, and tuff breccias. Observe the subtle changes in crystal content and fragment compositions and abundances while moving down the hill toward the two-track trail.</i> |
| 2.90 miles | <i>We will reassemble the group on the two-track trail. We will proceed to walk east-northeast along the two-track road for approximately 0.42 miles.</i> |
| 3.32 miles | <i>Cross bridge – there will be a large unnamed pond to the east. Continue 0.09 miles up hill to intersection and wait where the two-track road makes a turn from north-trending to east-trending.</i> |
| 3.41 miles | <i>This is where we parked our truck each day during our 2010 capstone mapping project for the PRC Field Camp. We will now walk northeast along the road for 0.07 miles.</i> |
| 3.48 miles | <i>Follow the field trip leader approximately 0.01-0.02 miles southeast into the bush to Stop 8.</i> |
| 3.50 miles | <i>Stop 8.</i> |



Stop 8: Gafvert Lake Sequence Tuffs and Lapilli Tuffs

Location: T. 62N, R. 15W, sec. 23, SE, NE,

Soudan 7.5-minute quadrangle

UTM: 559,675E / 5,298,700N

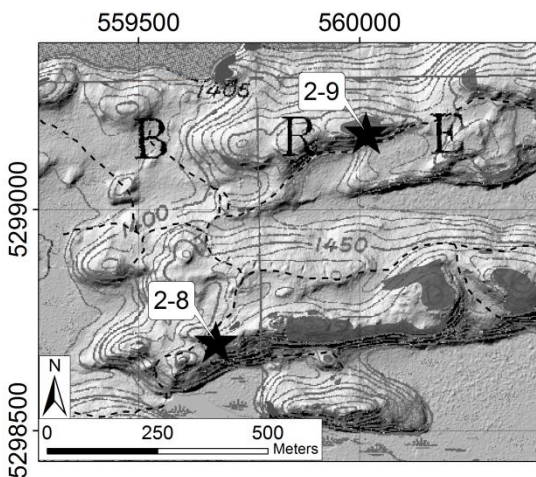
We will stop here to observe several small outcrops of the Gafvert Lake Sequence tuffs and lapilli tuffs. These deposits comprise very thickly bedded, light gray, quartz- and plagioclase-phyric dacitic to rhyodacitic tuffs and lapilli tuffs. The light gray recrystallized matrix generally contains 10-15% <1-2mm subhedral to euhedral tabular plagioclase phenocrysts which locally appear to be broken, as well as 3-8% <1-2mm pale gray anhedral, locally broken, anhedral to subhedral quartz phenocrysts. Various

types of lapilli may be observed, including: 1) 10-20% 1-3cm diameter quartz- and plagioclase-phyric coherent dacite to rhyodacite lapilli; 2) 5-7% ≤ 3 cm diameter pale gray green, lens-shaped, locally quartz- and plagioclase-phyric pumice lapilli; 3) ≤ 1 mm dark gray to light gray angular chert lapilli ranging from 0.5-3cm in diameter; and 4) 1-3% 0.5-5cm dark gray to black magnetite-rich banded iron formation lapilli.

3.50 miles *Traverse approximately 0.01 miles north through the bush back on to the two-track road.*

3.51 miles *Proceed northeast, then northwest, then northeast along the two-track trail for approximately 0.48 miles. We will assemble the group at this location before the group follows the leader on a 0.01mile traverse north-northwest into the bush to Stop 9.*

4.00 miles *Stop 9.*



Stop 9: Quartz $\pm$ plagioclase-phyric Rhyodacite Sill (informally named the Gafvert Lake Intrusive Complex)

Location: T. 62N, R. 15W, sec. 24, NW, NW,

Soudan 7.5-minute quadrangle

UTM: 560,015E / 5,299,175N

At this location we will observe a spectacular light gray, massive, quartz- $\pm$ plagioclase-phyric coherent rhyodacite which, based on regional mapping (Peterson and Jirsa, 1999; Peterson, 2001; Hudak et al., 2002b; Heim et al., 2011) comprises a sill-dike complex that extends from the northern extents of Lake Vermilion State Park over 20km eastward to Mitchell Lake. This intrusion is most prevalent in the vicinity of Gafvert Lake, where it

comprises several sills and dikes that intrude into the thickest section of Gafvert Lake Sequence volcanoclastic rocks. Based on the distribution of sills and dikes, coherent-facies Gafvert Lake Sequence deposits, and an abundance of coarse polymict breccias in this region, Peterson (2001) has interpreted this area to be the remnants of a stratovolcano that produced the Gafvert Lake Sequence dacitic to rhyodacitic volcanoclastic rocks. For this reason, this unique quartz-feldspar porphyry intrusion has been informally named the Gafvert Lake Intrusive Complex (GLIC). Lithogeochemical work recently completed at the University of Wisconsin Eau Claire by Geoff Pignotta and Kelly Schwierske indicates that the GLIC and Gafvert Lake volcanoclastic rocks have very similar major, trace and rare earth element characteristics suggesting that they may be genetically related. However, geochronological studies will need to be

performed to determine unambiguously if the GLIC and Gafvert Lake volcanoclastic rocks are genetically related.

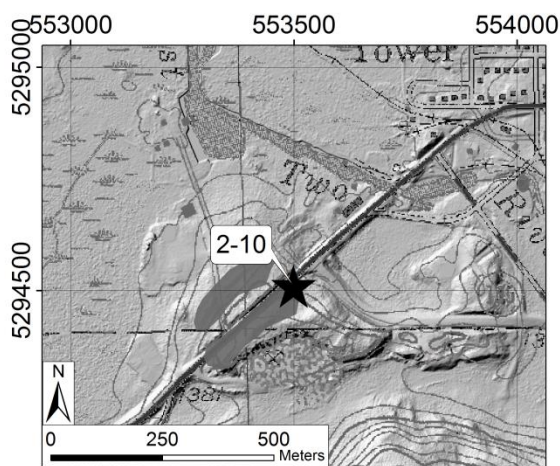
The GLIC comprises light gray, massive, quartz $\pm$ plagioclase-phyric coherent rhyodacite. The light gray aphanitic groundmass contains 3-7% gray to light blue subhedral rounded to euhedral square quartz phenocrysts that range from 3-10mm in diameter, and 10% pale gray to tan, subhedral to euhedral tabular plagioclase phenocrysts ranging from 1-4mm in length. A variety of xenoliths may be found in this intrusion, including: 1) brown mudstone lapilli; 2) green to gray-green massive and/or amygdaloidal basalt lapilli; and 3) light gray aphyric coherent rhyodacite lapilli. In the field, the presence of large 5mm-10mm diameter gray to blue gray quartz phenocrysts distinguishes the GLIC from other quartz-feldspar-porphyry intrusions in the Vermilion District.

- 4.00 miles *Traverse south through the bush 0.01 miles back to the two-track trail.*
- 4.01 miles *Once you return to the two-track trail, walk 0.58 miles east-northeast. At the end of the two-track trail you will intersect a well-maintained dirt road. The bus will be parked at this location.*
- 4.59 miles *Obtain refreshments and get on the bus.*

End of traverse log

Bus Log (Continued)

- 56.9 miles** *Pick up field trip participants after their traverse through the park. Drive east on well-maintained dirt road 0.48 miles back to the intersection with Old Ely Road.*
- 57.7 miles** *Turn right (southwest) and follow the Old Ely Road approximately 1.3 miles to the intersection with the dirt road immediately before the gate to Lake Vermilion State Park.*
- 59.0 miles** *Turn south on dirt road and return to Highway 1/169.*
- 59.1 miles** *Turn west on Highway 1/169 South. Proceed approximately 5.2 miles past Soudan and through Tower to the intersection between Highway 1/169 and Highway 135.*
- 64.3 miles** *Pull bus off on to shoulder of Highway 1/169 South just west of the intersection of Highways 1/169 and Highway 135. The outcrop on the south side of the road is Field Trip Stop 10.*



Stop 10: Recently Exposed Outcrop of Gafvert Lake Sequence Lapilli-tuffs and Tuff-Breccias

Location: T. 62N, R. 15W, sec. 32, SW, SW,
Tower 7.5-minute quadrangle
UTM: 553,500E / 5,294,510N

NOTE: Take extreme care when crossing the highway at this location!

The final stop on our field trip is to a recently exposed (~2012) outcrop comprising polymict Gafvert Lake Sequence tuff, lapilli-tuff and tuff-breccia deposits. Although never mapped in detail by any of the co-authors in this guidebook, this exposure appears to contain several individual volcanoclastic units that may

be distinguished by the size and abundance of the phenocrysts and fragments present. The pale gray to gray aphanitic matrix contains variable percentages and sizes of quartz and plagioclase phenocrysts. Locally, abundant (up to 10%) <1mm euhedral pyrite cubes are disseminated in the matrix. Fragment

composition is also variable, with gray chert, light gray quartz- and plagioclase-phyric coherent rhyodacite, light gray to tan pumice, and rare massive pyrrhotite fragments present. The rock also possesses a moderately- to well-defined schistosity with lineations that plunge moderately to steeply to the northeast (Sims, 1973).

64.3 miles *Field trip participants should return to bus. Take extreme care when crossing Highway 1/169. Follow Highway 1/169 south and retrace route to Vermilion District back to the Hibbing Park Hotel in Hibbing, Minnesota.*

112.5 miles *End of field trip at parking lot in Hibbing Park Hotel.*

Acknowledgements

Characterizing and evaluating the detailed geology of Lake Vermilion State Park involved a team effort between Minnesota Department of Natural Resources (MDNR) personnel, the Minnesota Geological Survey, NRRI geologists, and students and faculty from the Precambrian Research Center Field Camp as well as the University of Wisconsin Eau Claire. The lead author would like to thank Jim Essig (Manager, Soudan Underground Mine State Park and Lake Vermilion State Park) and James Pointer (Interpretive Supervisor, Soudan Underground Mine State Park and Lake Vermilion State Park) from the MDNR for their support, assistance, and guidance while planning and conducting detailed geological mapping by PRC students and faculty in Lake Vermilion State Park in 2010 and 2011. Also, Minnesota Geological Survey geologists Amy Radakovich, Mark Jirsa, and Terry Boerboom are thanked for their assistance (and patience!) during the development of this field trip guide. As well, Dean Peterson, the late Richard Patelke, Mark Severson, John Heine, Peter Jongewaard, Steve Hovis and Adam Hoffman are thanked for their excellent mapping in the southern part of what was to become Lake Vermilion State Park. This work by former and current NRRI colleagues became the foundation upon which new mapping in the park was based. Additionally, Geoff Pignotta, Kelly Schwierske, and the Department of Geology at the University of Wisconsin, Eau Claire are thanked for intellectual efforts and financial support to further evaluate the geology and geochemistry of Lake Vermilion State Park. Finally, PRC students Chris Heim, Rob Kilduff, Chris Mahr, Charlie Parent, Molly Partidge, Rita Pierce, Amy Radakovich, Christine Rahtz, Andrew Ritts, Heather Scott and Andrew Vial are thanked for their outstanding field mapping, compiling, computer map generation, and companionship during the four weeks in 2010 and 2011 that it took to produce the recent detailed geologic maps in Lake Vermilion State Park. Without these exceptional students, our knowledge of the fascinating geology of Lake Vermilion State Park would not be nearly what it is today.

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FIELD TRIP 3

Wednesday, May 14, 2014

WESTERN MESABI RANGE MINING OPERATIONS

LEADERS:

*Douglas Halverson (Cliffs Natural Resources—Duluth)
Daniel Cervin (Cliffs Natural Resources—Hibbing Taconite),
William Everett and Kevin Kangas (Essar Steel); and
Joseph Nielsen (Magnetation).*

INTRODUCTION

This field trip will visit three distinct iron mining operations along the western part of the Mesabi Iron Range (Fig. 1). The morning will be spent at the Hibbing Taconite operation managed by Cliffs Natural Resources. There, participants will get a pit tour of the mining and reclamation operations, followed by a tour of the processing facility. The afternoon will include a tour through the new construction of Essar Steel's taconite processing facility near Nashwauk (near the old Butler Taconite site), and may view some core that intersects strata including what is inferred to be the 1850 Ma Sudbury Impact Layer. This will be followed by a tour of Magnetation's two-year old, 1.2 million ton per-year facility located near Bovey, where iron concentrate is being extracted from the tailings of historic mining of natural (hematite) ores.

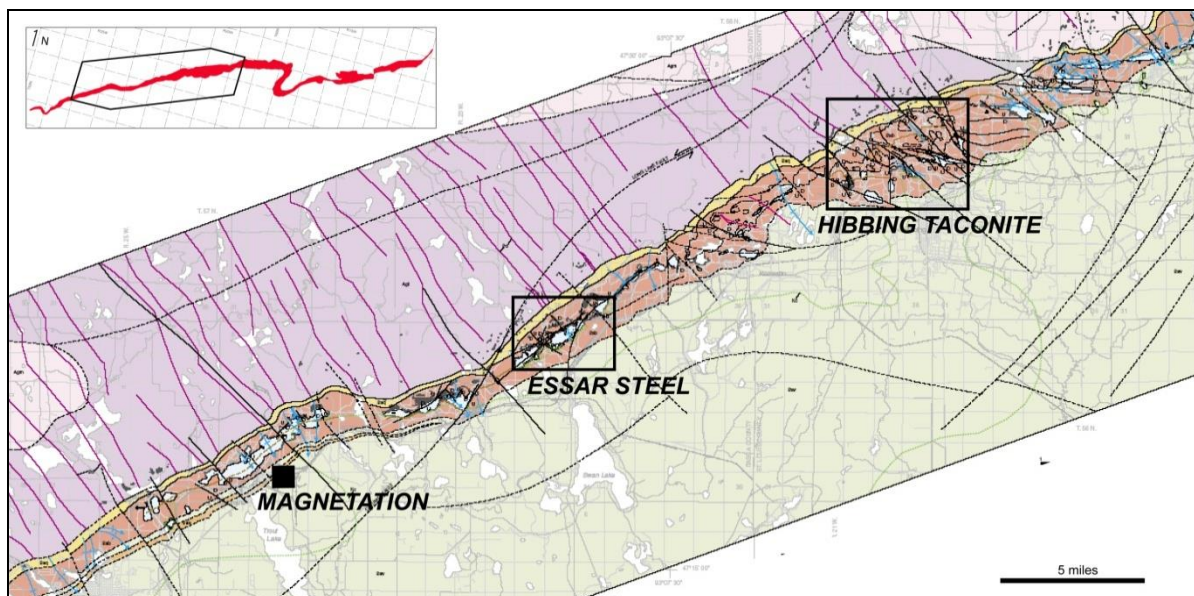


Figure 1. Bedrock geologic map of the western Mesabi Iron Range showing the 3 operations that will be visited during this trip. Map is clipped from MGS Miscellaneous Map M-163 (Jirsa and others, 2005; published scale=1:100,000). Reddish unit represents subsurface extent of Biwabik Iron Formation.

HIBBING TACONITE PIT AND PLANT

Hibbing Taconite Company, managed by Cliffs Natural Resources

Doug Halverson and Dan Cervin (Cliffs Natural Resources)

Field Trip participants will get a pit tour of the mining and reclamation operations. There may be an opportunity to view a blast, dependent upon blasting schedule and blast location. Hibbing Taconite Company (HTC) is managed by Cliffs Natural Resources, an international mining and natural resources company. Cliffs is the largest producer of iron ore pellets in North America, a major supplier of direct-shipping lump and fines iron ore out of Australia and a significant producer of metallurgical coal.

The annual production capacity of Hibbing Taconite Company is 8.0 million tons of taconite pellets, operating 24 hours per day, year-round, employing 770 people. Through 2013, HTC has produced 260 million gross tons of pellets. HTC is jointly owned by Arcelor Mittal (62.3%), Cliffs Natural Resources (23%), and US Steel Canada (14.7%).

The HTC plant is located approximately four miles northwest of Hibbing, Minnesota (Fig.2), just north of the Laurentian Divide. The initial taconite pit was developed in 1975. Since inception, this pit has expanded east, west, and south along the northern crest of the historic Hull-Rust Mahoning natural ore mine. The Hull-Rust Mahoning Mine, actually a combination of 30 separate mines, was developed along an east/west-trending fault structure and operated from 1895 to 1979. Material movement from this "largest open pit iron mine in the world" totaled more than 1.1 billion tons.



Figure 2. Airphoto image of Hibbing Taconite's plant, tailings facility, and mines that extend nearly 7 miles along the strike of iron-formation.

The four main subdivisions of the Biwabik Iron Formation are present in the vicinity of HTC. From bottom to top they are Lower Cherty, Lower Slaty, Upper Cherty, Upper Slaty. Erosion has removed the Upper Slaty and most of the Upper Cherty members within the area of the current pit. Where present, the

Lower Slaty member and the upper 30 feet of the Lower Cherty member are stripped as rock waste. Approximately 150 feet of cherty and slaty taconite is mined from the central portion of the Lower Cherty. The formation strikes northeast and dips 6°-8° to the southeast. The Pokegama Quartzite forms the footwall of the iron-formation and outcrops to the north, along the south edge of the divide. The only significant structural features are common, but minor, northwest-trending normal faults. Numerous natural ore mines were located along these oxidized structures.

The taconite mined by HTC averages 20 percent magnetic iron, with the general mineralogy consisting of quartz, magnetite, siderite, ankerite, minnesotaite, stilpnomelane and hematite. Ore units in the Lower Cherty (120 feet thick) are predominantly "cherty" taconite, with 6- to 12-inch-thick massive silicate-chert zones separated by 1/8- to 2-inch-thick slaty bands. The lower two units (30 feet thick) are predominantly "slaty" taconite with inter-bedded argillite, magnetite, and minor hematite forming slaty bands from 2-10 inches in thickness separated by 2- to 4-inch massive cherty zones.

Stripping materials include glacial overburden, waste rock, lean oxidized taconite, and old stockpiles, all varying widely in thickness from area to area. Standard rotary drilling, blasting, electric shovel loading, and 240-ton truck haulage are the mining methods utilized. The processing flowsheet differs significantly from the standard Mesabi Range taconite plant in the area that involves crushing and grinding. By contrast, the HTC plant utilizes autogenous mills, which do not contain grinding media. A single stage of gyratory crushing in one of two 60-inch crushers reduces the crude to 10 inches. This is followed by autogenous grinding in one of nine 36-foot-diameter mills. Water is added and as the ore tumbles, it reduces itself to powder fineness.

Field Trip Stops at Hibbing Taconite:

Stop 1 – The first stop in the field trip will be in Hibbing Taconite's Group 4 to view the stratigraphy of the Biwabik Iron Formation at the mine site. In this currently inactive portion of the pit, overlying stockpiles from natural ore mines, glacial till, Lower Slaty rock stripping and Lower Cherty ore horizons are exposed in the high wall of the mine pit. Opportunity for the collection of typical cherty taconite ore will be available at this site.

Stop 2 – The tour will travel east to Group 2 and view the "footwall" of the mine. Opportunity to collect samples of more slaty, jasper rich ore horizons will be available while discussing the in-pit enrichment processing that is used to improve these horizons prior to plant processing.

Stop 3 – The tour will continue east to view active mining activities in Group 1. Typical mine production activities such as production drilling, loading and hauling will be observed from this location.

Stop 4 – View a production blast or plant tour. Depending on the production blasting schedule and the visibility of the blast from a safe location the tour may have the opportunity to view a production blast, typically involving 500,000 to 1,000,000 tons of taconite ore or overlying rock.

Alternate - If scheduling or location does not allow the viewing of a blast, the tour will view the Hibbing Taconite processing plant. Portions of the process that include crushing, grinding, concentration, balling and induration will be toured to show the flow of material from crude ore to finished product.

Stop 5 – Lunch at mine view in old North Hibbing. The scale of the Hibbing Taconite mining operation will be seen from this scenic overview .

ESSAR STEEL'S NEW TACONITE PIT AND PLANT

Essar Steel Minnesota, LLC, Nashwauk, MN

William Everett (Area Manager of Mining)

Essar Steel Minnesota, LLC (ESML) is an iron ore mining company engaged in the development of a fully integrated iron ore mine and pellet plant located on the western end of the Mesabi Range. The Project is adjacent to the City of Nashwauk located in Itasca County, approximately 15 miles (24 km) west of Hibbing and 20 miles (32 km) east of Grand Rapids. This mining project is probably the last major taconite facility to be built on the Mesabi Range.

The facility construction is nearly 65% completed and will cost \$1.8 billion dollars when finished, with an expected design capacity of 7.0 million tons per year of fluxed, standard and DR-grade pellets. All of the required permits required for construction and operation are in place for the designed capacity. The Crushing & Concentration Facility is separated from the Pelletizing Facility, as shown in the Figure 3. When completed, we believe this mining operation will be one of the lowest cost iron ore pellet producers in North America.

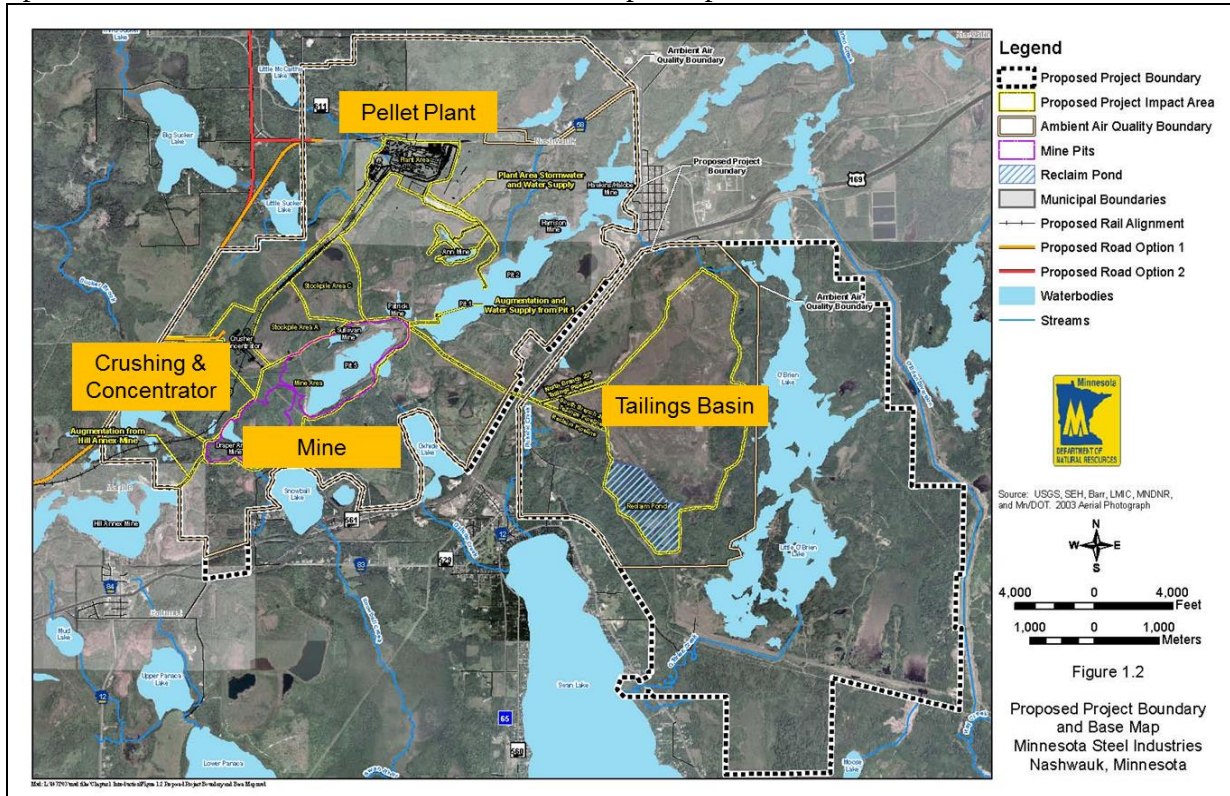


Figure 3 – Essar Steel Minnesota site layout.

The taconite resource at this project site was originally mined by Butler Taconite Mining Company, a company managed by Hanna Mining Company. Butler Taconite was a jointly owned mining operation and was in production from 1967 to 1985. When one of the owners declared bankruptcy, the other two owners closed the operation, and the facility underwent demolition. In 2007, Essar Steel Holdings acquired Minnesota Steel Industries (MSI), a development company owned by some of the mineral owners on the property. In 2008, Essar renamed the MSI to Essar Steel Minnesota, LLC (ESML)

The ESML deposit is a low grade iron-formation with magnetite as the predominate resource. In 2011, ESML conducted a diamond drilling program to bring the deposit into compliance with Canadian

NI43-101 ore reserve standards. A total of 63 diamond drill holes were drilled across the length of the deposit totaling 41,720 feet. The new drilling information combined with the historic drilling, defined the ore zone over the strike length and down dip of the deposit. The drilling program identified 1.7 billion tons of proven and probable ore, having an average stripping ratio of 1.69 and an average weight recovery of 29.1%.

Within the project area, the Biwabik Iron Formation is underlain by the Pokegama Quartzite and is overlain by the Virginia Formation. The Biwabik Iron Formation subcrop and the Virginia Formation are overlain by scattered Cretaceous marine deposits, and all these formations are covered by glacial drift. The Biwabik Formation strikes generally E-NE (065°) with a 5° S-SE dip. The stratigraphy in the ESML Project area was characterized in detail by Hanna Mining Company geologists. The Biwabik Iron Formation has four distinct members: Upper Slaty Member – Slaty non-magnetic taconite; Upper Cherty Member – Cherty weakly magnetic taconite; Lower Slaty Member – Slaty non-magnetic taconite; and Lower Cherty Member – Cherty magnetic taconite. The Lower Cherty Member has been divided into ten distinct subunits. The ore zone lies in the LC4A, LC4B, LC4C, and LC5A subunits of the Lower Cherty Member, and averages about 200 feet thick.

The La Rue fault system runs along the strike length of the formation, bisecting the historic pits within the project area. The Patrick Shear Zone crosscuts the deposit between Pits 2 and 5 of the original Butler Taconite pits. The taconite has been locally oxidized along these fault zones. Figure 4 is a map depicting the geology and historic drilling across the project site. A basic geologic column at the ESML Project site is shown in Figure 5.

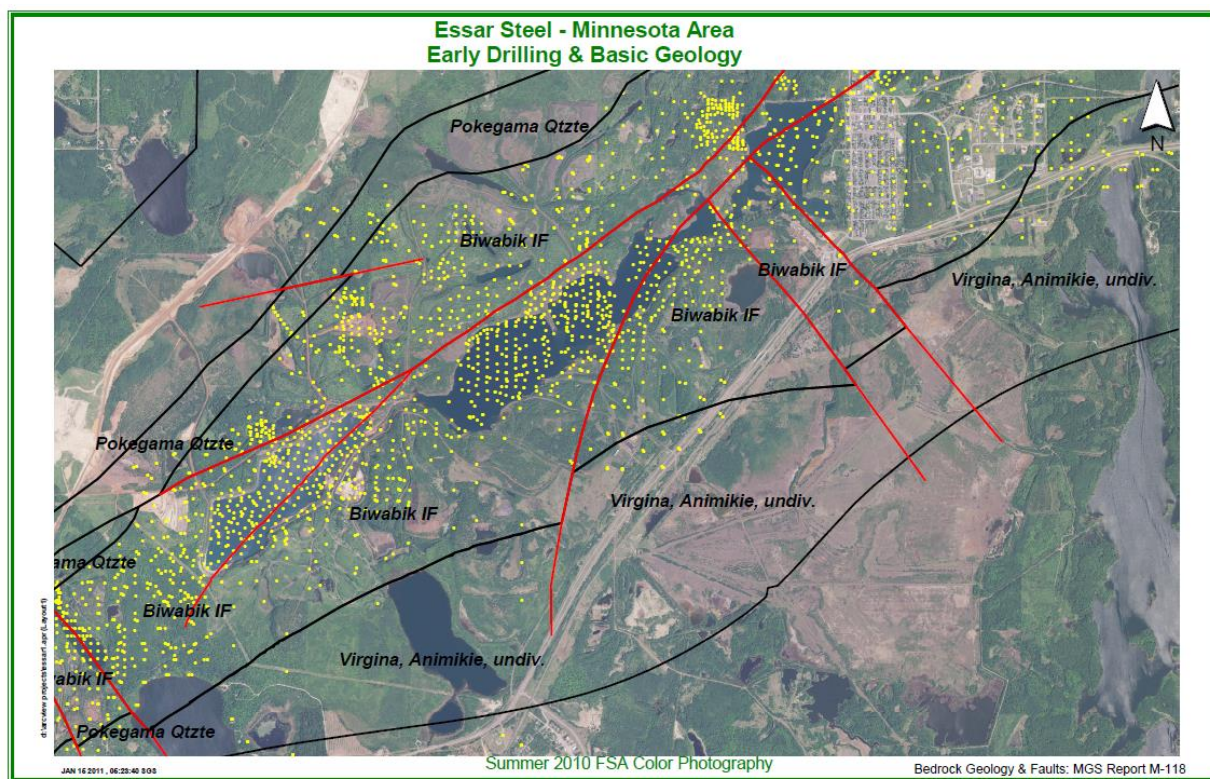


Figure 4. Map of bedrock geology and historic drilling (yellow dots) within the project site. Geologic contacts are black; faults are red.

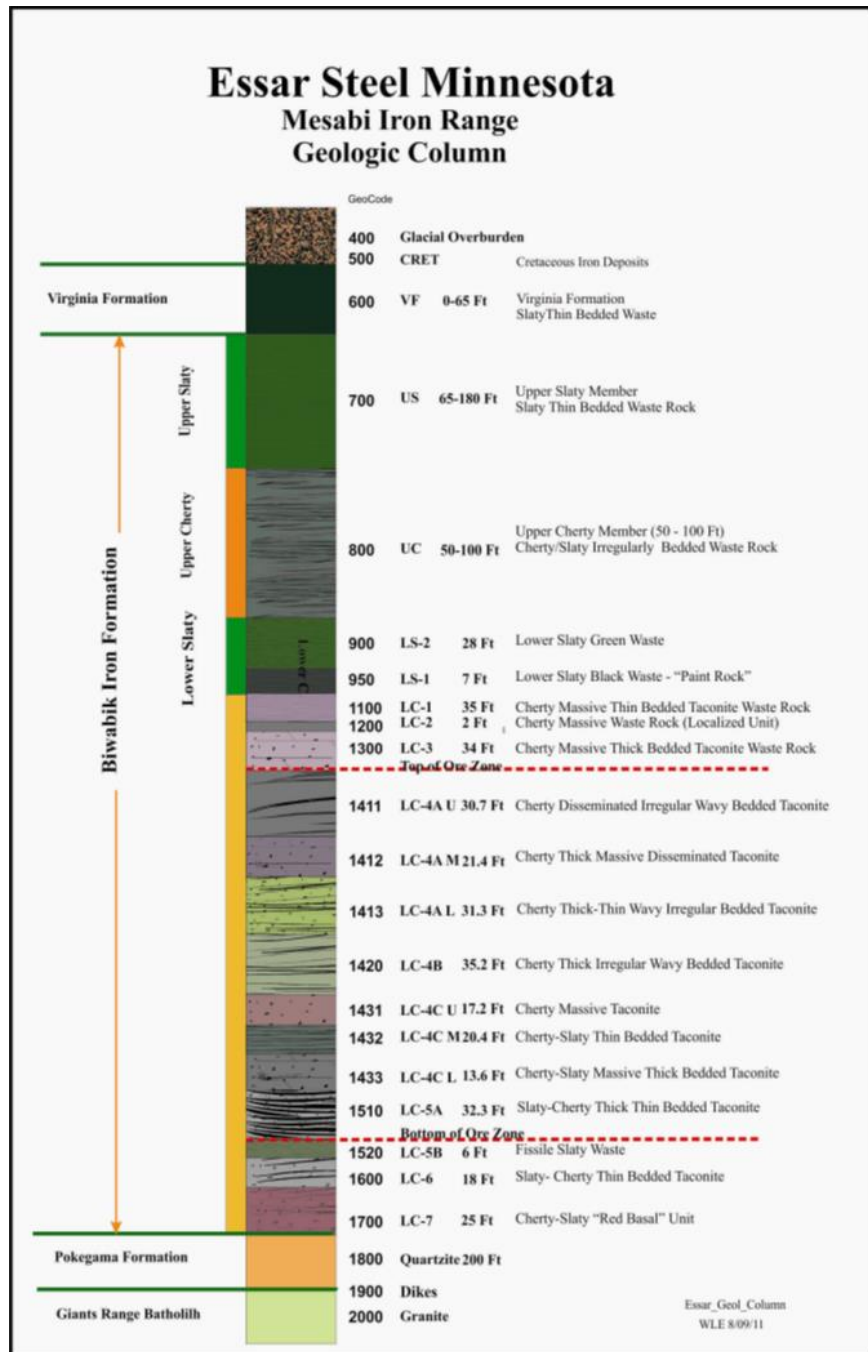


Figure 5. Geologic Column at the Essar property.

During the 2011 diamond drilling campaign, down-dip drill holes intersected a well-defined contact between the Virginia Formation and the Upper Slaty Horizon of the Biwabik Iron Formation. This very distinctive contact zone is characterized by a chaotic mix of broken and deformed strata, uncharacteristic for the uniformly bedded strata above and below this zone. The zone of turbulence matches contacts found in the Gunflint Iron Formation which are attributed to the Sudbury Meteorite Impact event. Above the zone, little evidence of magnetic beds would seem to suggest a disruption in the deposition of ferrous iron minerals by this catastrophic event. Figure 6 provides photographic documentation of the contact in two separate diamond drill holes.



P128 - 361.5 Ft Upper Slaty - Virginia Contact



P130 - 345.5 Ft Upper Slaty - Virginia Contact

Figure 6. Images of core that intersected the Upper Slaty – Virginia Contact.

Taconite mining will start with a development cut adjacent to the new crushing complex and progress down dip into the old Butler Pit 5 taconite pit. Pumping is presently in progress to remove water from the historic mine pit. To date, the water level in Pit 5 has been lowered approximately 60 feet, with another 80 feet needed to reach pit bottom. The mine will be developed using a level bench configuration, hydraulic shovels, and a small fleet of 240-ton haul trucks. The initial mining area has been pre-stripped of glacial overburden north of the Butler Pit 5, and the first blast will be directly in the taconite production zone, as shown in Figure 7 (labeled “Mine Development”).

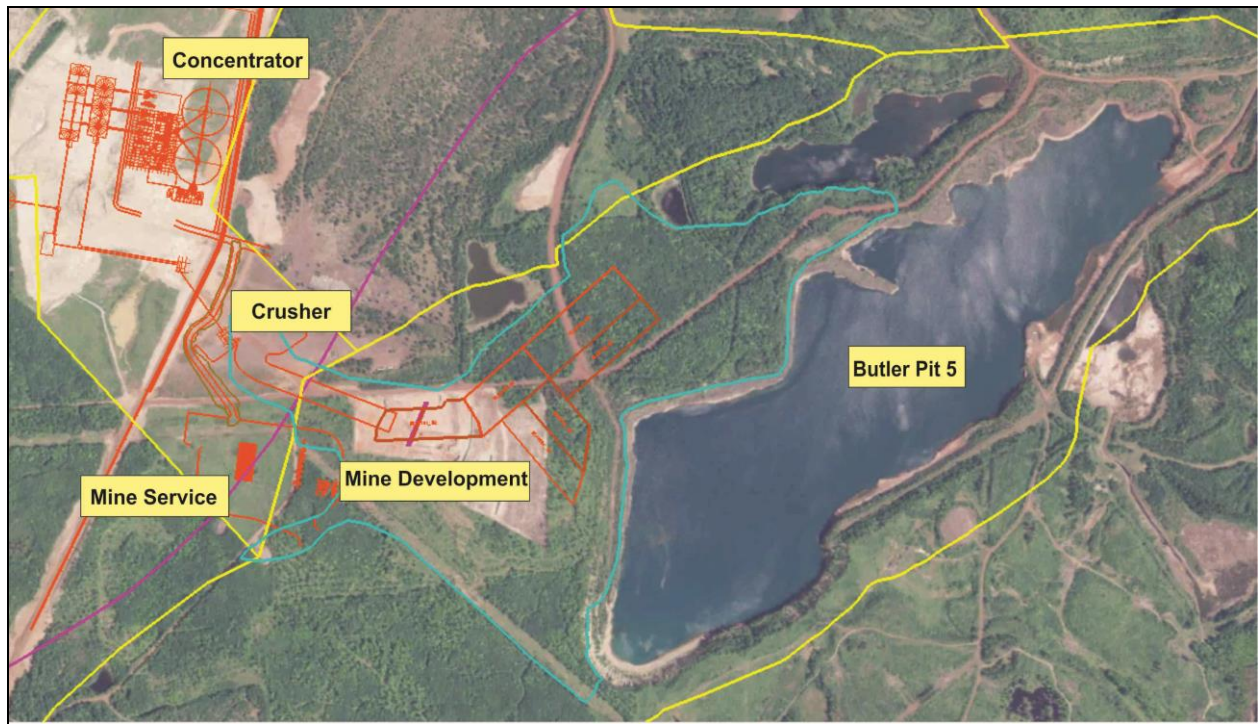


Figure 7. Map showing facilities and planned location for initial mine development adjacent to the historic Butler Taconite pit.

MAGNETATION
Joe Nielsen (Magnetation Inc.).



This trip will visit a unique new iron ore venture currently operating in the Bovey area. Founded in 2006, the privately-held Magnetation Inc. was created with the intention of utilizing magnetic separation technology to capture iron ore particles left over from previous mining operations that existed on the iron range dating back to the 1890s. Owners Al Fritz and Rod Hunt focused the company's early efforts on research and development of a beneficiation process centered on the Ferrous Wheel®, a technology Al Fritz invented in the 1970s that uses permanent magnets to separate iron ore from waste materials and produce an upgraded iron ore concentrate. The location near Bovey is the second of three plants operating on the Mesabi Iron Range; the others are near Keewatin and Chisholm. The Bovey plant commenced operations in May 2012, and produces about 1.2 million tons per-year of iron ore concentrate. The operations consist of excavation and transport of iron-bearing tailings to the concentrator facility, where the iron-rich portions are reclaimed. The resulting iron ore concentrate is then trucked to the Jessie Load-Out, where it is shipped by rail to Magnetation customers. A new pelletizing plant under construction in Reynolds, Indiana, will begin processing ore concentrate late 2014.



The Ore

Magnetation mines the iron ore tailings from mining days long ago. We dig up all the discarded tailings in the **Pit** and bring them to the **Plant** one truck load at a time. In the plant the tailings go through various stations to become Iron Concentrate. The trucks then take the Concentrate to our train loading station to ship to our customers.



The Pit

Unlike a conventional open mine pit our material requires no drilling and blasting. We dig up what was left behind as waste, remove the iron, and return the material back to the basin to be eventually replanted with vegetation.



The Plant

In the plant, we are using Magnetation's unique processes and equipment to remove the iron from our feed material. We are constantly adjusting processes to accommodate the variations in feed from the pit and continually improving to further the iron yield from the feed material.

www.magnetation.com

FIELD TRIP 5

Saturday, May 17, 2014

VISIONS OF MATURI: THE GEOLOGY OF THE SOUTH KAWISHIWI INTRUSION

LEADER: Dean M. Peterson (Duluth Metals, Ltd.)

INTRODUCTION

Twin Metals Minnesota (TMM), a private joint venture company owned by Duluth Metals Ltd (60%) and Antofagasta plc (40%), is currently finishing a robust prefeasibility study (due mid 2014) to develop the Maturi Cu-Ni-PGE deposit in the northern part of the South Kawishiwi Intrusion (SKI). The mineralization at Maturi is confined to the basal 500' of the intrusion and has been the focus of numerous ILSG-based field trips in recent years. In October 2013, UMD's Precambrian Research Center hosted a workshop on Cu-Ni-PGE deposits in the Lake Superior area and the Duluth Complex fieldtrip guidebook of that workshop (Severson et al., 2013) is perhaps the most up-to-date geologic description of Duluth Complex Cu-Ni-PGE deposits. As such, this field trip is focused mainly on the "rest of the rocks" of the SKI as a way to put the Cu-Ni-PGE ores in better context to the vast majority of the rocks of the intrusion.

The importance of understanding these rocks and the overlying glacial deposits will be ever increasing as the TMM Project goes into bankable feasibility, environmental review and permitting, since virtually all of the water in the region (surficial and deep groundwater) interacts mostly with the "other rocks" of the SKI. Duluth Metals' understanding of the SKI has been facilitated by extensive bedrock geologic mapping in the local region (22,200 outcrops mapped), drilling (2,300,000 feet in 1,556 holes), geochemistry (4,500 tills, 1,800 rocks, and ~110,000 drill core assays), and geophysics (3 recent airborne VTEM surveys covering 178,600 acres, borehole EM surveys, Titan-24 survey, and RIM cross-hole imaging).

REGIONAL GEOLOGIC SETTING, DULUTH COMPLEX

The Duluth Complex and associated intrusions of Keweenawan age (~1.1 billion years) in northeastern Minnesota constitute one of the largest mafic intrusive complexes in the world, second only to the Bushveld Complex of South Africa (Miller et al., 2002). These rocks cover a 2,200 square mile (5,700 square km) arcuate area associated with the two strongest gravity anomalies (+50 and +70 milligals) in North America, implying intrusive roots over 8 miles (13 km) deep (Allen and others, 1997). The comagmatic flood basalts and intrusive rocks underlying much of northeastern Minnesota were emplaced during development of the Mesoproterozoic Midcontinent rift, which can be traced geophysically from exposures in the Lake Superior region along a 1250 mile (2,000 km) long, segmented, arcuate path to Kansas and Lower Michigan. The Duluth Complex is defined as the more or less continuous mass of mafic to felsic plutonic rocks that extends for >170 miles (275 km) in an arcuate fashion from Duluth nearly to Grand Portage (**Fig. 1**). It is bounded by a footwall of Paleoproterozoic sedimentary rocks and Archean granite-greenstone terranes (Peterson and Severson, 2002), and a hanging wall largely of comagmatic, rift-related flood basalts and hypabyssal intrusions of the Beaver Bay Complex. In genetic terms, the Duluth Complex is composed of multiple discrete intrusions of mafic to felsic tholeiitic magmas that were episodically emplaced into the base of a volcanic edifice between 1108 and 1098 Ma.

The geology of the Duluth Complex and adjacent areas has recently been described in two major publications by the Minnesota Geological Survey (MGS). These include a 1:200,000 scale regional bedrock geological map of northeastern Minnesota (Miller et al., 2001), and a comprehensive written description of the geology depicted on this map (Miller et al., 2002), commonly referred to as the "bible" by geologists working on Duluth Complex geology. Readers' interested in more detailed descriptions of

the geologic setting of the Duluth Complex should begin their quest for knowledge by downloading these publications from the MGS website (<http://mgssun6.mngs.umn.edu/pub2/>).

Within the nearly continuous mass of intrusive igneous rock forming the Duluth Complex, four general rock series are distinguished on the basis of age, dominant lithology, internal structure, and structural position within the complex.

Felsic series—Massive granophyric granite and smaller amounts of intermediate rock that occur as a semi-continuous mass of intrusions strung along the eastern and central roof zone of the complex, that were emplaced during early stage magmatism (~1108 Ma).

Early gabbro series—Layered sequences of dominantly gabbroic rocks that occur along the northeastern contact of the Duluth Complex, emplaced during early stage magmatism (~1108 Ma).

Anorthositic series—Structurally complex suite of foliated, but rarely layered, plagioclase-rich gabbroic anorthosite emplaced throughout the complex during main stage magmatism (~1099 Ma).

Layered series—Suite of stratiform troctolitic intrusions that comprises at least 11 variably differentiated mafic layered intrusions that occur mostly along the base of the Duluth Complex. These intrusions were emplaced shortly after the Anorthositic series (~1099 Ma).

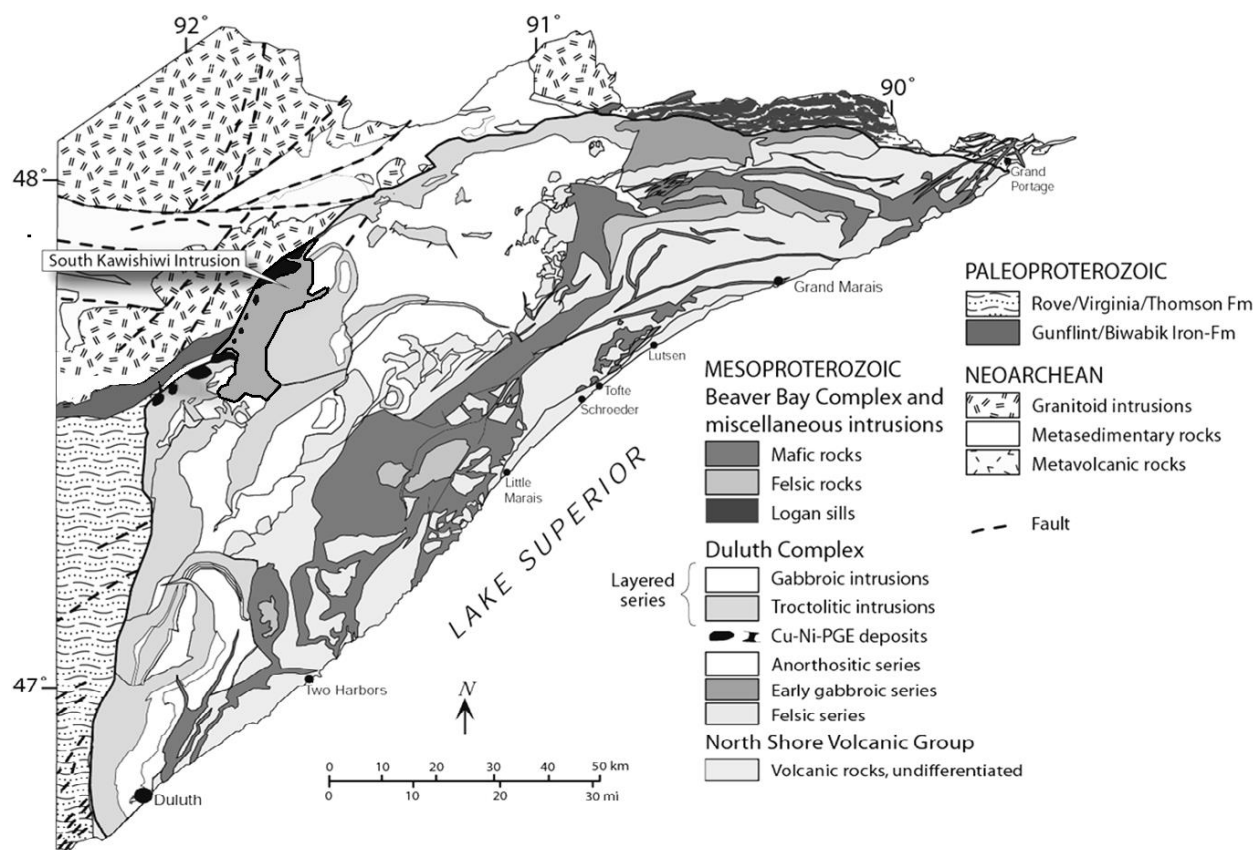


Figure 1. Generalized geologic map of northeastern Minnesota (modified from Miller et al., 2002).

LOCAL GEOLOGIC SETTING, THE SOUTH KAWISHIWI INTRUSION

The SKI consists almost entirely of troctolitic rocks that generally dip gently to the southeast. However, it is not well known that shallow southwesterly dipping troctolite of the upper SKI in the northeastern portion of the intrusion defines an asymmetric funnel-shaped body that emerged from the Nickel Lake macrodiike. The basal mineralization of the SKI is exposed in an arc-shaped area that extends from the Serpentine deposit, in the southwest, to the Spruce Road deposit, in the northeast (**Fig. 2**). Footwall rocks include the Paleoproterozoic Virginia Formation, Biwabik Iron Formation and Archean Giants Range Batholith, the latter is the dominant footwall rock type.

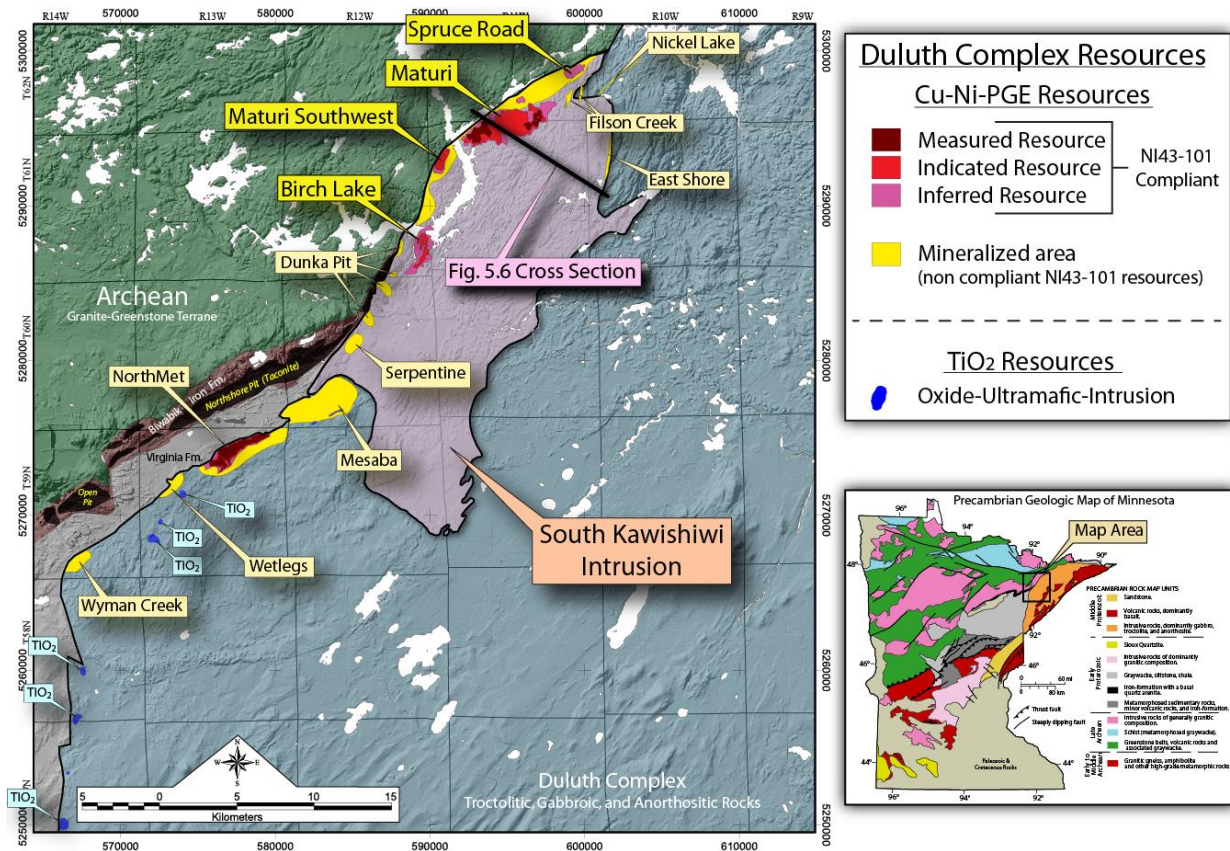


Figure 2. Simplified geological and ore deposit map of the northwestern Duluth Complex.

GEOLOGIC MAPPING

Detailed geological mapping, generally at a scale of 1:5,000 or greater, has been ***the most important component*** of Duluth Metals' understanding of the geology of the SKI. Geologists associated with the company have mapped over 17,000 outcrops (~1,400 total acres of outcrop covering >77,000 acres of ground) within the SKI and adjacent rock units. True understanding of the SKI rocks in their natural environment, ***In The Field***, has led to much improved interpretation of the rocks observed in drill core, the geochemistry of glacial tills, and the interpretation of geophysical studies. Geologists working on the mineral deposits within the Duluth Complex that have not spent a considerable amount of time in the field mapping the rocks will have a difficult time interpreting what they see and log in drill core. Such *in the field* knowledge is especially important as projects advance and true 3D geological models have to be

constructed for mine planning in feasibility studies. All geological interpretations will be scrutinized and audited once the banks get involved in project financing.

A historical account of geological mapping programs within the SKI is presented in **Figure 3**, and the aerial extent of mapped rock types within the bounds of the SKI are given in **Table 1**. It is important to note that there are nearly 4,600 acres of sulfide-bearing bedrock exposed on the Earth's surface within the SKI (1,230 gossanous outcrops mapped).

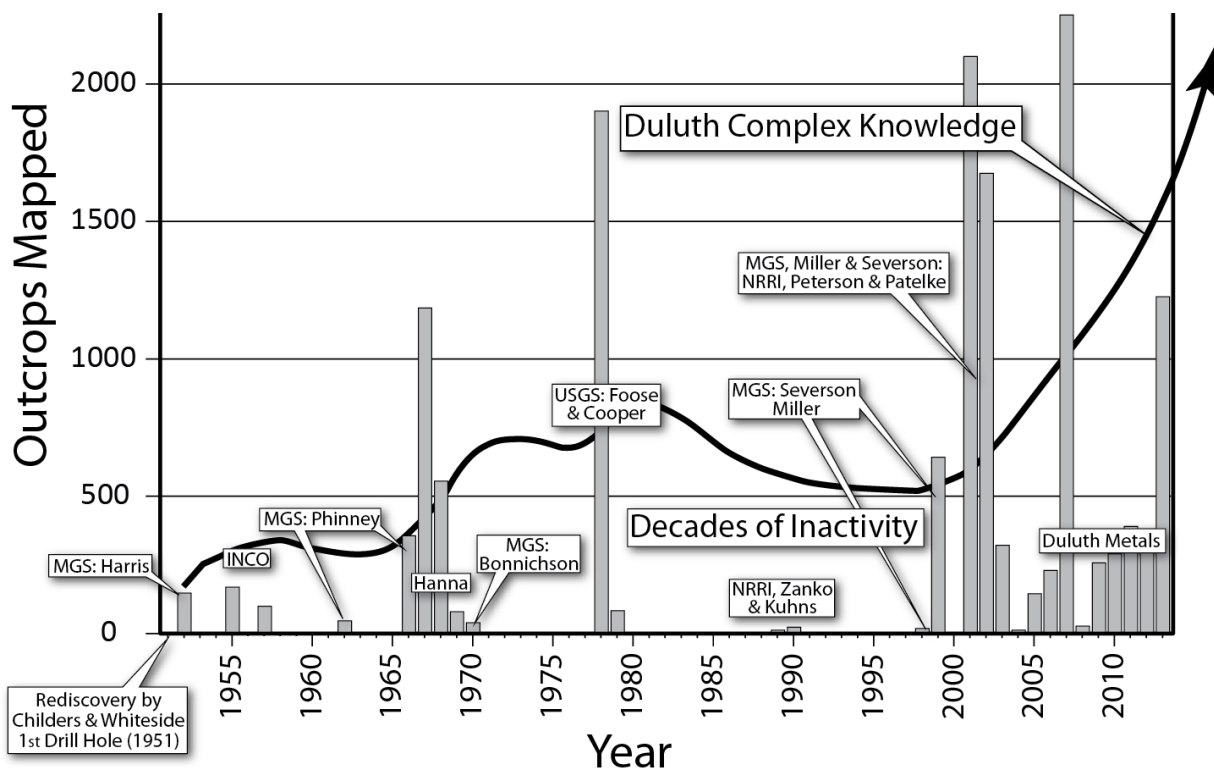


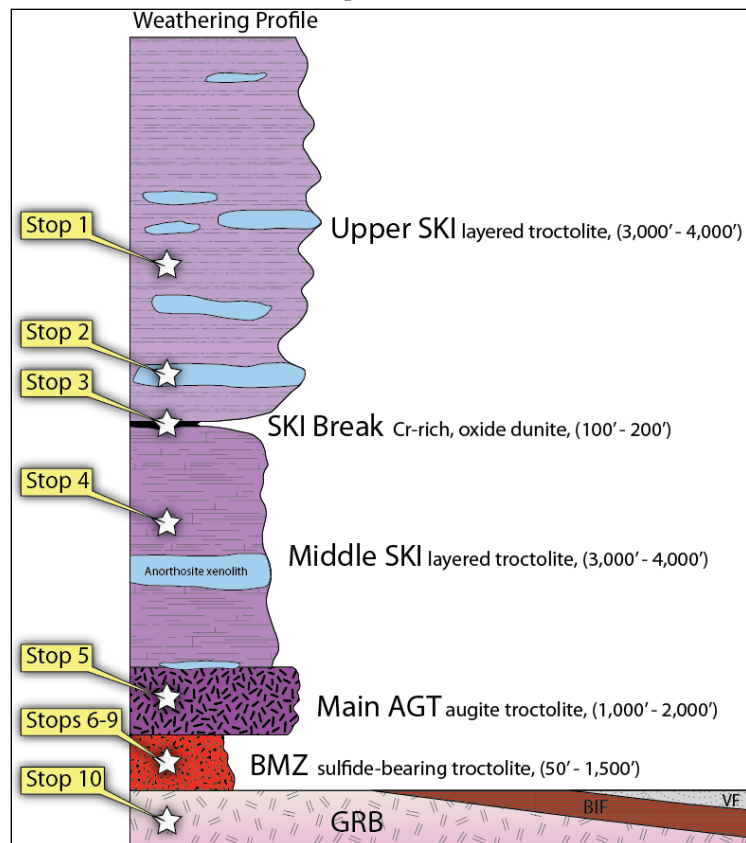
Figure 3. History of geological mapping in the SKI.

Table 1. Distribution of rock types exposed on the surface in the SKI.

Rock Type	Acres	% Area
Diabase dike	1	0.00%
Iron Formation xenoliths	6	0.01%
Sandstone xenoliths	14	0.02%
Ultramafic rocks	14	0.02%
Gabbroic xenoliths	206	0.31%
Basalt xenoliths	482	0.72%
Anorthosite xenoliths	3,112	4.67%
Sulfide-bearing Troctolite	4,594	6.89%
Augite Troctolite	5,554	8.34%
Anorthositic Troctolite	20,364	30.56%
Troctolite	32,286	48.45%
Grand	66,633	100.00%

IGNEOUS STRATIGRAPHY & LITHOGEOCHEMISTRY

Integration of geological, geochemical, and geophysical data over the last few years by Duluth Metals has resulted, via integration, in a new interpretation of the bulk igneous stratigraphy of the SKI. The proposed new stratigraphy of the intrusion is presented in **Figure 4** and consists of five regionally extensive units. From the top down these units include:



Upper SKI – Medium to coarse-grained, locally layered troctolite and anorthositic troctolite. Well layered as defined by olivine-rich horizons.

SKI Break – Chromium oxide-rich, heterogeneous dunite and melatroctolite. Interpreted to be a magmatic unconformity within the SKI.

Middle SKI – Medium to coarse-grained, locally layered troctolite and anorthositic troctolite. Layering defined by olivine.

Main AGT – Coarse-grained, homogeneous, augite troctolite with high-density ophitic Augite grains. This unit is never layered and is interpreted to be the solidified basaltic liquid that carried the phenocrysts and immiscible sulfide droplets of the BMZ.

BMZ – Heterogeneous, sulfide-bearing troctolitic rocks. Interpreted to have formed from a sulfide-rich, crystal-laden magmatic slurry.

Figure 4 Interpreted igneous stratigraphy of the SKI.

A simplified composite lithogeochemical compilation profile through the SKI is presented in Figure 5. This geochemical compilation clearly displays the strong correlation of economically important base (Cu, Ni) and precious (Pt, Pd, Au) metals into the basal mineralized zone (BMZ) and adjacent footwall granitoids. In addition, the common 3:1 Cu:Ni ratio of the BMZ is clearly completely different than the vast majority of the intrusion, where in fact Cu averages about 100 ppm and Ni averages about 200 ppm. Strong olivine layering in the Middle and Upper SKI can easily be seen in the Mg % profile and the break between the Middle and Upper SKI is clearly displayed in the Cr (ppm) profile.

A geologic cross section roughly along Minnesota State Highway #1 is presented in Figure 6, and integrates hundreds of thousands of feet of drilling into this new regional context. Note the extremely large xenolith of Anorthositic Series rocks and North Shore Volcanic Group lavas in the center of the intrusion. This xenolith of older rocks played an important role in the development of higher grades of Cu, Ni, and PGEs in the Maturi deposit compared to other deposits in the district (Peterson and Boerst, 2013).

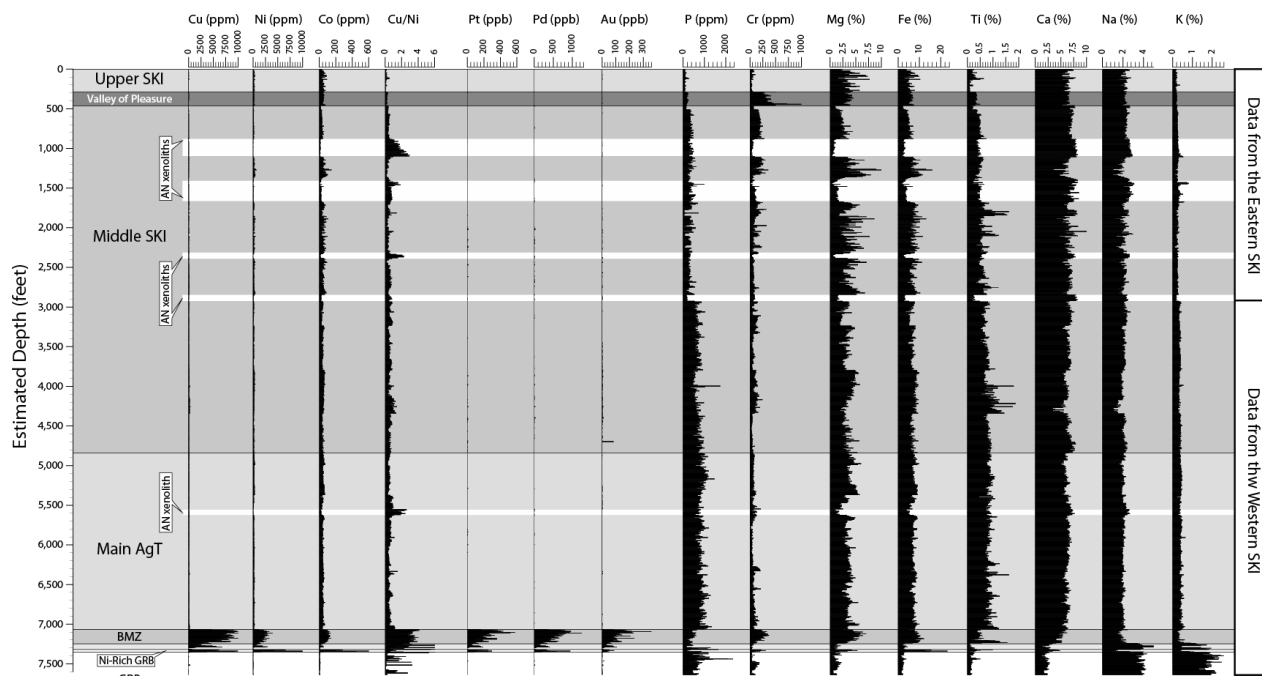


Figure 5. Generalized lithogeochemical compilation profile through the SKI. Data from continuous sampling of Duluth Metals drill holes within the western SKI in the Maturi Deposit in 2007 – 2009 (drill holes MEX-072, Mex-109, and MEX-155), and along the northeastern margin of the intrusion in 2012-2013 (drill holes 12-DM-14, 12-DM-15, and 13-DM-45).

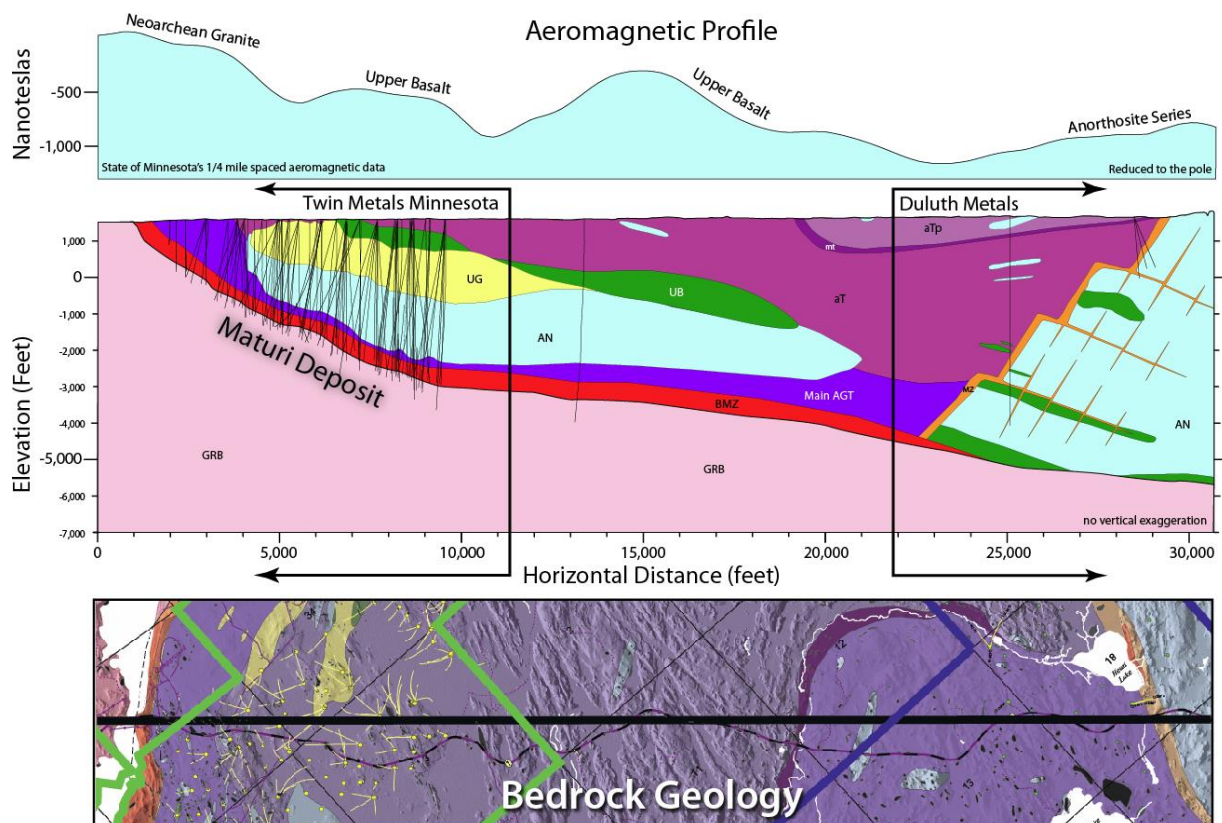


Figure 6. Geologic cross section through the northern South Kawishiwi Intrusion.

DRILLING

Exploration for Cu-Ni deposits at the base of the Duluth Complex began in 1948, about 12 miles southeast of Ely, MN, when strongly mineralized rocks were uncovered in an excavation used to source road material for Spruce Road. Local prospector Fred S. Childers of Ely noted copper stains in the material and he, along with Roger V. Whiteside of Duluth, began searching along the basal contact in the vicinity of the Kawishiwi River. In 1951, they diamond drilled a 188 foot deep hole and intersected mineralized troctolite that averaged 0.36% Cu and 0.13% Ni. In 1952, the International Nickel Company (INCO) began intensive exploration efforts along the zone that coincided with the basal contact and eventually picked up the Childers-Whiteside properties (Spruce Road and Maturi deposits).

Since this initial hole, an additional 1,555 holes (**Fig. 7**) have been drilled in the intrusion (2,300,000 feet of total drilling) in eight prospective areas (Maturi, Spruce Road, Dunka Pit, Maturi SW, Serpentine, Filson Creek, Birch Lake, and East Shore).

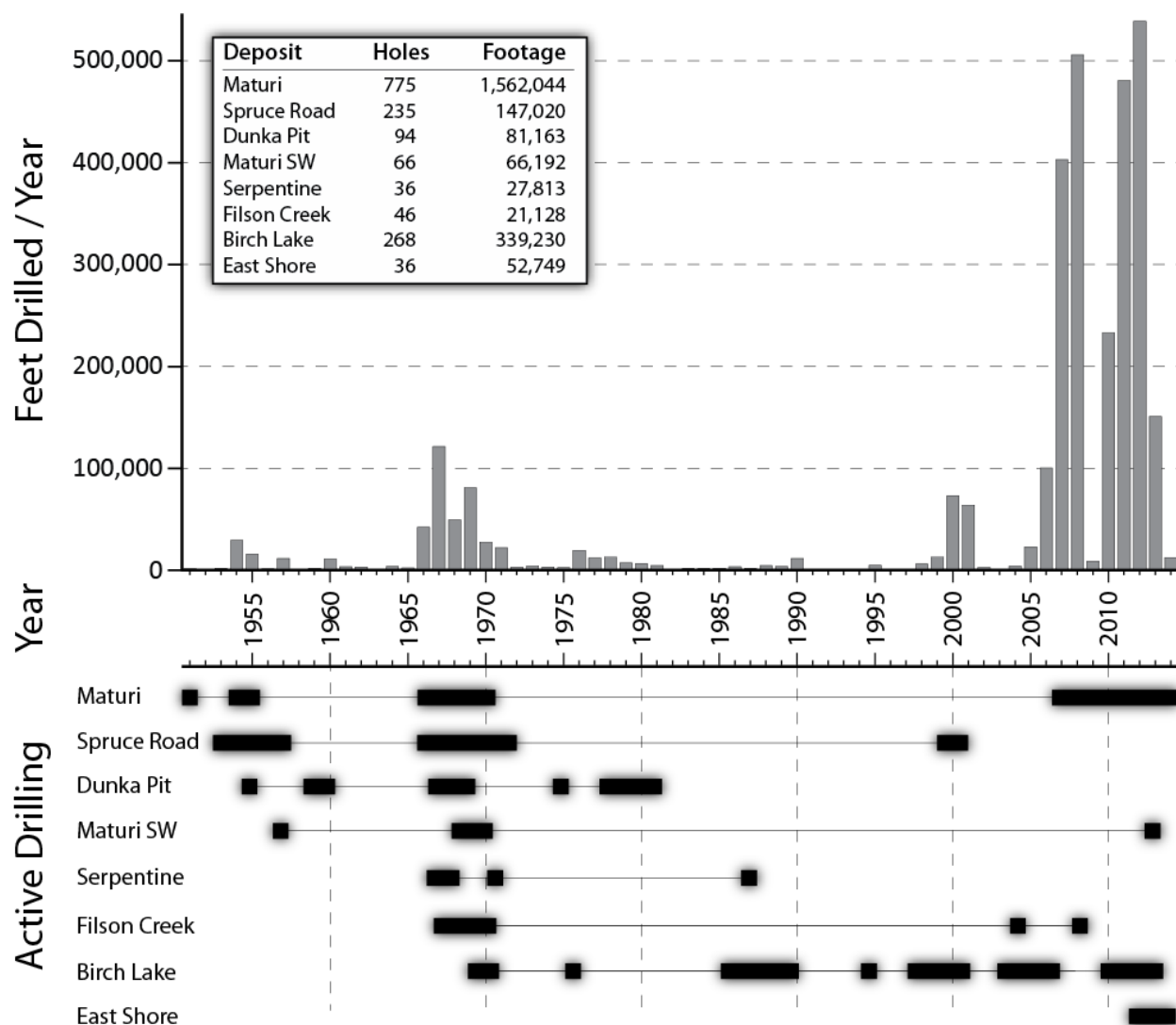


Figure 7. Historic perspective of the amount of drilling completed in the SKI since 1951.

ORE DEPOSITS, TWIN METALS MINNESOTA

A detailed description of Maturi deposit has recently been published (Peterson and Boerst, 2013) and field trip participants that wish to delve deeper into the geology and geochemistry of the basal mineralized zone (BMZ) should acquire that guidebook. However, since publication of that deposit description, Duluth Metals has received an updated independent NI 43-101 Technical Report completed by AMEC

E&C Services Inc. (AMEC) on the Maturi and Maturi SW deposits. The extent of the resource categories for the Maturi, Maturi SW, Birch Lake, and Spruce Road deposits are presented in **Figure 8**.

The updated study utilizes 922 drill holes and 312 wedge offsets, and reports a significant portion of the Maturi deposit upgraded to the Measured Resource category. The mineral resources have been estimated using CIM Definition Standards for Mineral Resources and Reserves dated November 2010.

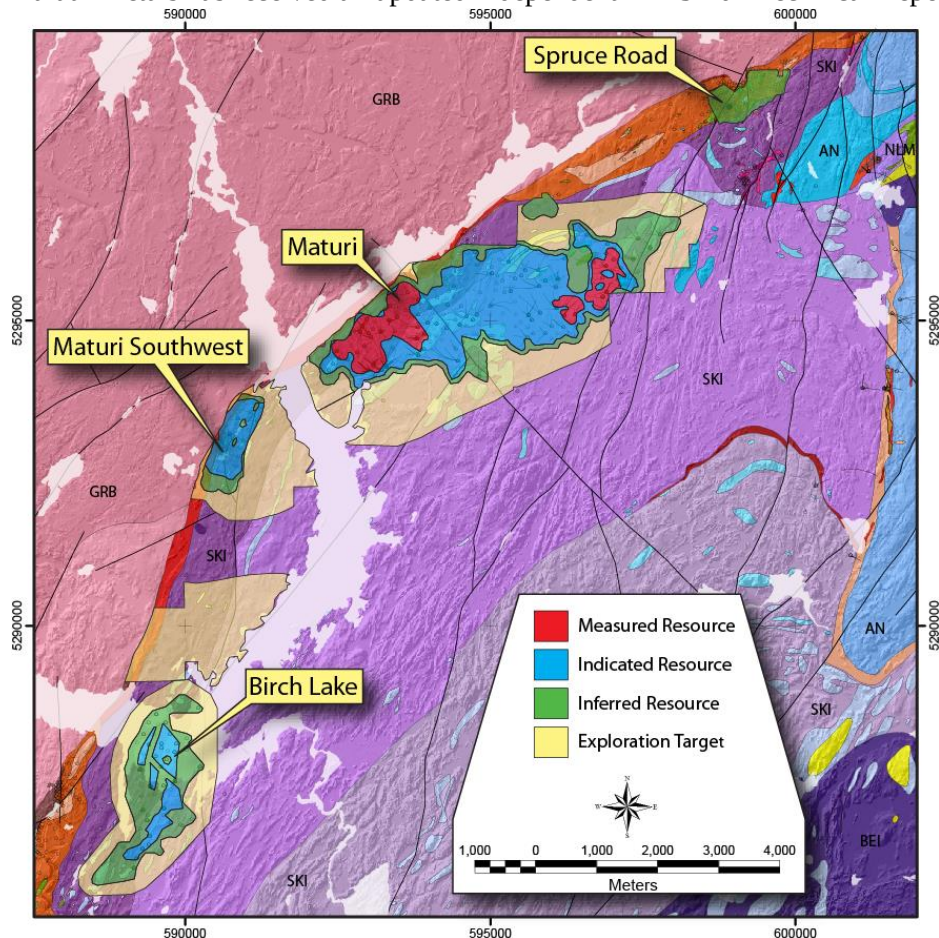


Figure 8. Map of the NI 43-101 qualified resources of Twin Metals Minnesota within the SKI.

The majority of the increase in total contained metals in the 2014 resource estimates reflects the addition of the Maturi Southwest Deposit. The updated mineral resources estimate has 295 million tons in the Measured category at a 0.3% copper cut-off in the Maturi Deposit, which may potentially provide an early start-up area for future mining. The change in category for a significant portion of the Indicated Resource to Measured Resource reflects the excellent continuity of the resource demonstrated by the close-spaced fence drilling completed at Maturi.

Base case qualified resources for these deposits are given in Table 2, and the combined metal contents for the measured, indicated, and inferred mineral resources are provided in Table 3. The enormity of the metal resource in these deposits is certainly quite staggering and clearly shows why so much time, effort, and money has been spent on these deposits in recent years.

Table 2. Twin Metals Minnesota's NI 43-101 base case Qualified Mineral Resources

Deposit Name	Resource Class	Cut-off Cu (%)	Tons (Mst)	Cu (%)	Ni (%)	Pt (ppm)	Pd (ppm)	Au (ppm)
Maturi	Measured	0.3	295	0.63	0.20	0.148	0.345	0.084
Maturi	Indicated	0.3	774	0.58	0.19	0.160	0.360	0.085
Maturi	Inferred	0.3	562	0.51	0.17	0.138	0.317	0.071
Maturi Southwest	Indicated	0.3	103	0.48	0.17	0.080	0.185	0.048
Maturi Southwest	Inferred	0.3	32	0.43	0.15	0.065	0.157	0.041
Birch Lake	Indicated	0.3	100	0.52	0.16	0.233	0.511	0.114
Birch Lake	Inferred	0.3	239	0.46	0.15	0.180	0.370	0.087
Spruce Road	Inferred	0.3	480	0.43	0.16			

Table 3. Contained Metals in the TMM Resource (effective date October 8, 2013) *

Metal	Measured Resource	Indicated Resource	Measured + Indicated	Inferred Resource
Copper	3.7 billion lbs.	11.0 billion lbs.	14.7 billion lbs.	12.3 billion lbs.
Nickel	1.2 billion lbs.	3.5 billion lbs.	4.7 billion lbs.	4.2 billion lbs.
Platinum	1.3 million ozs.	4.5 million ozs.	5.8 million ozs.	3.6 million ozs.**
Palladium	3.0 million ozs.	10.2 million ozs.	13.2 million ozs.	8.0 million ozs.**
Gold	0.7 million ozs.	2.5 million ozs.	3.2 million ozs.	1.8 million ozs.**

* Based on mineral resources estimated at base case 0.3% Cu cut-off grade; for tons and grade see Tables 2 further below.

** Contained ounces of platinum, palladium, and gold in the Inferred category do not include the Spruce Road deposit.

Additional exploration potential highlighted by AMEC outside of the four mineral resources (Maturi, Maturi Southwest, Birch Lake and Spruce Road deposits) and in addition to the TMM defined mineral resource are considered targets for further exploration. An estimate of the exploration potential is between 1.3 to 2.1 billion tons contiguous to the boundaries of the four deposits (**Fig. 8**).

For Maturi and Maturi Southwest, AMEC Assumed that mining, processing and G+A costs would be approximately \$15/t, \$8/t and \$2.50/t respectively for a total of \$25.50/t. At Birch Lake and Spruce Road, AMEC assumed that mining, processing and G+A costs would be approximately \$16/t, \$12/t and \$2/t respectively for a total of \$30/t. This indicates a breakeven NSR of approximately \$30 per ton. Resources meeting an NSR cutoff of \$30/t approximately equate to a copper cutoff of 0.3%.

The geological subunit within basal mineralized zone (BMZ) in the Maturi Deposit, known as the Stage 3 (Peterson and Boerst, 2013), hosts higher grades ore (172 million short tons at 0.72% Cu, 0.23% Ni, 0.188 ppm Pt, 0.438 ppm Pd and 0.104 ppm Au in the Measured category) that is a subset of the base case mineral resource estimate that may have potential as an early start-up area.

The AMEC 2014 Technical Report update on the Measured, Indicated and Inferred categories is presented below for Maturi (**Table 4**), Maturi SW (**Table 5**), Birch Lake (**Table 6**), and Spruce Road (**Table 7**) deposits.

Table 4. Maturi Deposit Mineral Resources by Copper Cutoff Grade (base case is highlighted)

Measured Mineral Resources						
Cut-off Cu (%)	Tons (Mst)	Cu (%)	Ni (%)	Pt (ppm)	Pd (ppm)	Au (ppm)
0.2	312.5	0.61	0.20	0.143	0.334	0.081
0.3	295.3	0.63	0.20	0.148	0.345	0.084
0.4	262.2	0.66	0.21	0.157	0.366	0.089
0.5	224.9	0.70	0.22	0.168	0.392	0.094
0.6	174	0.74	0.24	0.179	0.419	0.101
Indicated Mineral Resources						
0.2	829.4	0.56	0.18	0.154	0.345	0.082
0.3	774.2	0.58	0.19	0.160	0.360	0.085
0.4	678.2	0.61	0.20	0.171	0.384	0.091
0.5	518.2	0.66	0.21	0.192	0.431	0.101
0.6	366.8	0.71	0.22	0.209	0.470	0.109
Inferred Mineral Resource						
0.2	804	0.43	0.14	0.117	0.265	0.060
0.3	562	0.51	0.17	0.138	0.317	0.071
0.4	399	0.57	0.19	0.162	0.370	0.083
0.5	266	0.63	0.20	0.194	0.437	0.097
0.6	147	0.70	0.22	0.233	0.523	0.114

*Effective Date is 8 October 2013.

*Dr. Harry Parker, RM SME, is the QP for the estimate and is a Professional Geologist licensed in Minnesota.

*The resources are based on a US\$30/t NSR that assumes a mining cost of \$15.00/t, a process cost of \$8.00/t and G&A charges of \$2.50/t; global metallurgical recoveries of 94.3% (Cu), 60.8% (Ni), 82.3% (Au), 36.1% (Pd), and 42.5% (Pt); and long-term consensus metal prices of \$3.00/lb Cu, \$9.50/lb Ni, \$1,200/troy oz Au, \$700/troy oz Pd and \$1,650/troy oz Pt.

*The NSR equates to a 0.3% Cu cut-off grade.

*Figures have been rounded and may not sum.

* Mst = million short tons.

Table 5. Maturi Southwest Deposit Mineral Resources by Copper Cutoff Grade (base case is highlighted)

Indicated Mineral Resources						
Cut-off Cu (%)	Tons (Mst)	Cu (%)	Ni (%)	Pt (ppm)	Pd (ppm)	Au (ppm)
0.2	131	0.43	0.15	0.071	0.164	0.042
0.3	103	0.48	0.17	0.080	0.185	0.048
0.4	71	0.53	0.18	0.093	0.217	0.055
0.5	40	0.59	0.20	0.108	0.256	0.064
0.6	16	0.67	0.22	0.124	0.294	0.071
Inferred Mineral Resource						
0.2	57	0.35	0.13	0.052	0.126	0.033
0.3	32	0.43	0.15	0.065	0.157	0.041
0.4	16	0.51	0.17	0.082	0.197	0.050
0.5	7.2	0.60	0.20	0.102	0.251	0.063
0.6	3.2	0.66	0.22	0.115	0.279	0.069

*Footnotes the same as in Table 4.

Table 6. Birch Lake Deposit Mineral Resources by Copper Cutoff Grade (base case is highlighted)

Indicated Mineral Resources						
Cut-off Cu (%)	Tons (Mst)	Cu (%)	Ni (%)	Pt (ppm)	Pd (ppm)	Au (ppm)
0.2	111.9	0.49	0.15	0.217	0.474	0.106
0.3	99.7	0.52	0.16	0.233	0.511	0.114
0.4	85.4	0.55	0.17	0.247	0.543	0.120
0.5	54.9	0.60	0.18	0.269	0.591	0.130
Inferred Mineral Resource						
0.2	313.1	0.41	0.13	0.156	0.320	0.076
0.3	239.2	0.46	0.15	0.180	0.370	0.087
0.4	158.4	0.51	0.16	0.203	0.423	0.098
0.5	76.8	0.58	0.18	0.228	0.480	0.111

* Effective Date is 15 September 2012.

* Dr. Harry Parker, RM SME, is the QP for the estimate and is a Professional Geologist licensed in Minnesota.

* The resources are based on a US\$30/t NSR that assumes a mining cost of \$16.00/t, a process cost of \$12.00/t and G&A charges of \$2.00/t; global metallurgical recoveries of 90.8% (Cu), 57.4% (Ni), 863.3% (Au), 63.6% (Pd), and 55.2% (Pt); and long-term consensus metal prices of \$3.00/lb Cu, \$9.38/lb Ni, \$1,050/troy oz Au, \$850/troy oz Pd and \$1,840/troy oz Pt.

* The NSR equates to a 0.3% Cu cut-off grade.

* Figures have been rounded and may not sum.

* Mst = million short tons.

Table 7. Spruce Road Deposit Mineral Resources by Copper Cutoff Grade (base case is highlighted)

Inferred Mineral Resource			
Cut-off Cu (%)	Tons (Mst)	Cu (%)	Ni (%)
0.2	674	0.38	0.14
0.3	480	0.43	0.16
0.4	254	0.50	0.18
0.5	101	0.57	0.21

* Effective Date is 15 September 2012.

* Dr. Harry Parker, RM SME, is the QP for the estimate and is a Professional Geologist licensed in Minnesota.

* The resources are based on a US\$30/t NSR that assumes a mining cost of \$16.00/t, a process cost of \$12.00/t and G&A charges of \$2.00/t; global metallurgical recoveries of 90.8% (Cu), 57.4% (Ni); and long-term consensus metal prices of \$3.00/lb Cu, \$9.38/lb Ni.

* The NSR equates to a 0.3% Cu cut-off grade.

* Figures have been rounded and may not sum.

* Mst = million short tons.

DESCRIPTION OF FIELD TRIP STOPS

The location of the South Kawishiwi Intrusion field trip stops is presented in **Figure 9**, and short descriptions of the geology and important take-away knowledge (bedrock geology, glacial geology and dynamics, exploration geochemistry, environmental review, etc.) from each stop is given in the following descriptions. It is important to note that the whole northern SKI is located in the scoured bedrock terrain of the Wisconsin cycle of the Pleistocene Laurentide ice sheet (herein the Rainy Lobe about 12,000 years ago). The local end moraine of the Rainy Lobe is located immediately south of Stop #1 along the new Tomahawk Road. Work by Duluth Metals has positively shown that the mean transport length (the distance where $\frac{1}{2}$ of material in the glacial deposits is sourced from) in the field trip area is <0.5 miles.

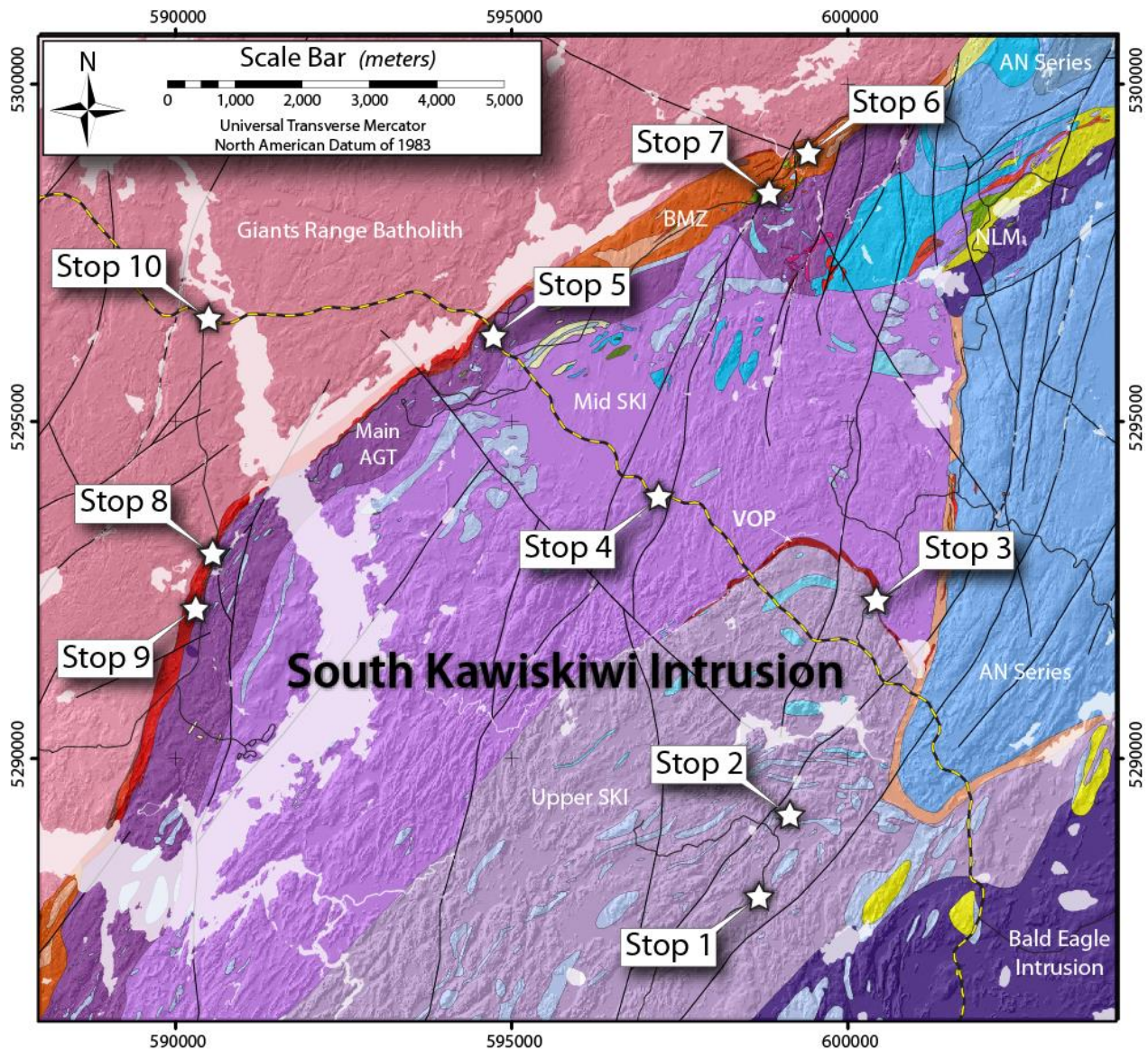


Figure 9. Location map of the field trip stops in the South Kawishiwi Intrusion.

STOP 1: Upper SKI Troctolite

UTM NAD83 Coordinates: 598664E, 5287937N. PLS: T61N, R11W, S26

Glacially scoured outcrops of weakly layered anorthositic troctolite perfectly exposed at the bottom of a State of Minnesota gravel pit. As is typical in much of the Upper and Middle SKI, the rocks dip shallowly to the southeast with olivine layers striking 35° and dipping 8° to the southeast. Exposures such as this are extremely important in environmental reviews of proposed mining operations as they display the natural occurrence of bedrock surfaces beneath the thin veneer of glacial sediments. As we at Duluth Metals have become aware during the course of prefeasibility studies of TMM's Maturi deposit, many hydrogeologists (consulting, governmental, academic) believe that there has to be several hundred feet of fractured bedrock immediately below the Quaternary-Mesoproterozoic contact. If these envisioned fractured bedrock zones exist, they would clearly be permeable (even to the Precambrian geologist author) to near-surface groundwater flow, thus requiring identification and study any environmental review of a mining operation.

Please note the glacial scours and examine the glacial till bank on the north edge of the gravel pit. PGE-enriched sulfide-bearing boulders (with >0.5% Cu) in these glacial deposits have been obtained from this till bank. The nearest identified and exposed up-ice Cu-Ni-PGE occurrence is the Spruce Road deposit, approximately 7 miles to the NNE, which is over 15 times the mean glacial transport distance.

STOP 2: Mesabi Black Quarry, Coldspring Granite Company

UTM NAD83 Coordinates: 599131E, 5289159N. PLS: T61N, R11W, S24

Coldspring's Mesabi Black® quarry opened in 2000 and furnishes the dimension stone industry with poikilitic gabbroic anorthosite. The company utilizes a mix of mining techniques at the quarry to harvest the blocks but the technique that will most interest field trip participants is the diamond wire cuts.

Get ready, this is perhaps the best locality in existence where one can examine Anorthositic Series rocks of the Duluth Complex in 3D and wrap your mind around the viscosity of a plagioclase crystal mush. The quarry is in the heart of the USGS mapping of the Harris Lake area (Foose and Cooper, 1978). Troctolitic rocks in the area dip (as defined by olivine layering and plagioclase crystal foliation) moderately to the southeast at approximately 25°. The operation ships blocks of rocks from a large Anorthositic Series xenolith from within the Upper SKI.

For the ILSG regulars, an anecdote herein is required... Duluth Metals geologists invited Dr. Paul Weiblen to visit the quarry in the spring of 2013. After about an hour looking at Anorthositic Series rocks in perfect 3D cuts, Paul stated to the author (and I quote), "I learned by far more today about the Anorthosite Series than I have over the last 50 years mapping and studying these rocks".

STOP 3: The SKI Break - Middle to Upper SKI Troctolite Contact

UTM NAD83 Coordinates: 600417E, 5292338N. PLS: T61N, R10W, S18

Duluth Metals geologists first took interest in the rocks of this field trip stop in 2011 during follow up field work around several highly anomalous till geochemistry (Cu, Ni, Pt and Pd) samples. Numerous angular boulders of melatroctolite and peridotite were discovered containing highly anomalous Cu-Ni-PGE geochemistry. A detailed mapping and sampling program ensued and the locally sulfide-bearing, chromium oxide-rich ultramafic SKI Break was discovered in outcrops along the valley of Keeley Creek.

This stop will involve a moderately long walk (½ mile) walk through large exposures of anorthosite xenolith-rich, shallow dipping (here to the west-southwest), foliated troctolite of the basal portion of the Upper SKI into the recessive weathering SKI Break. As we walk along U.S. Forest Service road 1468 please note the rugged topography of the glacially scoured outcrops of the Upper SKI. The rugged nature of the topography in the Upper SKI becomes more subdued in glacially polished outcrops of the Middle SKI. Subtle differences in geochemistry/mineralogy/texture of these SKI units must have resulted in differential weathering and saprolite development of these similar troctolite units of the SKI.

STOP 4: Middle SKI Troctolite

UTM NAD83 Coordinates: 597134E, 5293919N. PLS: T61N, R11W, S11

In the summer of 2013, geologists from Duluth Metals and Twin Metals Minnesota completed detailed structural mapping of nearly 100 sites around the Maturi deposit in the initial field phase of an upcoming geohydrology program for the project. Nearly 800 structural elements were measured and 1,321 new outcrops were mapped in detail. The goal of this work was to attempt to define the dip direction and angle of numerous topographic lineaments that cross the SKI (refer back to **Figures 6, 8, and 9**). Many workers and NGO activists believe that each and every one of these topographic features represent deep-seated faults that directly transport groundwater throughout the entire mass of the intrusion. We at Duluth

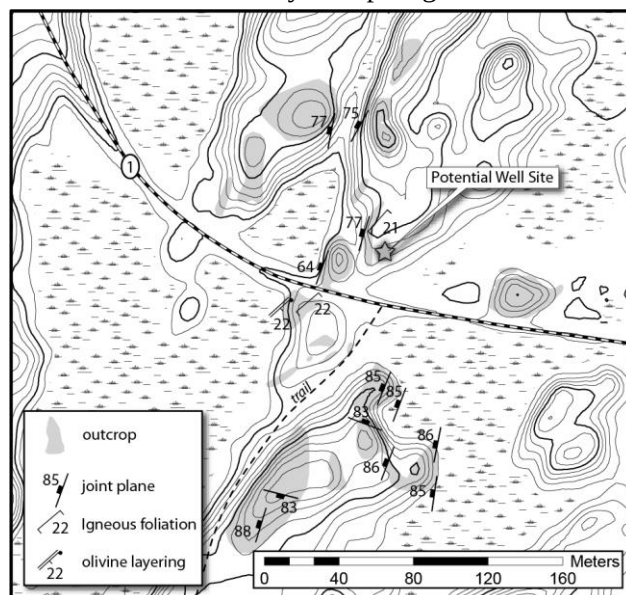


Figure 10. Example of detailed structural mapping.

Metals would beg to differ and it is hoped that a conversation on this topic originates on the outcrop. A simplified single example of this type of geological mapping is presented in Figure 10, and the original field sheet will be available to examine during the field trip.

At this stop, massive, moderately layered and foliated troctolite (strike 45° , dip 22°) of the Middle SKI is exposed in numerous outcrops. This stop epitomizes the “Sea of Troctolite” that occurs throughout the vast majority of the SKI. We will take a short walk to the south onto large massive exposures of glacially scoured troctolite to investigate weak jointing developed along the eastern margin of the bedrock exposures and discuss the interpretation of the NNE trending topographic lineaments.

STOP 5: Main Augite-Troctolite, the “Main AGT”

UTM NAD83 Coordinates: 594743E, 5296267N. PLS: T62N, R11W, S33

Recent road cut along the south side of Minnesota Highway #1 of massive, extremely homogeneous augite troctolite of the Main AGT unit of Severson (1994). Troctolite of the Main AGT unit differs from the Middle and Upper SKI troctolite in two distinctive ways: 1) ophitic augite crystals are black, distinctly associated with Fe-Ti oxides + apatite, and occur as high-density ophitic crystals from 1 to 3 inches in diameter. In the Middle and Upper SKI, ophitic augite crystals are brown, not associated with Fe-Ti oxides, and occur as large (up to 15 inches) low-density grains; and 2) The Main AGT is **never layered**.

Geologists at Duluth Metals interpret the units’ homogeneity and lack of layering as evidence that the Main AGT magma lacked phenocrysts of olivine and plagioclase and represents the end product of top-down and bottom-up solidification of a basaltic liquid. We currently interpret the Main AGT as the solidification of much of the “carrier liquid” of the underlying sulfide-bearing BMZ magmatic slurry.

STOP 6: Basalt Xenolith in the BMZ, the Spruce Road Deposit

UTM NAD83 Coordinates: 599404E, 5298990N. PLS: T62N, R11W, S24

A short field trip stop up onto a small knob of basalt hornfels within the center of the Spruce Road Cu-Ni deposit. Sulfide-bearing troctolitic rocks of the Spruce Road Cu-Ni deposit are distinctly different in several ways to similar rocks within the Maturi Cu-Ni-PGE deposit. First and foremost is the fact that the precious metal content (Pt-Pd-Au) of Spruce Road ores are much less than within Maturi. The second

fact is that the Spruce Road deposit contains a large amount of sulfide-barren xenoliths. In fact, the mapped proportion of barren xenoliths at Spruce Road approaches 15% of the total rocks within the heart of the deposit (**Table 8**). At Maturi, such accessory and exotic xenoliths account for <<1% of the Cu-Ni-PGE mineralized zone.

Table 5-8. Extent of mapped rock types in the Spruce Road deposit.

	Rock Type	Acres	Extent
SKI	Sulfide-bearing, heterogeneous troctolite	187.3	82.5%
	Sulfide-bearing melatroctolite to dunite	6.8	3.0%
Xenoliths	Basalt	14.7	6.5%
	Barren troctolite (early chill margin?)	9.9	4.4%
	Biwabik Iron Formation	5.1	2.3%
	Anorthosite	3.0	1.3%
	Virginia Formation	0.2	0.1%

The difference between these deposits is interpreted to be the result of the timing of magma injection. The Spruce Road deposit is believed to have formed prior to Maturi and the lithology and amount of xenoliths are the end product of the system cleaning out the pathways that the magma traveled upwards from depth (Peterson and Boerst, 2013).

STOP 7: U.S. Forest Service Borrow Pit, BMZ in the Spruce Road Deposit

UTM NAD83 Coordinates: 598826E, 5298384N. PLS: T62N, R11W, S25

Beginning in the late 1940s, the U.S. Forest Service utilized locally derived glacial tills and weathered bedrock gossans as road building materials during the construction of the Spruce Road. As we take a short hike into one of these borrow pits, we will walk by the 1973 INCO bulk sample site in the Spruce Road deposit. This short stop will examine the bottom of an old borrow pit where participants can walk on and sample sulfide-bearing troctolite gossans. Please note the friable nature of the rocks in the weakly saprolitic exposure and look for rounded core-stones where weathering over the eons was less intense.

STOP 8: Sulfide-bearing Troctolite & layered Melatroctolite, Maturi SW Deposit

UTM NAD83 Coordinates: 590572E, 5293036N. PLS: T61N, R11W, S7

Classic roadside exposures of heterogeneous sulfide-bearing troctolite and layered melatroctolite of Severson's (1994) Basal Heterogeneous (BH) and Ultramafic 3 (U3) units of the SKI. A large core-stone is well exposed in the weakly saprolitic heterogeneous troctolite outcrop. Several small xenoliths of fine-grained troctolite can be observed on top of the outcrop and are interpreted as Stage 1 chilled margin autoliths (Peterson and Boerst, 2013). Within the exposure of the overlying U3 layered melatroctolite, olivine layers strike 17° and dip steeply 51° to the ESE. The steep dip is apparently associated with two defined north-south trending faults east of these exposures. Recent drilling by Twin Metals Minnesota in this area has led to the definition of the Maturi SW deposit (see **Fig. 8** and **Tables 2** and **5**).

STOP 9: Basal Heterogeneous, Sulfide-poor Troctolite

UTM NAD83 Coordinates: 590313E, 5292211N. PLS: T61N, R11W, S18

At this stop, we'll examine perhaps the best exposure of Severson's (1994) BH unit in the whole SKI. The heterogeneous troctolitic rocks at this stop are generally poorly mineralized and thus lack a gossanous saprolitic weathering profile which lets one see the true nature of the heterogeneity within the troctolite. I believe that all geologists who ever will log drill core within the Cu-Ni deposits of the Duluth Complex (or who attempt to model such deposits for mine planning purposes) should be required to spend several days examining the rocks within the area around both Stops 8 and 9. All participants should imagine a drill core cutting this exposure and how they would interpret the geology of that core without first examining this outcrop. Such thoughts are why the Precambrian Research Center's field camp has for many years had its students complete a 1:5,000 scale bedrock geology map of this area.

STOP 10: Giants Range Batholith, the Footwall

UTM NAD83 Coordinates: 590518E, 5296544N. PLS: T62N, R11W, S31

The footwall rocks of the whole northern SKI consist of the Neoproterozoic Giants Range batholith (GRB), such as is exposed along Highway 1 at this field trip stop. The rocks here consist of porphyritic hornblende quartz monzonite with distinctive 1-2 cm potassium feldspar phenocrysts. The massive nature of this unit creates an excellent footwall for the intrusions Cu-Ni-PGE deposits as it lacks bedding and thus rarely (if ever) gets incorporated into the mineralized zone as barren xenoliths. In addition, the melting of the GRB beneath long-lived magma channels (Peterson and Boerst, 2013) at the base of the Maturi deposit has contaminated the SKI and induced additional sulfide immiscibility and the formation of Ni- and Co-rich massive sulfide bodies (see the bottom of **Fig. 5**).

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FIELD TRIP 6

Saturday, May 17, 2014

THE ST. LOUIS SUBLOBE AND GLACIAL LAKE UPHAM

LEADERS:

Phil Larson (Duluth Metals Limited)

Alan Knaeble (Minnesota Geological Survey)

Howard Mooers (University of Minnesota Duluth)

with contributions by

Lisa Marlo (Halcon Resources Corporation)

INTRODUCTION

The flat plain lying to the south of the Mesabi Iron Range at first glance stands in stark contrast to the varied topography and geology of northeastern Minnesota. Largely covered by peatland, early travelers on the St. Louis and Savanna Rivers no doubt appreciated their quick passage through the mosquito-infested terrain north to Lake Vermilion and west to Sandy Lake. After iron ore was discovered on the Mesabi, mine developers at first appreciated the gentle grades and straight lines of the railroads vital to transporting iron ore to market, but learned to respect the difficulties inherent in maintaining lines across water-logged ground prone to sink beneath the heavy traffic. Agricultural settlement was almost an afterthought, memorable more for its heroic efforts than lasting success. Even so, on a long drive south from the Range, a traveler cannot help to wonder, “Why so flat, and why here?”

As it turns out, the Glacial Lake Upham Plain (as it is called) is the location of one of the more interesting episodes in the long history of glaciation in Minnesota and the Upper Midwest (Fig. 1). As the last Laurentide Ice Sheet retreated to the north, the long-standing pattern of ice flowing from Hudson Bay in the north to central Minnesota and beyond to the south was dramatically disrupted by a surge of ice from the Red River Valley to the west. As the ice entered the flat Upham Plain, flow took right-turns to either flank, advancing to the southwest as far as Aitkin, and more remarkably, advancing from south to north as far as the Giant’s Range. Despite this dramatic entry, almost as rapidly, it melted away and was gone.

The glacial advance by the St. Louis sublobe was rapid, short-lived, and barely left a mark on the landscape. Nevertheless, the signs of this light touch are there for the careful observer to see. This field trip explores sites illustrating the landforms and sediments associated with the St. Louis sublobe and its associated glacial lakes.

HISTORICAL BACKGROUND

Leverett (1932) first noted the presence of a distinct clayey, calcareous grey till in the region south of the Mesabi Iron Range. He correlated the till with similar fine-textured, calcareous till associated with the Des Moines lobe to the west, and first applied the name St. Louis sublobe to the corresponding glacial advance. He also noted the presence of similar reddish clayey till in the immediate vicinity of the Mesabi, attributing the red color to local incorporation of red hematite iron-ore. He also mapped the extent of the lobe, as well as the associated overlying glacial lake basins, named Glacial Lakes Aitkin and Upham.

Although Wright (1955) initially attributed the red color in the drift to reddish drift from a Superior lobe advance from the southeast, later work by Baker (1964) and Wright and Watts (1969) returned to Leverett’s conclusion that the St. Louis sublobe advanced from the northwest. Baker (1964) recognized that two tills were associated with the St. Louis sublobe advance (named the Alborn Phase by him): a reddish clayey till, and an overlying grey (brown when oxidized) silty till, named the Alborn till and Prairie Lake till, respectively. Wright and Watts (1969) attributed the red color of the Alborn till to

incorporation and mixing of reddish lake clay from Glacial Lake Upham I into a glacial debris load composed of grey (-brown) silty till. The red lake clay was believed due to reddish sediment ultimately sourced from the northern margin of the Superior lobe. Farnham and others (1964) published two radiocarbon dates attributing relatively young dates to the Alborn phase. Wright and Watts (1969) recognized that meltwater from the St. Louis sublobe and its successor lakes flowed around the northern margin of the Thomson-Nickerson moraine during the Nickerson phase of the Superior lobe.

Winter (1971) and Winter and others (1973) investigated St. Louis sublobe deposits along its northern extent in detail. They questioned that the grey(-brown) Prairie Lake till and reddish Alborn till could be sourced from the same provenance, noting the difference in texture and clast composition.

Hobbs (1983) described the history of meltwater flow associated with the St. Louis sublobe and its successor lakes, including the flow of meltwater from Lake Koochiching (potentially a Glacial Lake Agassiz precursor) through the Upham and Aitkin basins. Ballantine (1991) conducted additional stratigraphic investigations in the southern Upham basin. Meyer (1993) defined the St. Louis sublobe as referring only to ice flowing east of the Giant's Range, distinguishing it from the much more extensive northwestern provenance ice advance to the west. More recently, Knaeble and others (2004) and Knaeble and Hobbs (2009) conducted the first detailed modern investigations of the southern margin of the St. Louis sublobe in Crow Wing and Carlton Counties. Marlow (2004) investigated the widespread occurrence of eolian deposits on the beds of Glacial Lakes Aitkin and Upham II, and Jennings and Reynolds (2005) re-mapped glacial stratigraphy along the axis of the Mesabi Iron Range.

GEOLOGY

Regional Background

Bedrock in the area glaciated by the St. Louis sublobe is underlain by shales, mudstone, and greywackes of the Paleoproterozoic Virginia Formation. This formation has proven relatively less resistant to weathering and erosion than the iron-formation and Archean granite-greenstone terrane to the north, the Mesoproterozoic Duluth Complex to the east, and the fold-and-thrust belt of the Paleoproterozoic Penokean orogen to the south. Consequently, a natural basin existed in the area prior to glaciation.

The area was repeatedly glaciated over the course of the Pleistocene. Most recently, during the Wisconsin glacialiation, ice of the Rainy lobe advanced from the north-northeast from the Labradoran sector of the Laurentide ice sheet (bearing approximately 215°), carrying relatively coarse-grained, sandy textured sediment dominated by crystalline igneous and metamorphic rocks of the Canadian Shield. At about the same time, to the south, the Labradoran ice funneled into the Lake Superior basin as the Superior lobe advanced roughly parallel to the Rainy lobe. The Superior lobe carried an abundance of reddish sediment eroded from rift-filling sedimentary rocks of the Mesoproterozoic Midcontinent Rift.

During the waning stages of the last glaciation, the Rainy lobe margin retreated back to the north-northeast across the area. Retreat was characterized by a relatively steady rate of margin retreat, punctuated by the occasional minor re-advance and moraine building event. When the line of retreat reached the region underlain by the Virginia Formation (Animikie Basin), proglacial lakes developed in the natural topographic low, Glacial Lake Aitkin to the southwest and Glacial Lake Upham to the northeast (Fig. 1). These lakes received meltwater-borne sediment from both the Rainy lobe to the northeast and the Superior lobe to the south.

Glacial Lake Upham I

Lake Upham I formed as a proglacial lake as the Rainy lobe ice margin retreated to the northeast. Little is known about the extent or duration of the lake, as its existence is largely inferred from the red clayey lacustrine sediment incorporated into the debris load of the later St. Louis sublobe advance. *In situ* Upham I sediment is rarely observed.

The distinctive red color of Upham I sediment is due to the presence of hematite in the clay size fraction. Leverett (1932) believed this red color reflected erosion and incorporation of soft hematite iron ore from the Mesabi Iron Range to the north. Leith (1903) recognized that significant glacial erosion of soft iron ores had taken place, and suggested that much of this eroded material would be resident in fine-grained glacial sediments. Wright (1955) correlated Upham I sediments with the Superior lobe, believing the red color came from incorporation of red shale from the Mesoproterozoic Midcontinent Rift System. Both are viable hypotheses, and it is not unlikely that both sources contributed to the red color of Upham I sediments.

The lake basin was bounded by stagnant ice-cored topography to the north and west, and glacial ice of the Rainy lobe to the northeast and the Superior lobe to the south. Large patches of ice-cored topography likely persisted in the lake basin, particularly along northwest-southeast oriented moraines formed by the retreating Rainy lobe. The outlet to the lake was likely to the southwest through Lake Aitkin I and ultimately the Mississippi River.

The age of Upham I is constrained by two bracketing events: retreat of the Rainy lobe from the St. Croix moraine in central Minnesota, tentatively dated at no later than ~ 15.1 ^{14}C kyr BP (Birks, 1976; Mooers and Lehr, 1997), and the Alborn phase advance of the St. Louis sublobe. The lake was likely contemporaneous with the Automba phase of the Superior lobe.

St. Louis Sublobe

Gradual retreat of the Rainy lobe and deglaciation of the Upham basin was abruptly interrupted by advance of the St. Louis sublobe from the northwest (Fig. 1). In contrast to the Rainy lobe, formed by ice flowing from the northeast and the Labradoran accumulation center of the Laurentide Ice Sheet, the St. Louis sublobe originated from ice flow southward from the Lake Winnipeg basin into the Red River Valley. For most of the glacial cycle, a relatively high flux of ice over the Canadian Shield of northwestern Ontario and northeastern Minnesota blocked eastward expansion of ice streaming south from Lake Winnipeg, funneling this ice into the Minnesota River Valley. Retreat of the Rainy lobe ice margin from the Itasca-St. Croix moraines opened a low elevation, ice-free corridor to the north of the Itasca moraine. Continued high ice flux in the Red River Valley rapidly surged into this gap, expanding south of the Mesabi Range to the northeast and southwest across the fine-grained lacustrine sediments of Lakes Aitkin I and Upham I. This advance and its associated glacial deposits are grouped as the Alborn phase.

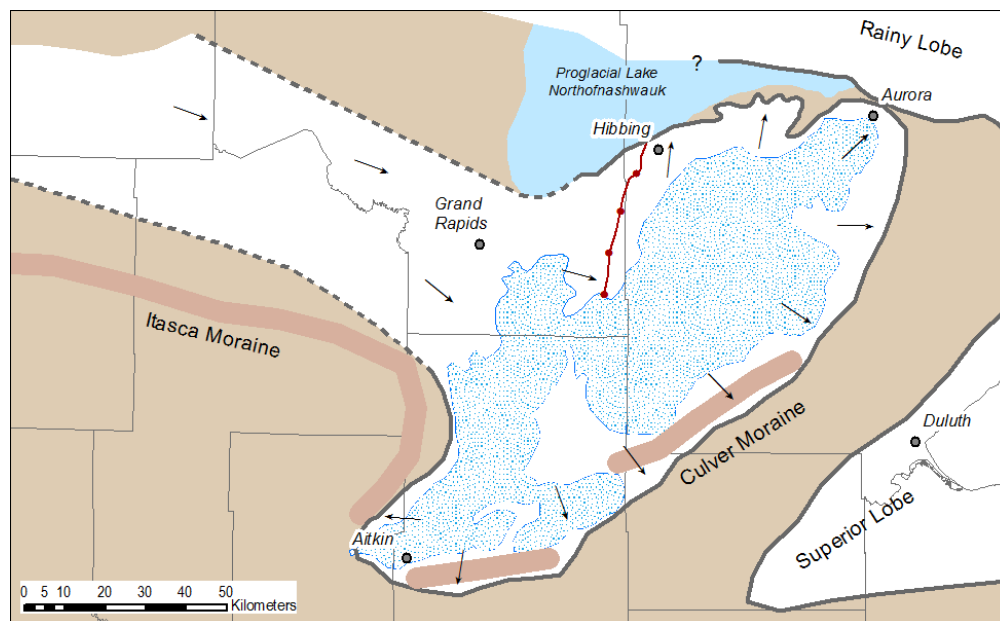


Figure 1. Maximum extent of St. Louis sublobe advances (white) and inferred location of Lakes Aitkin I and Upham I (stippled pattern). Arrows indicate inferred ice flow directions. Western limit of Alborn member/red clayey till along the Mesabi Range (red line) from Winter and others

(1973). Proglacial Lake North of Nashwauk is the high-level (elev. >1500') proglacial lake dammed by the initial phase of the St. Louis sublobe.

Chronology

The absolute timing of the Alborn phase is poorly constrained. As mentioned above, the advance could not have occurred prior to ~15.1 ^{14}C kyr BP. Two radiocarbon dates have been attributed directly to St. Louis sublobe associated deposits (Farnham and others, 1964). A buried soil within Lake Aitkin II sediment near Aitkin, MN was dated at 11.6 ^{14}C kyr BP. Wood recovered from a red clayey till at the Mariska Mine in Gilbert, MN, near the northeastern limit of the Alborn phase, was dated at 11.2 ^{14}C kyr BP. However, the Lake Aitkin II soil was developed on lacustrine sediment, and therefore significantly post-dates the Alborn phase and formation of Lake Aitkin II. The Mariska Mine wood is unlikely to have been incorporated during advance of the St. Louis sublobe into a forested environment. Rather, it perhaps represents wood incorporated into a flow till developed on ice-cored topography, and also significantly post-dates the Alborn phase. These young dates therefore are minimum bracketing dates for the Alborn phase.

Relative age relationships provide additional insight. During the Split Rock phase, meltwater from the Superior lobe apparently flowed northwest from the Cloquet moraine into the ice-free Upham I basin (Wright and Watts, 1969). In the later Nickerson phase, meltwater from the eastern margin of the St. Louis sublobe and from post-glacial Lake Upham II flowed down the proto-St. Louis River and around the Superior lobe margin (Thomson moraine) into the St. Croix River drainage. Significant meltwater flow down the St. Louis River persisted after retreat of the Superior lobe from the Thomson moraine and formation of Lake Duluth. However, meltwater inflow from Lake Upham II into Lake Duluth ceased prior to readvance of the Superior lobe to the western end of the Lake Duluth basin during the Marquette phase (Mooers and others, 2005), dated at 10.0 ^{14}C kyr BP (Lowell and others, 1999).

Glacial Dynamics

The St. Louis sublobe was a thin, temperate glacier. It was on the order of perhaps 100-200 m thick at its maximum extent (Knaeble and others, 2005). The apparent confinement of the lobe by topography during its advance is a direct consequence of this relative thinness; flow was apparently confined by stagnant ice in the Itasca and Outing moraines to the south and west, and by stagnant Rainy lobe ice to the north. Notwithstanding the relatively thin ice, expansion into the Lakes Aitkin I and Upham I basins was facilitated by the low shear strength lacustrine sediment substrate.

The base of the glacier was apparently at the pressure melting point throughout the Alborn phase. Preexisting proglacial lakes precluded development of permafrost in the Aitkin I and Upham I basins, so much of the glacial advance was over a 'warm' bed. Little evidence for freeze-on and entrainment of sediment into glacial ice exists, either in the form of thick Alborn phase glacial sediment accumulations or stagnation topography.

Alborn phase tills tend to be compositionally homogeneous, and relatively uniform in thickness. Local incorporation of basal sediment (lacustrine clays and silts) is evident in the basal Alborn member tills, however erosion and entrainment occurred at very low rates relative to the polythermal Rainy lobe just to the north. Smeared pods of Alborn till are often incorporated into the overlying Prairie Lake member till at the contact. However, the overall composition of Prairie Lake till is homogeneous and distinct from Alborn till, suggesting erosion, entrainment, and mixing of substrate into the debris load was occurred at very low rates. These features suggest sediment transport by the St. Louis sublobe is best explained as having occurred as a subglacial deforming layer.

Other than Lake Upham I lacustrine sediments, Alborn phase deposits show little evidence of erosion, entrainment, or other modification of the older Rainy lobe glacial deposits over which the St. Louis sublobe advanced. Rainy lobe landforms such as eskers, drumlins, and moraines are clearly visible beneath a veneer of Alborn phase drift, and the hummocky 'moraines' found at the margins of the St. Louis sublobe are composed primarily of older, relatively coarse-grained Rainy and Superior lobe drift.

The inability of the St. Louis sublobe to form its own landforms, or modify older landforms, is a consequence of a general lack of englacial sediment, and an inability to transmit shear stress into the substrate.

The relative thinness of the ice and the low shear strength substrate suggest that the St. Louis sublobe advance was as one, or possibly two, surge(s) from the main Red River lobe ice mass. By analogy with modern ice streams in Antarctica and Greenland, ice may have streamed into the Upham and Aitkin basins at flow rates up to 10 km/yr, and may have taken only a few decades for the St. Louis sublobe to reach its maximum extent.

It is unclear how many individual surges, or advances, from the main Red River lobe ice mass may have occurred over the history of the St. Louis sublobe. The two distinct tills associated with the Alborn phase likely record two distinct phases of ice flow through the St. Louis sublobe. The Goodland esker (Knaeble and others, 2005) formed in a large englacial meltwater conduit developed in response to the initial west-to-east St. Louis sublobe advance blocking southward meltwater flow from the Rainy lobe to the north. A reconstruction indicates around 75 m of ice was necessary to block this flow. However, the Goodland esker and the surrounding ice-cored Rainy lobe deposits are mantled by upper (later) Prairie Lake member tills, suggesting this area was later overrun by thicker ice in a later phase of the St. Louis sublobe advance.

Similar to other surging glaciers, the lobe may have stagnated *en masse*, with an abrupt retreat of the 'active' ice margin by as much as 150 km. Ice down-flow of the 'active' margin ceased flowing and melted without forming recessional moraines. The lack of significant englacial debris transport precluded formation of significant supraglacial sediment accumulation, even in marginal areas. Consequently, during stagnation and wastage of the glacier an insulating debris layer did not form to retard melting. Perhaps 5 m of surface melting of the clean ice may have occurred each season, meaning even a 200 m thick glacier would have persisted only 40 years after cessation of ice flow.

Alborn Phase Deposits

Glacial deposits associated with the Alborn phase are placed in the Aitkin formation lithostratigraphic unit (Johnson and others, in press). Two members have been formally defined: the Alborn member, and the Prairie Lake member (Baker, 1964). The Alborn member is the lower member, and is interpreted to have been derived to a great degree from erosion of underlying Lake Upham I lacustrine sediment. The overlying Prairie Lake member contains a significant component of Paleozoic carbonate and Cretaceous shale lithologies, derived from the Red River Valley and Winnipeg basin. The distribution of the two members is poorly understood within and adjacent to the Upham basin, however the Prairie Lake member is less extensive than the underlying Alborn member. Both members are recognized primarily as till lithofacies, however minor glaciofluvial and glaciolacustrine elements locally occur.

Alborn Member

The Alborn member consists predominantly of reddish-brown to dark reddish-grey, clay to clay loam till (Fig. 2a). The pebble, cobble, and boulder clast content is distinctly lower than underlying Rainy lobe deposits (Knaeble and Hobbs, 2009) (Fig. 2b). Alborn member tills along the northern margin of the Upham basin, adjacent to the Mesabi Range, are distinctly more clay rich than tills along the southern margin of the Upham basin (Winter and others, 1973) (Fig. 2a). This may reflect incorporation of a greater amount of clayey lacustrine sediment by the glacier along a longer flow path across the bed of Upham I.

Alborn till is patchily distributed near the margins of the St. Louis sublobe. The once continuous till sheet deposited on ice-cored topography has been disrupted by subsequent meltout and collapse of the underlying substrate.

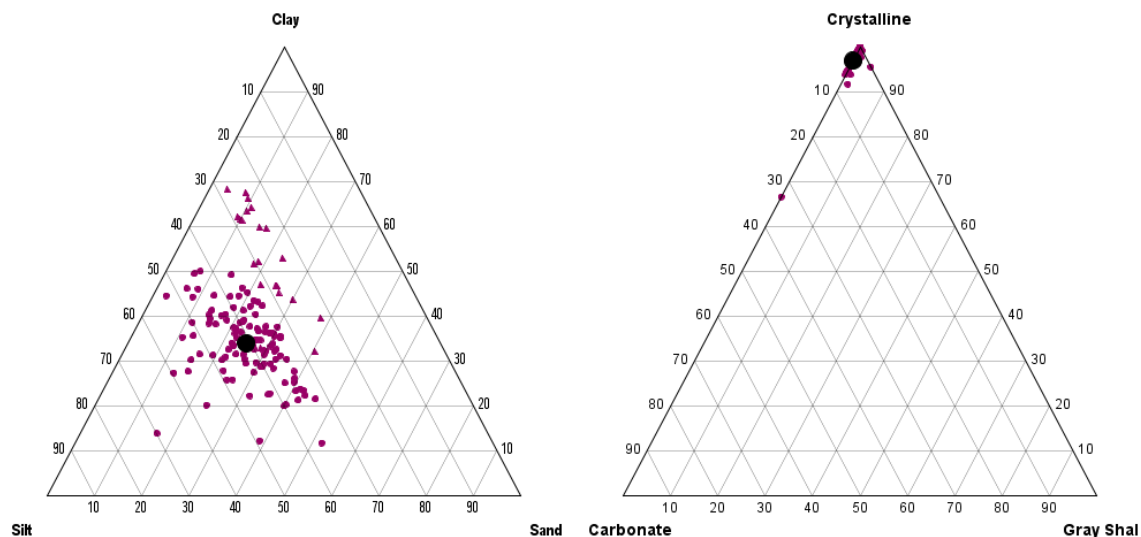


Figure 2. Matrix texture (a. left triangle) and lithologic composition (b. right triangle) of the 1-2 mm sand fraction of Alborn member tills. Circles are from QDI (Quaternary Data Index) database (Minnesota Geological Survey), triangles are ‘red clayey till’ of Winter and others (1973), and large circle is mean in Carlton County (Knaeble and Hobbs, 2009). The ‘red clayey till’ from the northern end of the Upham basin is distinctly more clay-rich than Alborn member tills from the southern end of the basin. The tills contain only a minor amount of carbonate, and no gray shale, reflecting the composition of recycled underlying northeastern provenance (Rainy lobe) drift.

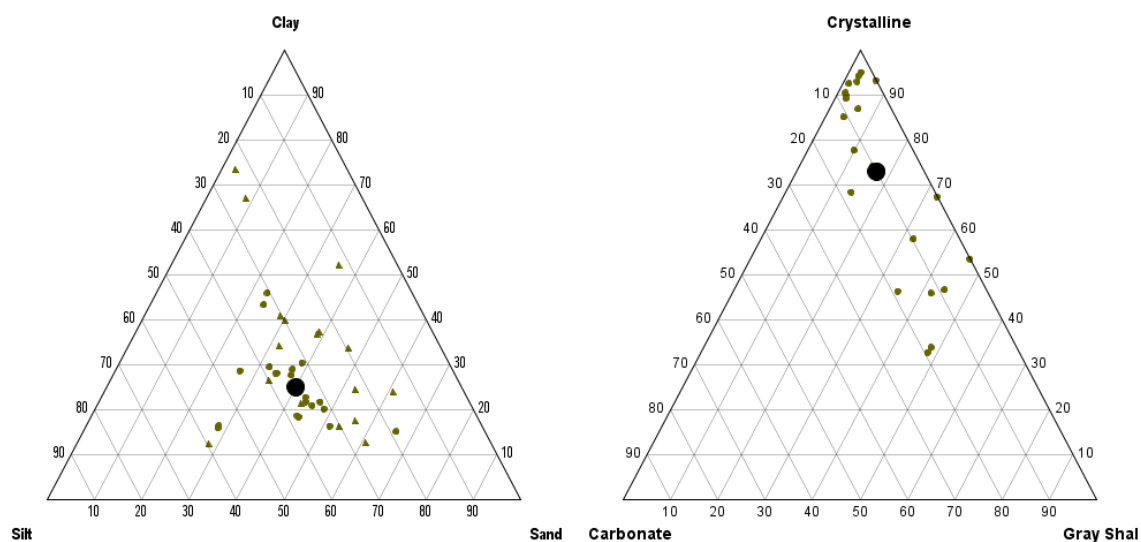


Figure 3. Matrix texture (a. left triangle) and lithologic composition (b. right triangle) of the 1-2 mm sand fraction of Prairie Lake member tills. Circles are from QDI (Quaternary Data Index) database (Minnesota Geological Survey), triangles are ‘brown silty till’ of Winter and others (1973), large circle is mean in Carlton County (Knaeble and Hobbs, 2009). The tills are variably enriched in carbonate and gray shale, reflecting a northwest (Red River Valley) provenance.

Prairie Lake Member

The Prairie Lake member consists predominantly of yellow-brown to brown to dark grey, loam to clay loam till (Fig. 3a). The pebble, cobble, and boulder clast content is similar to Alborn member till, and likewise distinctly lower than underlying Rainy lobe deposits. Unleached (grey) Prairie Lake tills average ~10% carbonate and ~17% grey shale clasts in the very coarse sand fraction, indicative of a northwestern provenance (Fig. 3b). Carbonate leaching averages about 1 m depth (Knaeble and Hobbs, 2009). In contrast to Alborn till, Prairie Lake till shows no apparent systematic textural variation across the Upham basin.

Glacial Lake Upham II

Advance and stagnation of the St. Louis sublobe was followed by a rapid melting of the thin, warm ice, as discussed above. Melting of the ice was accompanied by formation of a series of proglacial lakes, which ultimately coalesced to form Lakes Aitkin II and Upham II (Fig. 4).

Initially, Aitkin II and Upham II were contiguous. Opening of successively lower elevation outlets led to drawdown and drainage of both lakes into the proto-St. Louis River. Continued isostatic rebound raised the sill between Aitkin II and Upham II, leading to re-inundation of Aitkin II. Aitkin II ultimately drained with the opening of an outlet into the Mississippi River on the southwestern end of the basin.

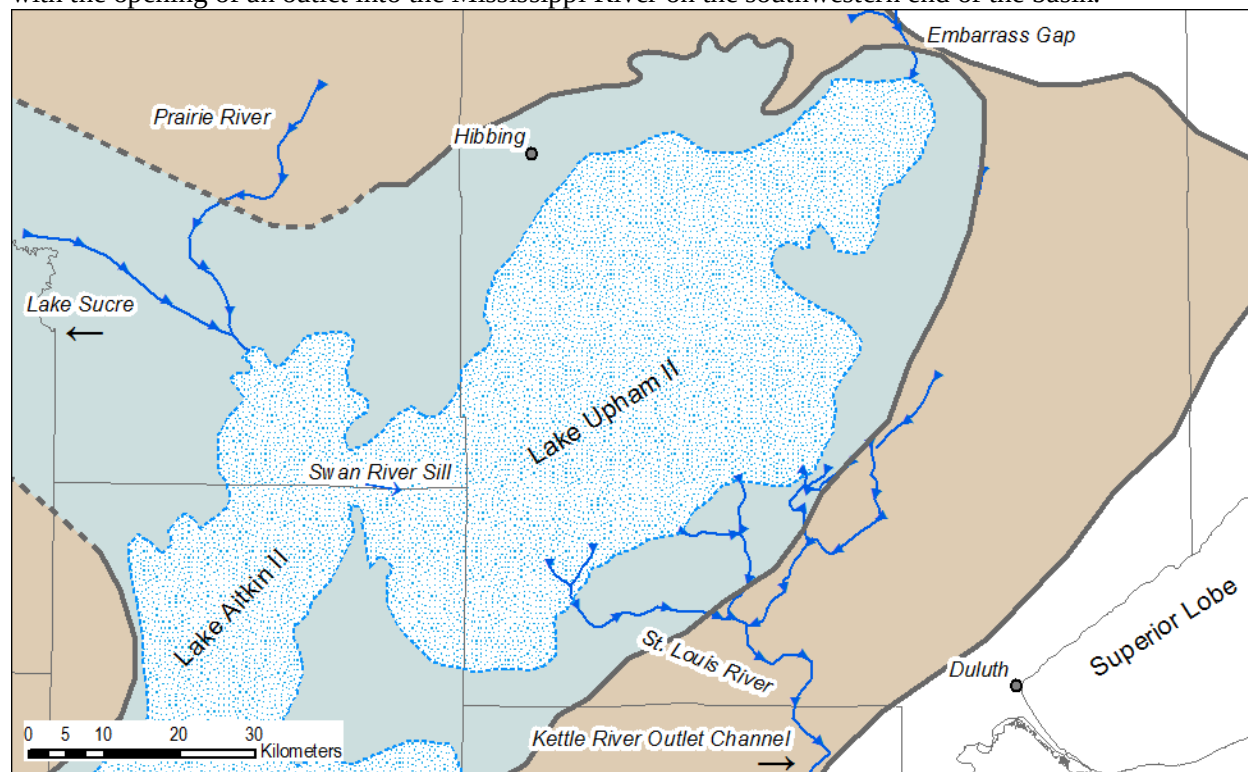


Figure 4. Location of major meltwater inflow and discharge points for Lakes Aitkin II and Upham II, St. Louis sublobe areal footprint (olive), and active Rainy and Superior lobe ice (white) during main phase of Upham II (>ca. 11.6 kyr BP).

As much as 67 m (220') of differential isostatic rebound occurred over the 160 km extending from the southwestern extent of Aitkin II to the northeastern extent of Upham II, based on strandline correlations (Marlow, 2004). Adjusting for isostatic rebound, the absolute elevation difference between the uppermost Upham II outlet and the final outlet is about 38 m (125'), a number also corresponding to the maximum depth of Upham II.

Meltwater Drainage

In addition to their immediate catchments, Aitkin II and Upham II received meltwater from up to 500 km of the margin of the Laurentide Ice Sheet (Fig. 5).

Advance of the St. Louis sublobe into the area recently vacated by the retreating Rainy lobe blocked the flow of meltwater south from the Rainy lobe ice margin. Meltwater apparently pooled in the interlobate area forming proglacial lakes, however discharge necessitated flow either through subaerial channels around the northeastern margin of the glacier, or through sub-, en-, or supra-glacial meltwater conduits in the glacier itself. Morphology of the Goodland esker suggests that a significant amount of meltwater flowed in a supraglacial channel southward across the surface of the glacier, at least during the earlier phase of the advance. Later meltwater flow may have been accommodated at least in part by subaerial channels along the northeastern margin of the glacier.

Following stagnation and melting of the glacier, and formation of Lakes Aitkin II and Upham II, meltwater entered the Aitkin II and Upham II basins at three major inlets. Retreat of the Rainy lobe north of the Laurentian divide led to formation of proglacial Lake Norwood (Winchell, 1901). Discharge from Norwood entered northeastern portion of the Upham II basin through the Embarrass Gap (Fig. 5). Norwood's initial outlet had an elevation of 1490', with successive outlets developed at 451 m (1480') and 450 m (1476') elevation (Lehr and Hobbs, 1992). A lower outlet at 443 m (1454') elevation is correlated with the Mizpah phase of Lake Koochiching (Hobbs, 1983; Lehr and Hobbs, 1992). The final outlet through the Gap has an elevation of around 428 m (1405'), however meltwater drainage ceased when the sill between the Pike and Embarrass Rivers (435 m (1427') elevation) was exposed.

Meltwater also flowed down the proto-Prairie River southward into Lake Aitkin II. This outlet has been nominated as a potential discharge point for the lower elevation Gemmell phase of Lake Koochiching, and even as a potential outlet for Lake Climax, an early, high level of Lake Agassiz (Hobbs, 1983). However, the sill over the Laurentian divide into the Prairie River drainage stands at 418 m (1373'), approximately the same as the lower Koochiching outlets in the Embarrass Gap (corrected for rebound), and channel morphology in the vicinity of the sill shows no evidence of high meltwater discharge. The lower reach of the Prairie River does show evidence for significant meltwater discharge. This was likely limited to a short interval after stagnation and melting of the St. Louis sublobe, and before retreat of the Rainy lobe ice margin from the Laurentian divide and expansion of Lake Norwood.

West of Grand Rapids, MN, Lake Sucre formed in the area covered by the 'upstream' expanse of the stagnant St. Louis sublobe (Larson and others, 2004). A considerable amount of meltwater from the Red River lobe flowed into the western end of this lake, eventually reaching the western end of Aitkin II.

At the time of maximum meltwater discharge through Upham II, flow was ultimately channeled around the northern margin of the Nickerson phase Superior lobe into the Kettle River meltwater channel (Wright and Watts, 1969). Assuming an average melt rate of 2.5 m/yr over an ablation zone extending 200 km from the ice margin, average annual discharge through this channel was roughly 8000 m³/s. Discharge was seasonably variable, so maximum discharge rates were considerably higher. Carney (1996) evaluated peak discharge through this channel examining a variety of parameters, concluding that maximum peak discharges of 12000 to 17000 m³/s were reasonable. This number suggests that meltwater discharge through Upham II and the Kettle River channel solely reflected ablation from the ice margins immediately adjacent to the lakes, and does not contain a Lake Climax/Agassiz discharge component.

Significant meltwater discharge through Lakes Aitkin II and Upham II was ultimately restricted to the period between stagnation and melting of the St. Louis sublobe and opening of the Macintosh channel into Lake Climax in the Red River Valley, at the end of the Cass phase (c. 11.6 kyr BP) (Fenton and others, 1983).

High Level Outlets

A series of outlets to high level proto-Upham II proglacial lakes are present along the eastern margin of the Upham II basin, including include the Chicken Creek (three outlets; 1500' to 1465') and Us-Kab-

Wan-Ka River (428 m to 422 m; 1404' to 1386') outlets. Initially, these channels drained the eastern margin of the St. Louis sublobe, and a portion of the southern margin of the Rainy lobe.

Formation of Lake Norwood may have occurred about the time the Us-Kab-Wan-Ka River outlets formed. As Norwood expanded to the west, an increasing discharge of meltwater was channeled along the eastern margin of the St. Louis sublobe. Meltwater flowed over an unstable, ice-cored landscape, resulting in frequent shifts in outlet location, and drops in outlet elevation. The Us-Kab-Wan-Ka outlets were succeeded by a series of outlets formed in the vicinity of Hellwig Creek ranging from 422 m (1386') to 418 m (1370') elevation.

Main Lake Stage

The highest well-developed strandlines in the Upham II basin correlate to a series of outlets between 413 m (1354') and 407 m (1336') elevation at Hellwig Creek. The Hellwig Creek outlets were abandoned as ice-cored topography in the Culver moraine continued to melt and collapse, in favor of lower level outlets in the Artichoke River 404 m (1324') and Spider Creek 396 m (1300') channels. The highest strandlines in Upham II are commonly obscured by collapsed topography, indicating these shorelines formed on topography still underlain by stagnant Rainy lobe ice.

The final outlet to Upham II was established with opening of an outlet down the proto-St. Louis River channel. Terraces correlating with the uppermost outlet had an elevation of about 392 m (1285'). The broad, wide floodplain and terraces corresponding to the upper St. Louis River outlet indicates a significant amount of meltwater continued to flow through Aitkin II and Upham II at the time this channel was established. At least four subsequent stable lower outlets are present, at 390 m (1278'), 386 m (1266'), 381 m (1250'), and 377 m (1238') elevation. In contrast to the higher elevation outlets abruptly abandoned with opening of lower elevation outlets, the St. Louis River outlet records stepwise drops within a single channel to an ultimate elevation of 375 m (1230').

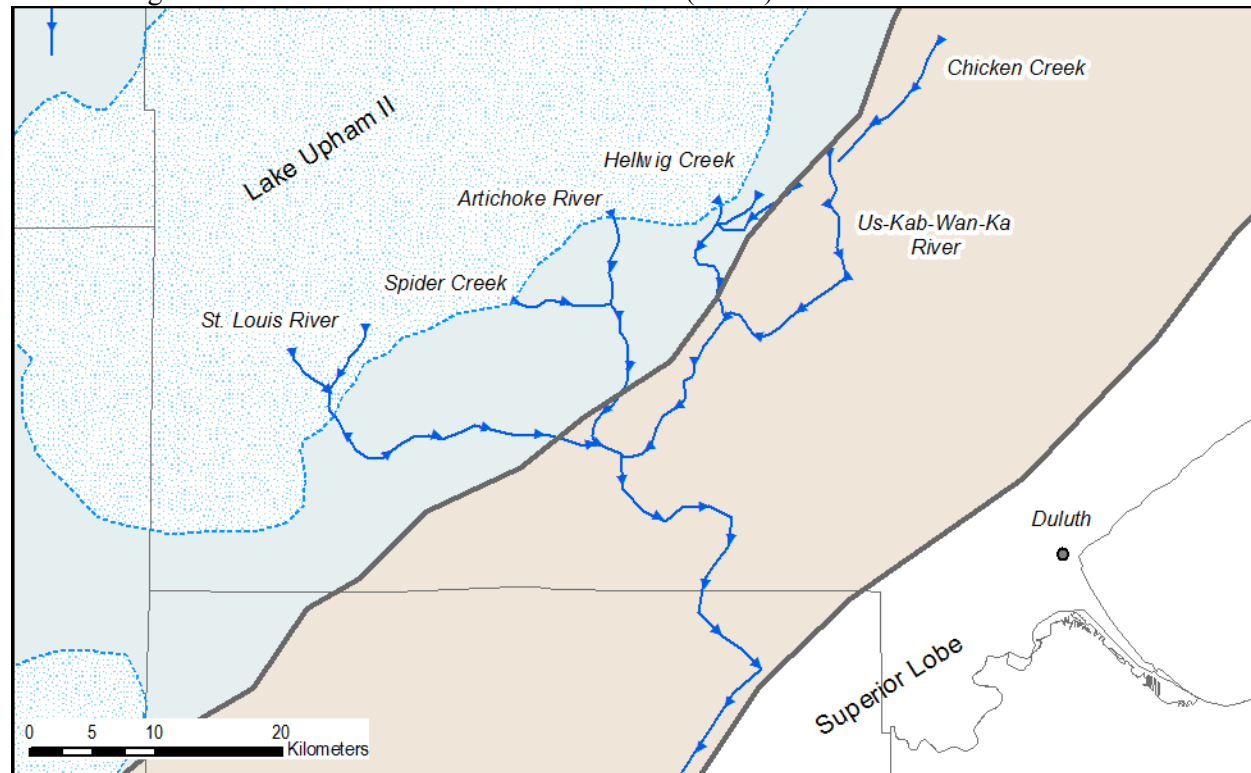


Figure 5. Location of outlets for early ice marginal drainage channels (Chicken, Us-Kab-Wan-Ka, Hellwig) and main stage Lake Upham II outlet channels (Hellwig, Artichoke, Spider, and St. Louis).

Eolian Activity in the Upham II basin

Upham II was apparently stable at the higher Hellwig Creek and Artichoke River outlet levels for a considerable length of time, long enough to allow deposition of well-sorted lacustrine sediments in the lake, ranging from very fine to fine sand in the nearshore areas to clay in the deepest portion of the basin. Later lowering of lake level and exposure of these nearshore sands triggered eolian activity, leading to extensive dune field development (Marlow, 2004).

Final Drainage of Lake Upham II

Separation of Upham II from Aitkin II occurred when the lake level in Upham II dropped below the elevation of the Swan River sill (382 m; 1252'). Outflow occurred through a broad, shallow channel, suggesting meltwater inflow into the Aitkin II basin had ceased. Downcutting of the St. Louis River outlet may have been triggered by cessation of meltwater influx into the lakes.

Discharge over the Swan River sill eroded dune fields developed on Upham II's bed, indicating significant eolian activity does not post-date final cessation of drainage from Aitkin II into Upham II. This relationship further indicates that significant eolian activity in the Upham II basin was largely confined to the interval between opening of the lower Hellwig Creek outlets and the 381 m (1250') elevation St. Louis River outlet.

Final drainage of Upham II occurred as the St. Louis River down-cut to its modern level of about 375 m (1230') at the final outlet to Upham II.

Final Drainage of Lake Aitkin II

Lake Aitkin II substantially drained as the outlet to Upham II dropped to below 382 m (1252'); this drawdown in lake level may have triggered development of the peat dated by Farnham and others (1964) (11.6 ¹⁴C kyr BP). Continued differential rebound between the Swan River sill on the northeastern side, and the southwestern side of the basin led to re-inundation of the lake (and deposition of the upper lacustrine sequence reported by Farnham and others (1964)). Hobbs (1983) reported a radiocarbon date of 9.1 ¹⁴C kyr BP from a snail shell recovered from a marl deposited in Aitkin II; this date indicates Aitkin II persisted for at least 2500 years after separation from Upham II. Aitkin II ultimately drained with opening of a new outlet into the Mississippi River in the southwest corner of the basin at about 366 m (1200') elevation.

DESCRIPTION OF FIELD TRIP STOPS

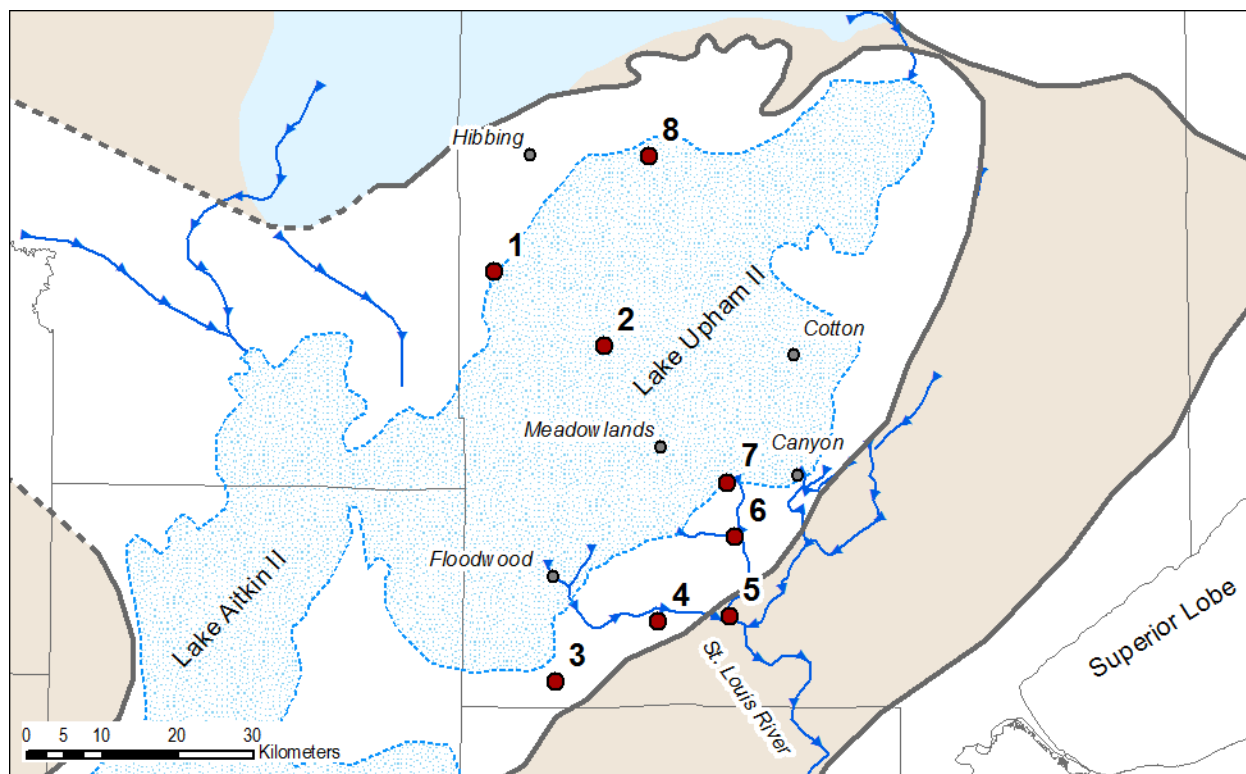


Figure 6. Location of field trip stops relative to major features of the Upham basin.

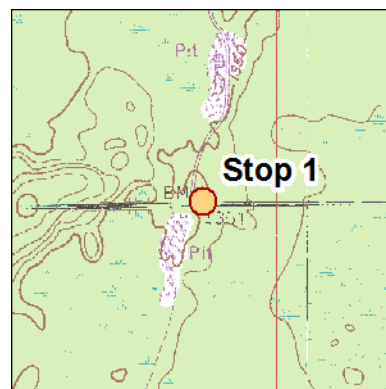
1. Glacial Lake Upham II Beach

499660E/5236650N (UTM Zone 15, NAD83 datum)

Silica 7.5' USGS Quadrangle

NENE, Section 4, T55N, R21W

This site is located on the uppermost relatively well-developed beach associated with Lake Upham II. It likely formed in response to establishment of a relatively stable outlet in the vicinity of Hellwig Creek, on the opposite time of the basin. Subsequent to the time of upper beach formation, Lake Upham II experienced a series of relatively gradual drops in water level at this site. Downward stepping clinoforms visible in a ground penetrating radar profile are interpreted to represent a forced regressive shoreface (Fig. 6). Progradation of the bedforms occurs as a result of shoreline regression, and indicates a constructional shoreline. At this site, regression was triggered by a combination of gradual relative lake level drop due to differential isostatic rebound relative to the more southerly outlets, and relatively abrupt lake level drops triggered by development of new, lower elevation outlets.



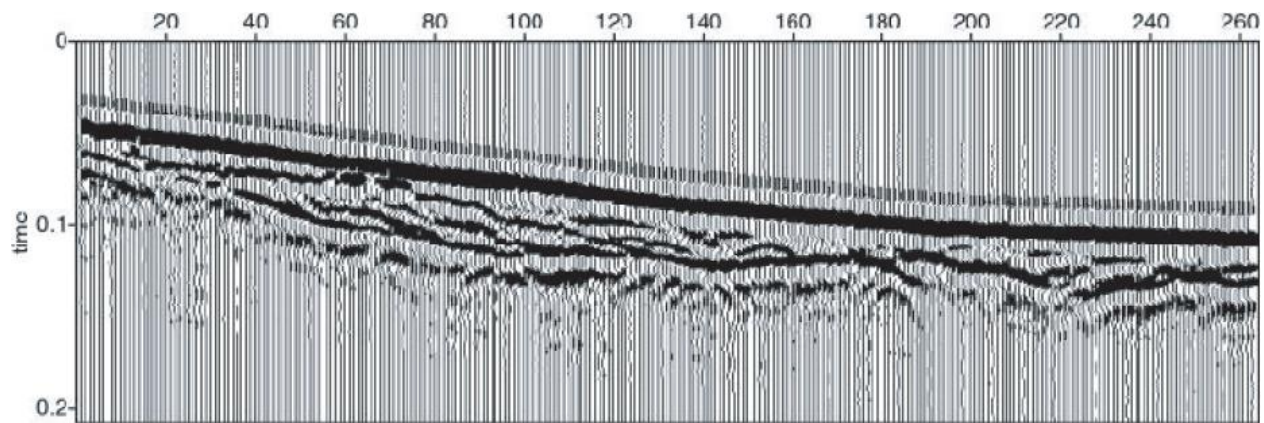
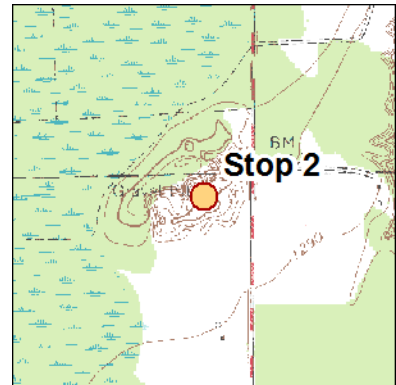


Figure 6. West to east ground penetrating radar profile of Site 1. Profile length 120 m, vertical scale ~9 m. From Knaeble and others (2005).

2. Toivola Esker

514200E/5226890N (UTM Zone 15, NAD83 datum)
Toivola 7.5' USGS Quadrangle
NENE, Section 1, T54N, R20W

Most of the landforms within the Glacial Lake Upham basin predate development of the lakes. This exposure is an example of an esker deposited during retreat of the Rainy lobe that was later overrun by the St. Louis sublobe and subsequently modified by wave action. This esker and others like it became wave-washed "islands" once Glacial Lakes Aitkin and Upham I and II formed. Exposed at the base of the sequence are coarse-grained gravels containing northeast provenance clasts, including granites, iron-formation and locally derived shale and greywacke of the Paleoproterozoic Virginia Formation. This esker segment, and numerous similar examples in the Upham basin, was deposited by a beaded esker system during retreat of the Rainy lobe. Overlying the gravels is a yellow-brown fine-grained till (Prairie Lake Member). Overlying the till is a sequence of nearshore sands and gravels. These presumably eroded from that portion of the esker rising above the level of glacial Lake Upham II; the strandline formed at about 397 meters (1,300 feet) elevation. The uppermost portion of the sequence is a blanket of eolian sand and silt derived from the surrounding lake plain to the upland after final drainage of Lake Upham II.



3. Prairie Lake Member Till

507820E/5182550N (UTM Zone 15, NAD83 datum)
Prairie Lake 7.5' USGS Quadrangle
SWSE, Section 20, T50N, R20W

There are two distinct tills at this road cut on the east side of State Highway 73 on the northeast side of Prairie Lake (Figure 1). Both units are deposits of the St. Louis sublobe. The elevation at the top of the exposure is about 1330 and the upper till, the Prairie Lake Member (Baker, 1964) of the Aitkin Formation (Johnson and others, in press),



is approximately 15 feet thick, yellow-brown (2.5Y5/4) to brown (10YR5/4), calcareous, and loam textured. Three samples have textures averaging 37-35-28 (sand-silt-clay percentages, respectively) and lithologic percentages of the 1-2 mm coarse-grained sand fraction, averaging 46-12-42 (crystalline-carbonate-gray shale, respectively). There are trace amounts (<1%) of red Superior-source sand grains. Five feet of the lower till, the Alborn Member (Baker, 1964) of the Aitkin Formation, is exposed at the base of the ditch. Above the contact between the two tills there are, in places, streaks of red-brown till incorporated into the base of the yellow-brown till. An auger boring in the ditch at the base of the outcrop penetrated another 30 feet. The upper 29 feet detected calcareous red-brown clay loam till. Six samples of this till have textures averaging 23-37-40 (sand-silt-clay percentages, respectively) with coarse-grained sand fraction amounts averaging 3% carbonate, no gray shale, and 11% red Superior-source. There are some, but not many pebbles in the till, which tends to become finer with depth, possibly due to incorporation of underlying lake sediment. The last foot was gray clayey silty lake sediment (possibly Glacial Lake Aitkin I).

Previous interpretations suggest that the tills of these two members were deposited by one ice advance (Baker, 1964; Wright, 1972). The Prairie Lake till represents the original yellow-brown and brown characteristics of the sediment in the ice as it advanced into glacial Lakes Aitkin I and Upham I, and the Alborn till depicts the incorporation and mixing of red lake sediments of glacial Lakes Aitkin I and Upham I into the basal portion of the ice as the glacier advanced across the basin. This produced ice deposits with brownish sediment overlying and/or intermixed with red sediments. In contrast, subsequent interpretations suggest that each member was a separate ice advance (Knaeble and Hobbs, 2009).

4. Alborn Member Till

521450E/5190450N (UTM Zone 15, NAD83 datum)

Martin Lake 7.5' USGS Quadrangle

NWNW, Section 35, T51N, R19W

At this private pit located just south of St. Louis CR 856 there are three tills exposed along the west wall (Figure 2). The elevation at the top of the exposure is about 1350. The upper 8 feet is composed of red-brown (5YR4/3 to 7.5YR4/3), non-calcareous, clay loam Alborn Member till with some pebbles. A sample at 6 foot has a textural analysis result of 22-44-34 (sand-silt-clay percentage, respectively), with no carbonate (leached) or gray shale, and ~10% red Superior-source in the coarse-grained sand fraction. Below a sharp contact there is 3 feet of red-brown (5YR4/3) sandy loam Cromwell Formation (Wright, 1972; Johnson and others, in press) till of the Automba phase of the Superior lobe. Textural and lithologic analysis results for two samples of this till average 39-47-14 (sand-silt-clay percentages, respectively) with 1% carbonate (one sample was leached), no gray shale, and 21% red Superior-source in the 1-2 mm coarse-grained sand fraction. Below another sharp contact is 2 feet of brown (10YR6/3) cobbly, sandy Independence Formation (Johnson and others, in press) till, a deposit of the Rainy lobe. Two samples of this till averaged 53-40-7 (sand-silt-clay percentage, respectively) with trace amounts of carbonate, no gray shale, and 17% red Superior source in the coarse-grained sand fraction. The pebble concentration basically doubled in each underlying till unit. Underlying the 3 till units at the base of the exposure there is pebbly, cobbly sand and gravel.



This site is about a mile or two north of the southern extent of St. Louis sublobe ice deposits. Here Alborn Member till thinly covers older Superior and Rainy lobe deposits.

This same stratigraphic sequence is evident in other pit exposures as far east as Brookston. The Toimi drumlins (Wright and Ruhe, 1965) east of Brookston and the St. Louis River are surface exposures of the Rainy lobe deposits that at this site were covered, first by Automba phase deposits of the Superior lobe and later by the Alborn Member deposits of the St. Louis sublobe.

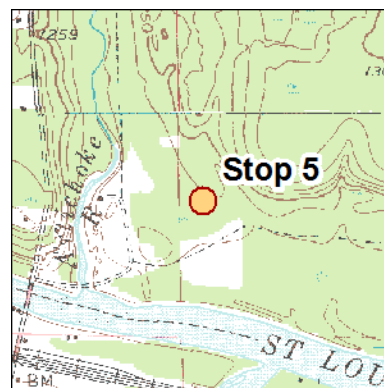
5. St. Louis River Outlet Channel

530800E/5191070N (UTM Zone 15, NAD83 datum)

Brookston 7.5' USGS Quadrangle

SWSW, Section 26, T51N, R18W

This stop is located at the intersection of the Artichoke and St. Louis Rivers, where there is a prominent terrace at 375 meters (1,230') elevation that extends 1 kilometer (0.6 mile) across. The terrace is predominantly composed of a thick deposit of relatively coarse-grained gravel, ultimately derived from erosion of Rainy lobe drift in the Culver moraine. The terrace gravels are overlain and partially infilled by loess. The loess likely originated from lacustrine sediment from the bed of the Lake Upham II, eroded by wind as the littoral zone was episodically exposed by rapid drawdown associated with establishment of new, lower elevation outlets.



6. Spider Creek Outlet Channel

531600E/5201600N (UTM Zone 15, NAD83 datum)

Alborn 7.5' USGS Quadrangle

NWNE, Section 26, T52N, R18W

Spider Creek occupies one of a series of around 10 successive outlets that drained Lakes Upham II, and by extension Aitkin II. The broad (800 m wide), flat-bottomed channel formed when collapse of underlying ice-cored Rainy lobe drift opened an outlet some 24' in elevation below the Artichoke River outlet. The channel morphology indicates a significant amount of meltwater was still discharging out of Upham II.



Baker (1965) reported a bulk radiocarbon date of $13,000 \pm 400$ ^{14}C yr bp from a sequence of lacustrine marl (sample W-1234) within the Spider Creek outlet. The marl must post-date the cessation of drainage through the channel because marl formation requires shallow, still water. Baker (1965) expressed concern that this date was too old due to possible contamination by lignite. However, this date is consistent with the other evidence presented in this field trip description. The Spider Creek date places the minimum age of Glacial Lakes Aitkin and Upham II, and therefore the maximum limit of the St. Louis sublobe, prior to 13.0 ^{14}C kyr B.P.

7. Birch Esker

530580E/5208810N (UTM Zone 15, NAD83 datum)

Payne 7.5' USGS Quadrangle

NESE, Section 34, T53N, R18W

There are multiple exposures in this large county pit (Figure 3). The eastern most exposure is a 25 foot cut adjacent to the railroad crossing revealing 3 separate tills. At a surface elevation of approximately 1350 the soil has been stripped from a 1 foot thick layer of leached mixed till and sand lenses. Underlying the leached layer is 5 feet of yellow-brown (2.5Y6/3 to 10YR6/4), calcareous, silt loam textured Prairie Lake Member till. Near the base of the unit there are shear bands (streaks, pods, and lenses) of incorporated material from the underlying red-brown till. A sample at a depth of 5 feet had a texture of 28-56-16 (sand-silt-clay percentages, respectively), and 8% carbonate, 2 % gray shale, and 1% red Superior-source in the 1-2 mm coarse-grained sand fraction. The underlying red-brown (7.5YR5/4 to 5YR4/4), slightly calcareous, loam textured Alborn Member till is about 3 feet thick with more pebbles and cobbles than the overlying till. There is a cobble-boulder stone line or lag at the base of the unit. A sample at the depth of 8 feet had a texture of 40-40-20 (sand-silt-clay percentages, respectively), and 5% carbonate, no shale, and 9% red Superior-source in the 1-2 mm coarse-grained sand fraction. The lowest unit is about 9 feet thick and exposes brown (10YR6/3) to gray-brown (10YR6/2) non-calcareous, sandy Independence Formation till with abundant pebbles and cobbles. A sample at the depth of 15 feet had a texture of 55-37-8 (sand-silt-clay percentages, respectively), and 1% carbonate, no gray shale, and 15% red Superior-source in the 1-2 mm coarse-grained sand fraction. There is about 10 feet of slump to the pit floor below these units.

At this site deposits of both members of the St. Louis sublobe are present in typical stratigraphic position. At Stop 2 the Alborn Member was above both the Cromwell Formation Automba phase till and the Independence Formation till. Here at Stop 3 we are further northeast beyond the depositional extent of the Automba phase deposits and therefore have only the Independence Formation till at the base.



8. Alborn Member Till on Collapsed Rainy Lobe Ice-cored Topography

520160E/5251950N (UTM Zone 15, NAD83 datum)

Kirk 7.5' USGS Quadrangle

NESE, Section 15, T57N, R19W

The Aitkin and Upham basins were occupied by glacial lakes on two separate occasions during the Late Wisconsin glaciation. Retreat of the Rainy lobe from the Mille Lacs and Outing moraines (Mooers 1988) led to the formation of Lakes Aitkin I and Upham I. The extent and timing of these lakes is poorly understood, as their presence is largely inferred from incorporation of lacustrine sediment into the overlying St. Louis sublobe till.

Exposed at the base of the sequence are steeply south-dipping foreset beds of a subaqueously deposited fan. These sediments are Rainy lobe provenance, deposited along the southern margin of stagnant ice lying on the southern slope of the Giant's Range, an area now characterized by collapsed ice-cored topography. The upper portion of the sequence is a St. Louis sublobe till. Between the fan sediments and till are a number of elongate slabs of fine-grained lacustrine sediment derived from Glacial Lake Upham I. This lacustrine sediment was eroded from deeper water and thrust onto the fan during the advance of the St. Louis sublobe (Figs. 11 and 12). The relationships visible in this exposure indicate that the St. Louis sublobe advance occurred while a substantial amount of debris-mantled, stagnant Rainy lobe ice was still present south of the Giant's Range.



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FIELD TRIP 7

Saturday, May 17, 2014

GEOLOGY AND GOLD MINERALIZATION OF THE VIRGINIA HORN AREA

LEADERS:

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William Rowell and Richard Sandri (Vermillion Gold LLC), and
Jason Richter (Minnesota Department of Transportation)*

INTRODUCTION

The local term “Virginia horn” applies to an area near the town of Virginia, where the generally east-trending, Paleoproterozoic, Biwabik Iron Formation makes an abrupt bend to the southwest, creating a marked anomaly in the map pattern (Fig. 1). The iron-formation unconformably overlies well-exposed Neoarchean bedrock within an uplifted, wedge-shaped block. This trip visits exposures that provide an overview of the Archean metavolcanic, meta-igneous, and metasedimentary rocks—including a Timiskaming-type successor-basin sequence, and Paleoproterozoic iron-formation and associated strata. Archean quartz-feldspar porphyry intrusions were the locus of deformation, alteration, and associated gold mineralization. The area has a long history of intermittent gold prospecting, and the potential for mineable quantities is currently under investigation by Vermillion Gold, LLC. Channel sampling and drill core from that exploration work will be displayed. Some of the most recently acquired core in the area is a product of highway relocation work underway to accommodate proposed new iron mining.

The following field guide is modified from Jirsa and Green (2011). The stops are ordered for expeditious travel, rather than stratigraphy or geochronology. All UTM coordinates are given using NAD 83, Zone 15N. It is likely that more stops are described here than can reasonably be covered in a single day. Stops missed during this excursion can be visited by individuals using the guide, with the caveat that

some stops may require permission from land owners.

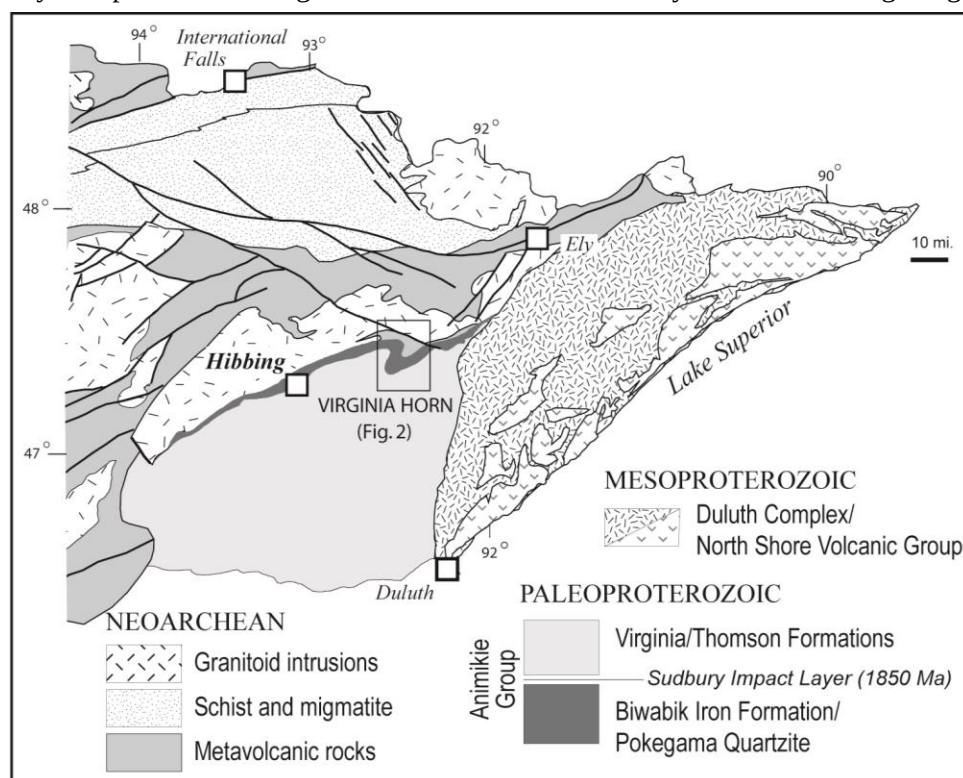


Figure 1.
Generalized geologic
map of northeastern
Minnesota showing
location of the
Virginia Horn area
(black outline).

Geology of Archean Rocks (STOPS 1-5, 9, 10)

The Archean rocks in the Virginia horn area (Fig. 2) are part of the Wawa subprovince of Superior Province, and are similar in most respects to other greenstone-granite terranes of the subprovince. The supracrustal rocks in the horn are separated from the well-known Vermilion district to the north by the Giants Range Batholith—a large, composite body consisting of several intrusive generations and compositions. The mafic volcanic and hypabyssal intrusive components in the Virginia horn may be equivalent to the Ely Greenstone in the Vermilion district, which has a $^{207}\text{Pb}/^{206}\text{Pb}$ zircon age of associated felsic strata of 2722.6 ± 0.9 Ma (Peterson and others, 2001). The supracrustal rocks in the horn are subdivided into northern and southern panels on the basis of metamorphic grade and deformation style. The northern panel, immediately south of the Giants Range batholith, contains intensely lineated, amphibolite-grade schist having volcanic, intrusive, and clastic protoliths (Minntac sequence). The southern panel contains lithologically and stratigraphically similar rocks that were metamorphosed to much lower grades, ranging from prehnite-pumpellyite to low greenschist (Mud Lake sequence). The two panels are separated by the east-trending, post-metamorphic, Laurentian fault. The two sequences are stratigraphically, lithologically, and geochemically identical; suggesting that they may represent different crustal exposure levels of the same stratigraphic package.

The metamorphic cleavage-forming event in both panels was the second (D_2) of three regional scale deformation events—no metamorphic effects are recognized from the other two deformations. The first (D_1) involved upright folding, soft-sediment deformation, and complex faulting. Strata of the southern panel form the broad, southwest-plunging, Mud Lake syncline (Fig. 2.B) and many smaller sympathetic folds—all inferred to be D_1 structures. The syncline is cored by graywacke, slate, and minor felsic tuff, and has outer limbs of calc-alkalic and tholeiitic volcanic strata. The Mud Lake strata and D_1 structures (F_1 fold axes) developed in it were cut by felsic quartz- and feldspar-phyric (QFP) intrusions. However, the QFP intrusions are lithologically similar to some layers within graywacke, suggesting the possibility of temporal overlap between dacitic magmatism and sedimentation. Strata in the Mud Lake syncline, and the quartzofeldspathic dikes that intrude them, are bisected by a fault- and unconformity-bounded, alluvial fan-fluvial-volcanic succession known as the Midway sequence (see discussion below). All three sequences described above were metamorphosed and deformed during D_2 , which has been bracketed locally between about 2674 Ma and 2682 Ma (Boerboom and Zartman, 1993). The third deformation event (D_3) produced localized semi-brittle crenulation of D_1 and D_2 structures, and selective reactivation of earlier-formed faults. Based on seismic and geochronologic work in adjacent Ontario (e.g., Percival and Helmstaedt, 2006), D_1 may be equated with the Shebandowanian orogeny at about 2695 Ma. It may represent collision of the Wawa subprovince with the composite Superior superterrane to the north. The second deformation (D_2) occurred at about 2680 Ma during the Minnesotan orogeny. It can be attributed to oblique collision of the Minnesota River Valley subprovince with the Superior superterrane along a north-dipping suture known as the Great Lakes Tectonic Zone in south-central MN (Percival and others, 2006). The three major deformation events are more or less coaxial, reflecting a continuum of roughly NW-SE-directed compression. The D_3 event appears to have been a late manifestation of this, perhaps at a time when rocks were at sufficiently high crustal levels to elicit mainly partitioned brittle responses.

The Midway sequence—a Timiskaming-type assemblage (STOPS 2, 5)

The Midway sequence forms a wedge of strata <500 meters thick that youngs consistently southward. The basal (NW) contact is not exposed, but intersections in 10 drill holes indicate the contact is a fault in some localities, and an unconformity in others. The latter is indicated in drill core by the presence of sand-filled fractures in the subjacent Viking porphyry intrusion, and the abundance of QFP clasts in the “overlying” Midways sequence. The Midway sequence contains attributes of Timiskaming-type successions, including temporal and geographic association with large fault systems, alkalic and calc-alkalic volcanic and intrusive rocks (hornblende trachyandesite), and conglomerate containing clasts of trachyandesite, together with those derived from older plutonic and strongly foliated substrate bedrock

(Jirsa and Boerboom, 2003). The origin of such Timiskaming-type assemblages has been variously ascribed to localized extension in regional transpressional regimes (Jirsa, 2000; Corcoran and Mueller, 2007), or regional extension in response to imbrication and crustal loading during terrane accretion (Bleeker, 2012). Isolated occurrences of sequences similar to the Midway are exposed in other parts of Minnesota—near International Falls (Seine Group), in the Vermilion district (Gafvert Lake sequence), and in the Knife Lake area (Ogishkemuncie conglomerate). The ages for deposition of the Seine Group are 2693 ± 1 to 2692 ± 1 Ma (Fralick and Davis, 1999); the Gafvert Lake sequence is 2689.7 ± 0.8 Ma (Lodge and others, 2013); and the Ogishkemuncie conglomerate contains clasts of 2690.83 Ma Saganaga Tonalite (Driese and others, 2011). The broad equivalence of these sequences, and the fact that most young southward, indicates more or less synchronous development, which is consistent with an origin involving a single event of regional extension.



ROTTEN TO THE CORE

Filler—taken from an old GAC GEOLOG publication

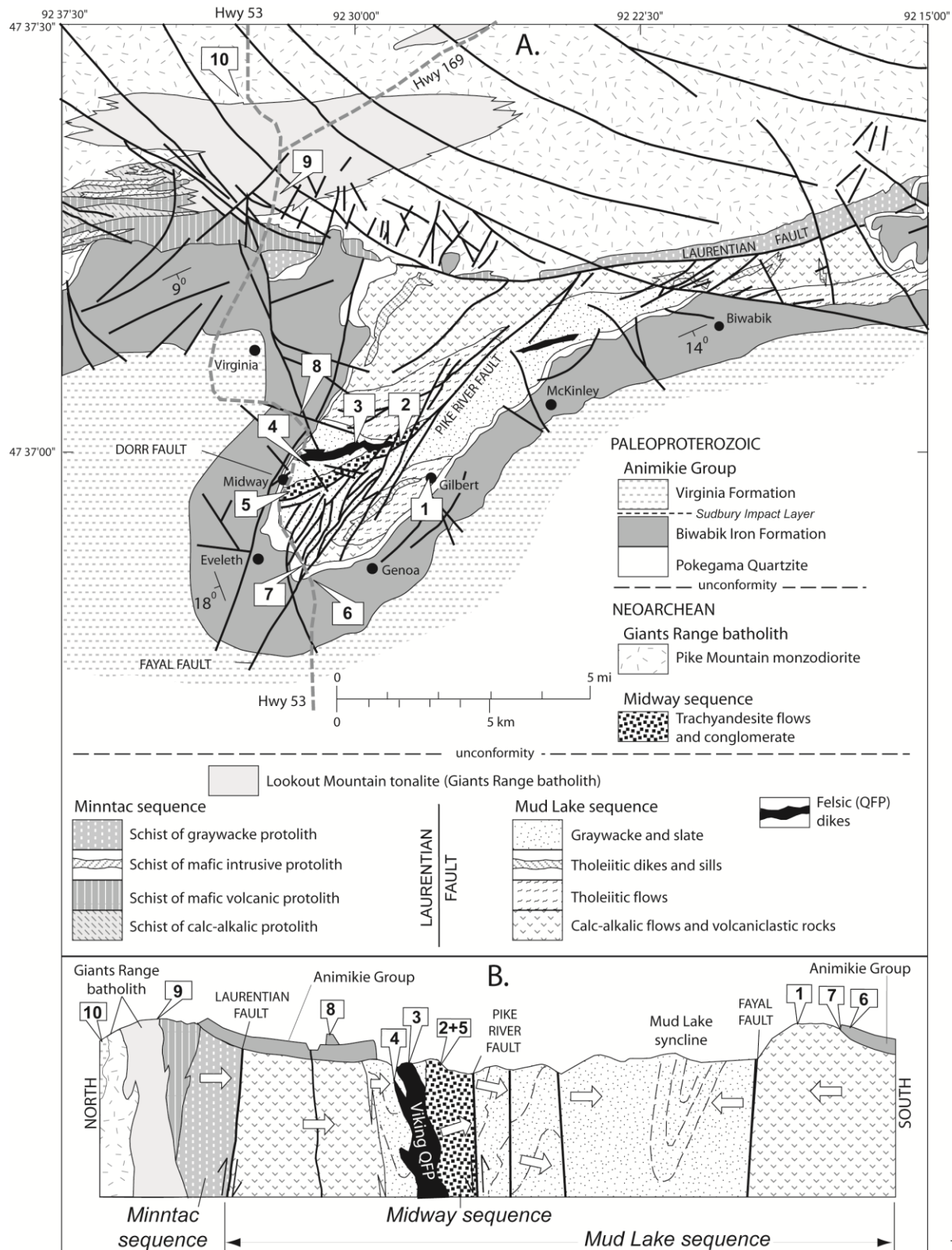


Figure 2. Geologic map (A.) and schematic cross-section (B.) of the Virginia Horn area (modified from Jirsa and others, 1998) showing details of field trip STOPS 1 to 10. Block arrows on B indicate directions of stratigraphic facing.

Gold Mineralization (STOP 3)

In the 1930s visible gold was discovered by Minnesota Geological Survey geologist J.W. Gruner (Grout, 1937) in Archean rocks adjacent to a railroad grade cutting through the Virginia Horn (Fig. 2A and STOP 3). Perhaps because of the regional emphasis on iron ore mining, there was no systematic exploration for gold in the Virginia Horn until the 1980s. During the 1980s Newmont Mining, Rhude and Fryberger, Resources Limited, and American Shield conducted exploration programs that included 43 drill holes totaling 20,000 feet, geologic mapping, soil and outcrop geochemical surveys, and ground geophysical surveys. Most of the 1980s exploration focused on well exposed knobs of what is known locally as the Viking quartz-feldspar porphyry (QFP), which strikes west-southwest from the western side of the Pike River fault (Fig. 2B). Within the QFP, gold is concentrated in zones of variable brittle-ductile deformation (likely both D_2 and D_3), with associated carbonate-sericite alteration and abundant quartz veins. Proximal to the Pike River fault, higher grade gold mineralization occurs in one to three cm-thick quartz veins with pyrite, arsenopyrite and free gold, and in greyish quartz flooded zones with acicular arsenopyrite but no visible gold. One kilometer to the west of the Pike River Fault, most of the known gold mineralization is associated with the Viking QFP and variably sericitized porphyritic dacite that rarely outcrops. Gold is predominantly concentrated in one to three cm anastomosing quartz veins with pyrite and arsenopyrite concentrated along vein margins. Arsenopyrite occurs in irregular masses and not in the acicular habit associated with gold mineralization one km to the east.

Recent studies of the geology of the Virginia Horn area by the Minnesota Geological Survey have shown that the Virginia Horn prospect is located within a Timiskaming-type geologic setting with good potential for gold mineralization beyond the porphyry and into adjacent metasedimentary and metavolcanic rocks (Jirsa and Boerboom, 2003; Bleeker, 2012). Subsequent analyses of samples from outcrop and drill core have confirmed that these rock types are also gold-enriched.

Geologic Setting of Paleoproterozoic rocks (STOPS 6-8)

The Paleoproterozoic strata exposed in the Virginia Horn are part of the Animikie Group, a sequence of sedimentary rocks, including basal quartzite and siltstone (Pokegama Quartzite), medial iron-bearing strata (Biwabik Iron Formation), and upper graywacke and shale of turbidite origin (Virginia Formation). Current models indicate deposition in a back-arc basin that evolved into a northward-migrating fore-deep during the compressional phase of the Penokean orogen—largely complete by about 1850 Ma (Schulz and Cannon, 2007; Pufahl and others, 2010). A depositional age for iron-formation can be inferred from interbedded volcanic tuff in the equivalent Gunflint Iron Formation to the northeast, which produced a U-Pb zircon date of approximately 1878 Ma (Fralick and others, 2002). The contact between iron-formation and overlying slate of the Virginia Formation is marked by the presence of breccia and ejecta formed during the 1850 Ma Sudbury meteorite impact event. The ejecta contain abundant petrographic evidence of impact origin, including the presence of zoned spherules and quartz fragments displaying multiple planar deformation features. The impact-related horizon, known as the Sudbury Impact Layer, is well exposed in the Gunflint Lake area of northeast Minnesota (Jirsa and others, 2011), in the Thunder Bay area of adjacent Ontario (Addison and others, 2005), and in Michigan (Cannon and others, 2010; Pufahl and other, 2007); however, it can be seen only in drill core on the Mesabi range. Tuffaceous layers in basal strata of Virginia Formation were sampled a few meters above the Sudbury Impact Layer and produced an age of 1832 ± 3 Ma (Addison and others, 2005).

The belt of exposure forming the Mesabi Iron Range defines a regional monocline striking ENE and dipping shallowly (0-12 degrees) southward. The exception to this trend is in the Virginia Horn, where strike varies from N-S to NE, and dips as great as 25 degrees are recorded. This paired syncline-anticline is inferred to be related to uplift of a horst, now manifest in the Archean core of the structure, which formed by a combination of folding and faulting (Morey, 2003). Offset along the Laurentian fault, which was south-side down after the D_2 Archean metamorphic and deformation event, was subsequently reactivated during the Paleoproterozoic to produce north-side down movement during and after deposition of the Animikie Group.

DESCRIPTION OF FIELD TRIP STOPS

STOP 1

Archean pillowed and massive greenstone—Old Gilbert school

Location: UTM (NAD 83, Zone 15): 539,820E/5,259,750N; north edge of athletic fields for former Gilbert Junior High School off Wisconsin Avenue, 4 blocks northwest of State Highway 37.

Description:

This outcrop of pillowed and massive basalt is part of the Archean Mud Lake sequence, metamorphosed to low greenschist-grade. Pillow shapes indicate stratigraphic facing is to the northwest, consistent with the location of this outcrop on the south side of a major D_1 structure known as the Mud Lake syncline. Note also the presence locally of fractures and shallow depressions on the outcrop surface that are filled with reddish jasper, presumably deposited by overstepping of Paleoproterozoic seas onto the eroded surface of Archean bedrock during deposition of the Biwabik Iron Formation.

Discussion:

Detailed structural study by Jirsa and others (1998) and Jirsa and Boerboom (2003b) demonstrate that the tholeiitic and calc-alkalic volcanic rocks and tholeiitic intrusions are conformably overlain by graywacke and slate of the Mud Lake sequence (STOPS 3 and 4). In detail, the Mud Lake sequence forms a broad, twice-deformed, west-plunging syncline that has been segmented by faults of several generations.

STOP 2

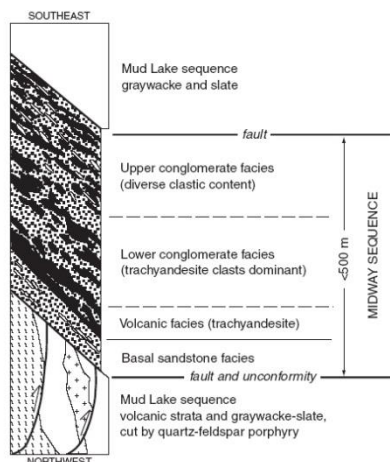
Archean volcanogenic conglomerate of the Midway sequence

Location: UTM: 539120E/5261040N; Old Railroad trail

Description:

This former railroad cut exposes parts of the volcanic and lower conglomerate facies of the Timiskaming-type Midway sequence. Figure 3 shows the position of this exposure within a composite stratigraphic section. The rocks here are characterized by disorganized beds of poorly sorted conglomerate and volcanic breccia. Dark red, green, and purple clasts of hornblende- and plagioclase-phyric trachyandesitic composition are most abundant. Both normal and reversed grading are preserved locally. Clasts are as large as 25cm, and diamictites containing outsized clasts are common in this unit. Flattening of clasts in the plane of D_2 is apparent, and a matrix of varied grain size locally displays anastomosing S_2 cleavage. Overall, the conglomerate contains clasts representing all lithologic components of the Mud Lake sequence and the porphyry dikes that intruded it. However, significant variations in clast content and internal organization of bedding characterize these units, and these attributes vary gradationally both

laterally and with stratigraphic height. Figure 4 shows the map and cross-section distribution of the Midway sequence based on both drill core and surface exposures. The Upper conglomerate facies will be visited at STOP 5.



Discussion:

Remarkably, the polychromatic trachyandesite clasts are identical with those in parts of the Shebandowan assemblage exposed some 240 km, more or less along strike to the NE in Ontario (e.g., Aubut and Campbell, 2012), despite the intervening Giants Range batholith and other terranes.

Figure 3. Composite cross section of the Midway sequence. Dark polygons represent clasts of trachyandesitic to trachybasaltic composition; white polygons represent clasts of quartzofeldspathic porphyry identical with the Viking QFP (STOP 3); which is represented diagrammatically by the dot pattern. (From Jirsa and Boerboom, 2003, Fig. 2.4).

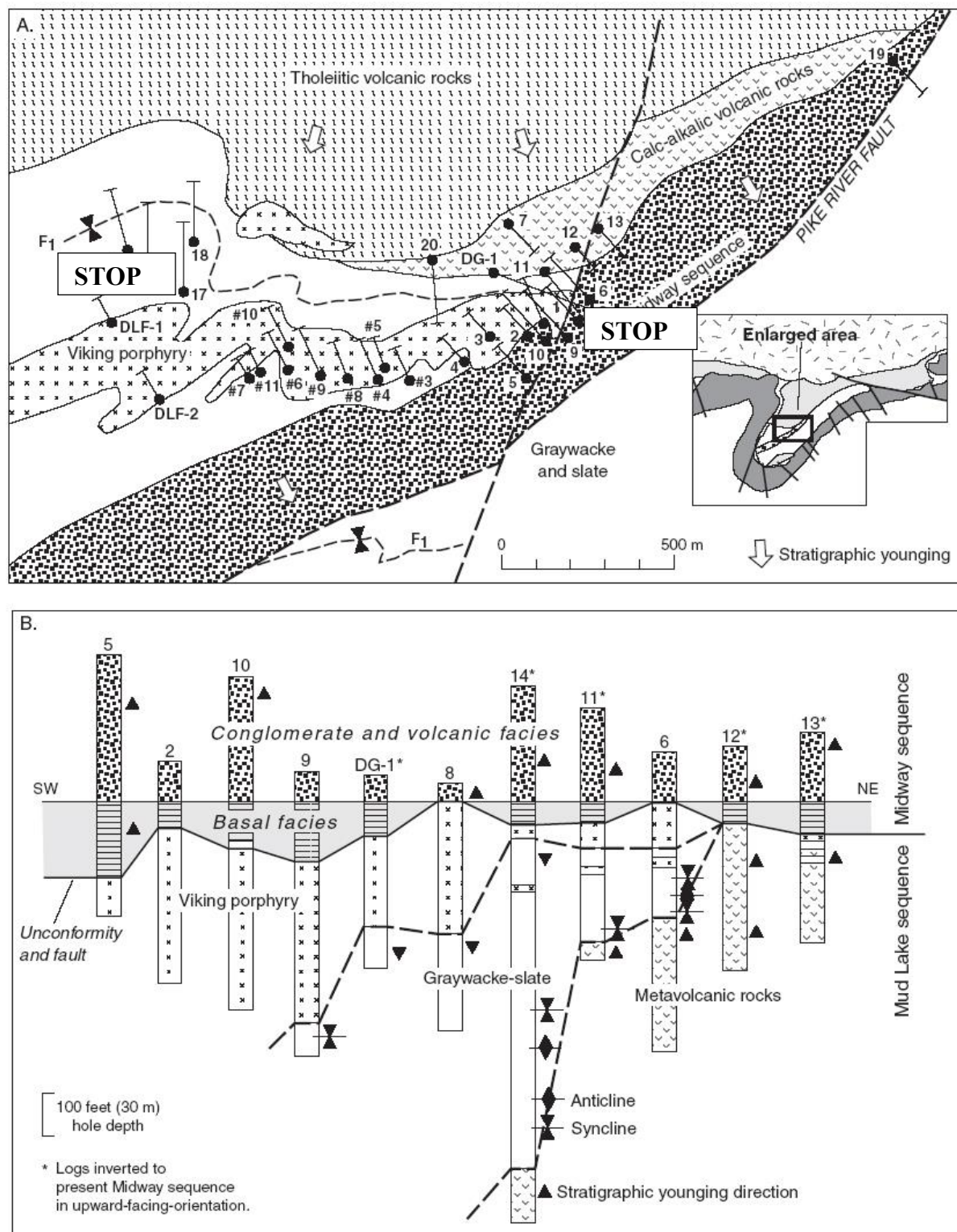


Figure 4. Surface and drill core-based geology of the “Viking Prospect area (STOPS 2 and 3). A) shows geologic map view; B) shows drill holes that intersected the northeastern (basal) contact of Midway sequence with adjacent rocks. From Jirsa and Boerboom, 2003, Fig. 2.6.

STOP 3

Archean Graywacke, argillite, and quartzofeldspathic porphyry with Au mineralization

Location: UTM: 537715E/5261240; Old Railroad trail

Description:

This former railroad cut and outcrops nearby expose interlayered graywacke and argillite of the Mud Lake sequence, intruded locally by quartzofeldspathic porphyry. One of the earliest reports of gold in Minnesota (1930) was made at this locality, and visible gold can still be found associated with small quartz veins. The porphyry intrusions here are identical with and presumably apophosial offshoots of the larger Viking QFP exposed to the east along and south of the trail (Fig. 4A). Regionally, the porphyry contains ornately embayed phenocrysts of quartz as large as 2cm, smaller albite phenocrysts, and minor mica in an aphanitic quartzofeldspathic groundmass. Groundmass is commonly crossed by anastomosing shear planes and altered to combinations of quartz, sericite, dolomite, iron-carbonate (ankerite and ferroan dolomite), pyrite, and locally arsenopyrite. Channel samples on this outcrop (Fig. 5) and core from several nearby exploration drill holes will be displayed and discussed.

Discussion:

During 2009 and 2010, Vermillion Gold, LLC (<http://vermilliongold.com>) completed nine drill holes designed to reevaluate areas where gold mineralization was intersected by the 1980s drill holes, and test new targets outside of the Viking QFP. Five of the nine drill holes have focused on gold mineralization in altered porphyritic dacite that outcrops adjacent to a railroad grade on the western side of the property (STOP 3). A channel sample of intercalated porphyritic dacite and metasediments taken from the outcrop by Newmont in the 1980s averaged 1.2 gpt over a sample length of 77.5 ft. Vermillion Gold's 2009 drill hole (Fig. 6) beneath the railroad grade outcrop intersected 1.1 gpt gold over an interval of 195.7 feet and includes intersections of 16.1 gpt/3.6 ft and 11.4 gpt/4.2 ft. A 2010 drill hole, located approximately 150 m south of the railroad outcrop, intersected a thick unit of porphyritic dacite locally cut by thin fingers of Viking QFP. A 224.9 ft section of the drill hole averages 1.0 gpt gold and includes a 77.9 gpt/2.3 ft sample with visible gold in a quartz vein. Within the thick, gold-enriched intersections, values of 100 to 300 ppb are associated with disseminated pyrite and arsenopyrite. Samples with multi-gram gold values include quartz veins with free gold. There is a strong correlation between gold and arsenic values, however, samples with the highest grade gold values reflect free gold and not increased arsenopyrite content.

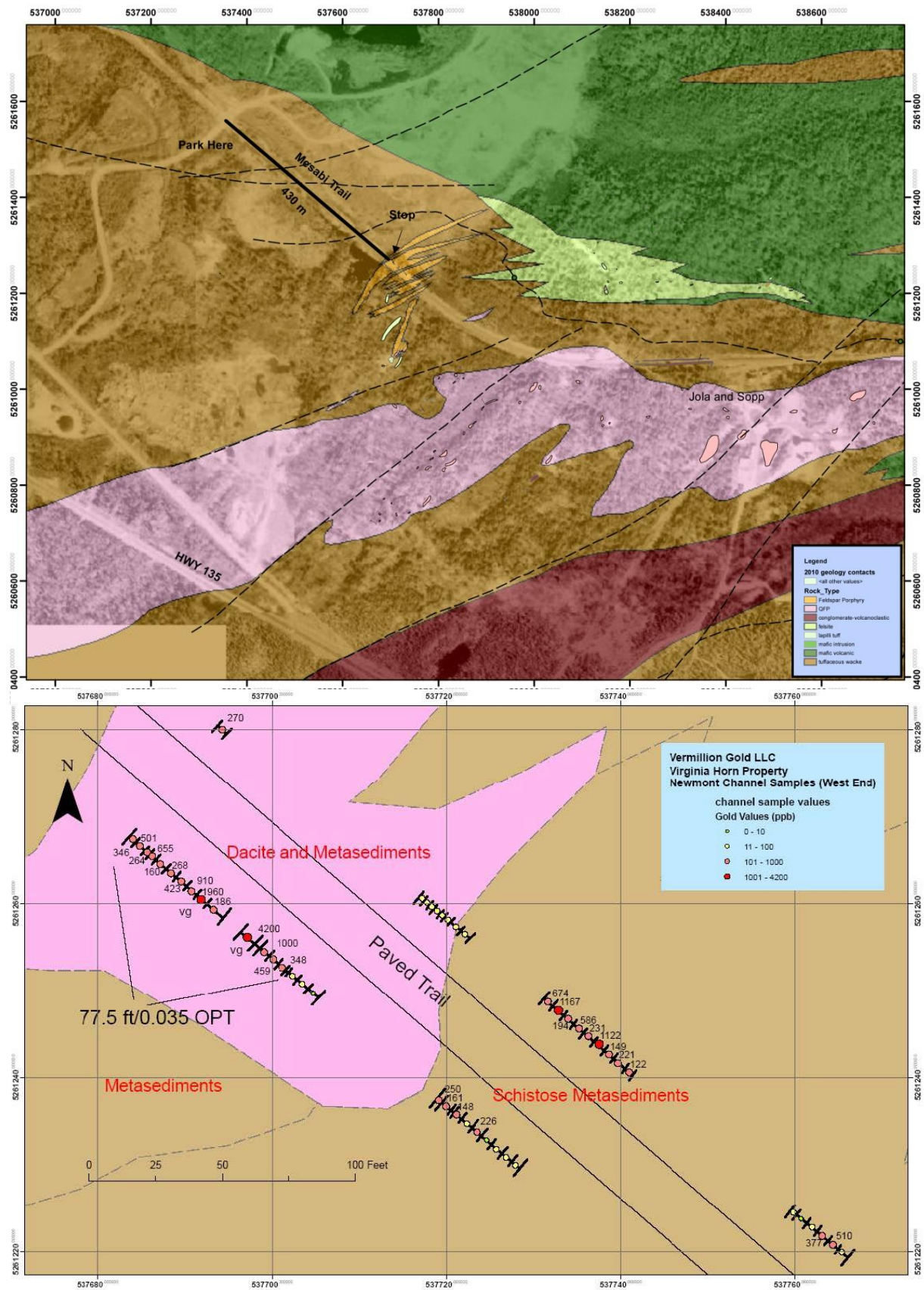


Figure 5. Location (upper image) and analytical results (lower image) of channel samples on outcrops at STOP 3 (imagery from Vermillion Gold, LLC)

Vermillion Gold LLC

Project - Virginia Horn

Drill Hole - VH-09-4

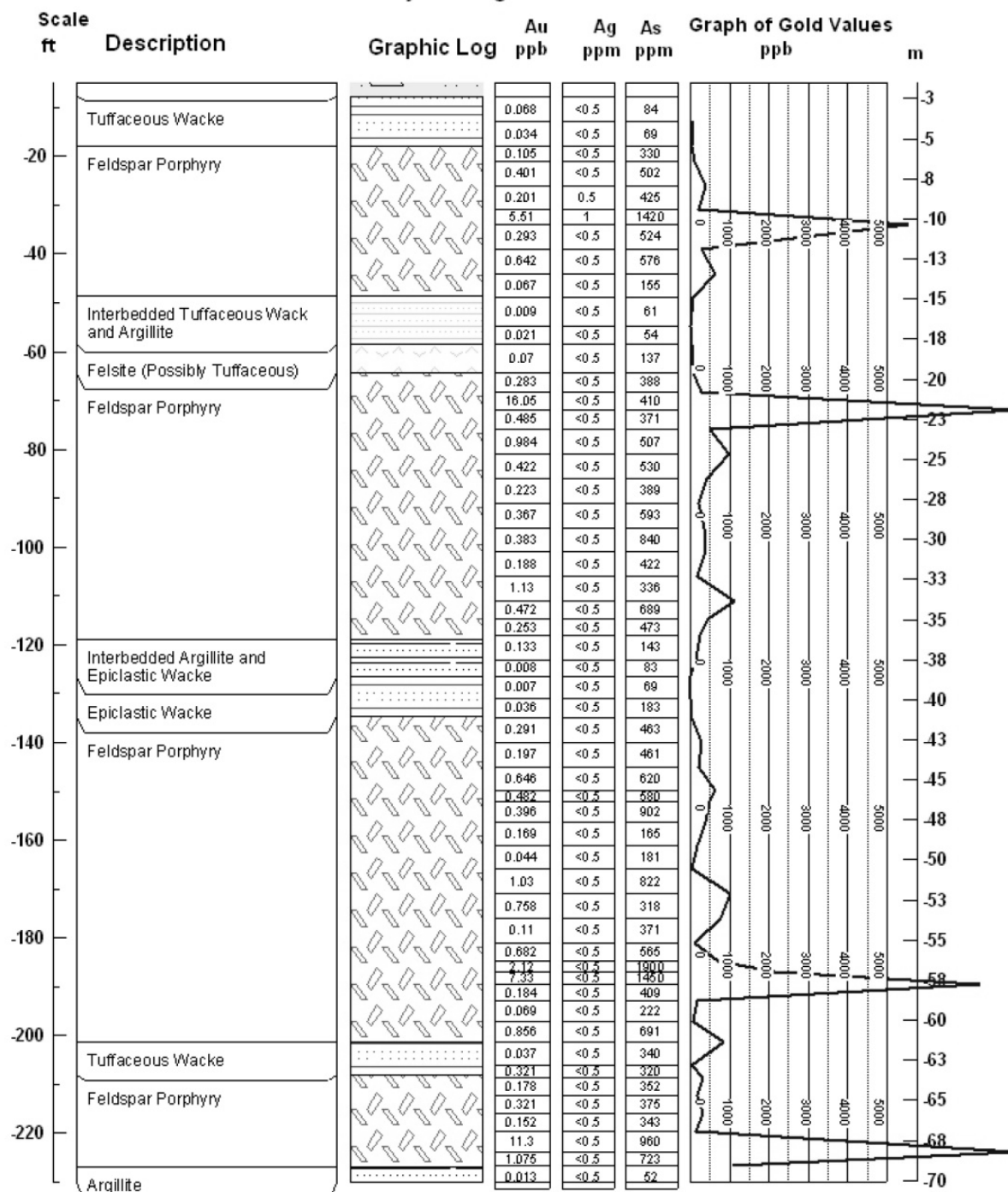


Figure 6. Partial log of core from drill hole VH-09-4, showing lithologic and analytical results (from records of Vermillion Gold, LLC).

STOP 4

Archean graywacke and slate, intruded by quartzofeldspathic porphyry—Bourgin Road

Location: UTM: 536,311E/5,260,659N; road cut on east side of Bourgin Road.

Description:

Outcrops along this side of the road expose quartzofeldspathic porphyry (QFP) intruded into variably deformed graywacke, siltstone, and slate of the Mud Lake sequence. The sedimentary rocks here are moderately deformed, but much of that deformation is inferred to predate the main cleavage-forming event D_2 , and some may have occurred prior to lithification. The QFP is large and continuous to the east, but at this locality, appears to be segmented into a zone of multiple anastomosing dikes. Both graywacke and QFP are intensely altered to some combination of iron-carbonate minerals (ankerite, ferroan dolomite) and sericite. Regionally, this style of alteration is commonly, though not always associated with the QFP intrusions—presumably because they remained more structurally rigid than the enclosing sedimentary rocks during the shear-related deformation event that accompanied alteration late in D_2 . Most gold mineralization in the area is closely allied to this alteration, yet this outcrop is apparently barren.

STOP 5

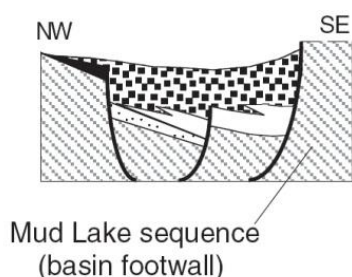
Archean conglomerate—Midway sequence

Location: UTM: 535,713E/5,259,459N; driveway at No. 7 Mesabi Lane, village of Midway.

NOTE: This is private property! Permission must be obtained before entering.

Description:

The Archean conglomerate and lithic sandstone that form this driveway surface represents the Upper conglomerate facies of the Timiskaming-type Midway sequence. It differs from the Lower conglomerate facies at STOP 2 in containing more diverse clast content, more rounded clasts, graded bedding, and more abundant sandy beds. These attributes imply submarine deposition. Taken in the context of the sequence stratigraphy (Fig. 3), this may indicate basin deepening over time, which is consistent with observations of other Timiskaming-type successions (Bleeker, 2012). The conglomerate contains clasts of basalt, graywacke, quartzofeldspathic porphyry (QFP), and porphyritic trachyandesite. This provenance indicates that the older Archean rocks of the Mud Lake sequence were intruded by QFP, deformed, uplifted and eroded to provide detritus to what was probably a “pull-apart” or extensional basin developed along a major structure now occupied by the Pike River fault zone. The southward-younging basin is bounded by both faults and an inferred unconformity (Fig. 7).



Discussion:

Note also the presence of red jasper in depressions and joints as at STOP 1, indicating that this outcrop surface represents paleo-seafloor during deposition of the Paleoproterozoic Biwabik Iron Formation.

Figure 7. Schematic illustration of Midway basin geometry prior to steepening by D_2 compressional deformation (from Jirsa and Boerboom, 2003).

STOP 6

Paleoproterozoic Biwabik Iron Formation-Highway 53

Location: UTM: 536,263E/5,256,200N; Outcrops along north-bound exit ramp from Hwy 53 to Hwy 37, Eveleth.

Description:

This exposure of gently south-dipping strata is part of the Lower Cherty member of the Biwabik Iron Formation. It lies nearly at the crest-line of the anticline that forms half of the Horn structure. The iron-formation forms a transitional contact with the underlying Pokegama Quartzite exposed at Stop 7. Both formations have fine- to coarse-grained sandy textures and cross-bedding, consistent with a high-energy,

near-shore depositional environment. Bimodal-bipolar cross-strata in the iron-formation indicate that tidal currents may have played an active role in deposition (Ojakangas, 1993), though tidal bundles have not been documented. The most significant difference between these two formations is the abrupt change in sediment source from the extrabasinal quartz grains in the Pokegama, to recycled, chemically precipitated chert nearly devoid of detrital grains in the Biwabik.

Discussions:

- 1) One possible explanation for the abrupt change in sediment source in the transition from Pokegama Quartzite to basal Biwabik Iron Formation may be related to topography of the watershed. If the terrane was a relatively flat peneplain, the continued rise of sea level may have essentially drowned the detrital source region.
- 2) The Biwabik Iron Formation is generally divided into 4 members; termed lower Cherty, Lower Slaty, Upper Cherty, and Upper Slaty (Fig. 8). These are convenient field names based on readily apparent bedding attributes; however, they are somewhat misleading. The cherty units are beds of recycled granular chemical precipitates including chert, iron oxides, iron carbonates, and iron silicates. They are interbedded on all scales with “slaty” units of fine-grained, laminated iron silicates and iron carbonates. In most of the Mesabi range, the iron-formation and associated strata were not significantly metamorphosed, and much of the textural and mineralogic attributes are products of diagenesis and subsequent fluid movement. As a result, no slaty cleavage exists, and thus the term “slate” is applied only as a field identifier. Most, though not all of the iron ore mined on the Mesabi range is extracted from cherty members (Fig. 8B).

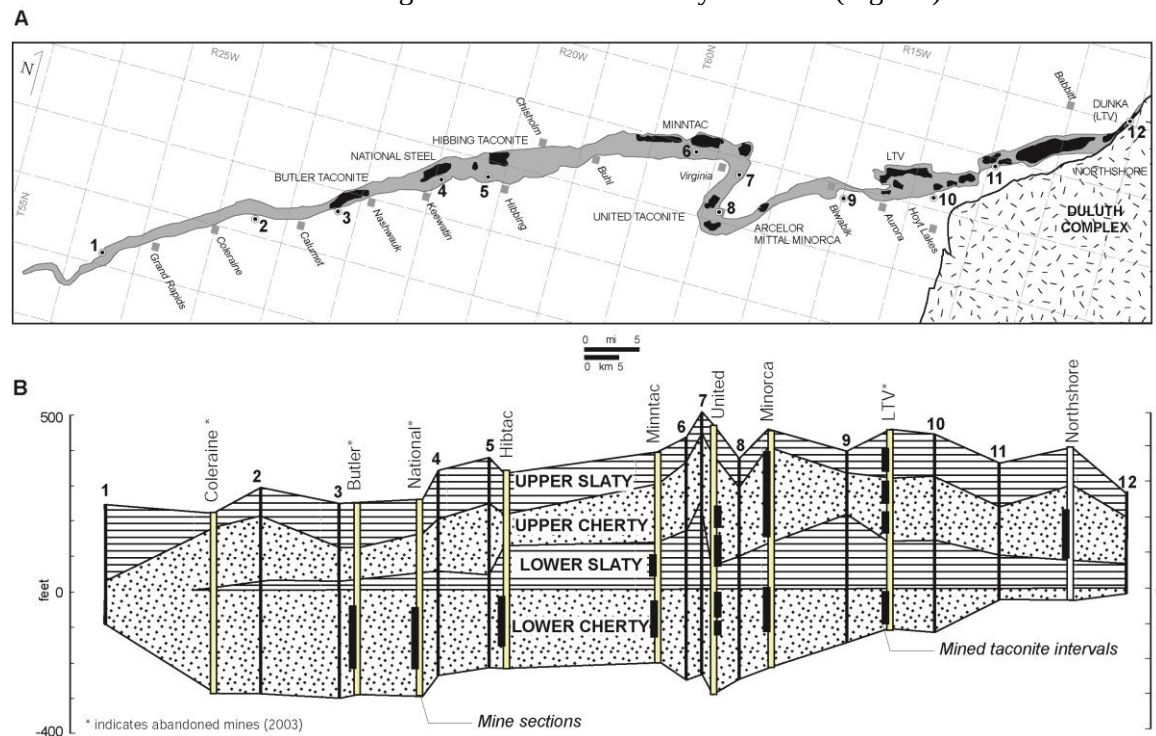


Figure 8. Simplified geologic map (A) of the Mesabi Iron Range showing locations of taconite mines (black) and drill holes (numbered), and cross-section (B) showing subdivision of the Biwabik Iron Formation based on mined sections and drill core, and approximate intervals mined for taconite at each locality (modified from Jirsa and others, 2008).

STOP 7

Paleoproterozoic Pokegama Quartzite-Highway 53

Location: UTM: 535,956E/5,256,913N; Outcrops along north-bound entrance ramp onto Hwy 53 from Hwy 37, Eveleth.

Description:

In the area of the Virginia horn, the Pokegama consists largely of siltstone and shale. This exposure represents the sandy, upper member of the Pokegama Quartzite, which is only a fraction of the units' total thickness of 26-51 m. It is quartz arenite characterized by coarse grain size, intraclasts of shale and siltstone, and massive beds as thick as 1.5 m, separated by thin beds of shale and siltstone. Ojakangas (1993) interpreted that the deposition of this facies occurred within a high-energy, lower tidal or subtidal environment. Because stratigraphic dip is southward, the outcrop at this location is inferred to be several meters stratigraphically beneath the iron-formation at STOP 6.

Discussions:

- 1) The basal strata of the Pokegama Quartzite is marked locally by conglomerate composed of a poorly sorted array of clasts derived from underlying Archean and Paleoproterozoic (diabase dikes) bedrock. The patchy distribution of conglomerate, and the presence of red jasper in fractures on some Archean exposures (as at STOPS 1 and 5), implies that chemical sedimentation abruptly overstepped clastic deposition during early evolution of the Animikie basin.
- 2) The contact between the Pokegama Quartzite and overlying Biwabik Iron Formation is conformable and gradational. In the transition zone, both units contain similar sedimentary structures and grain size, implying continuity of depositional process. The primary difference between them is grain composition—the Pokegama grains are epiclastic vs. those in the Biwabik are reworked from poorly lithified or unlithified chemical precipitates. The absence of epiclastic grains in the nearly 200 m-thick stratigraphic section of the Biwabik begs the question: How was this detritus abruptly shut off from the watershed?

STOP 8

Abandoned and flooded Rouchleau “natural ore” mine

Location: UTM: 535,710E/5,261,650N; Mineview in the Sky overlook near Virginia.

Description:

North from this overlook is a 3-mile-long complex of abandoned mining properties, known collectively as the Rouchleau mine, all developed within the Paleoproterozoic Biwabik Iron Formation. Actually, within this view there were some 15 separately named mines that collectively shipped ore during the period 1893-1986. All of them, and nearly 400 more along the 150-mile long Mesabi Iron Range, extracted oxidized (hematite- or goethite-rich) and leached (silica-depleted) iron-formation referred to locally as “natural ore.” Iron-formation at this point lies on the north-trending limb separating the syncline to our west and the anticline to the east. The natural ore deposits here are localized along a set of faults (Fig. 2A) that presumably provided the plumbing system for fluids that first oxidized the formation and produced permeability, then leached silica from the porous zones, thus increasing ore tenor. Natural ores typically contained as much as 50 percent iron and less than 10 percent silica. Since about the 1950's, the principal “ore of choice” has shifted from hematite- to magnetite-rich deposits. The mammoth open-pit mine in the distance to the northwest, and another just southwest of the highway, are developed in unoxidized magnetite ore containing about 30 percent iron, and 50 percent silica. The ore mined at these locations, and several others along the range is the source of the iron concentrate known as taconite. The name taconite has also been applied generally to magnetite-bearing iron-formation where it contains sufficient iron content to be mined for a profit using today's technology.

Discussions:

1) Origin of “natural ore”

Nearly 70 percent of the 3.6 billion metric tons of iron ore produced on the Mesabi range between years 1892 and about 2000 was extracted as natural ores. Although it is generally accepted that these ores formed by oxidation and leaching along folds, faults, and bedding planes, the source and composition of altering solutions and the timing of alteration have been subjects of considerable debate among economic geologists for nearly 80 years. Much of the literature and geologic observations on the issue are reviewed in Morey (2003). Many writers support the concept of descending meteoric waters to account for the dissolution of silica and oxidation of iron minerals. Others, including Gruner (1930) believed the geologic features were better explained by ascending hydrothermal solutions. Gruner’s theory failed to gain common acceptance, in part because no driving mechanism for such a hydrothermal system could be envisioned. The integration of Animikie Group strata into the tectonic context of the Penokean orogen in east-central Minnesota revived the theory of hydrothermal fluid flow within the Pokegama Quartzite and ultimately the iron-formation, as part of a continent-scale, gravity-driven ground-water system (Morey, 1999). The debate continues—fueled in part by the observation that most of the alteration occurs near the present land surface. *Field trip #1* in this guide explores these issues more fully.

2) Highway relocation

This overlook will soon be gone, as taconite mining to the southwest is slated to expand into the Rouchleau pit area. As a result, U.S. Highway 53 will be rerouted to skirt the new mining. Currently the Minnesota Department of Transportation (MnDOT), the Department of Natural Resources, and the mining company are engaged in geotechnical work and discussions to evaluate the various potential new routes for the highway. Exploratory drilling (for both ferrous and non-ferrous metallic mineral potential), geotechnical drilling, and geophysical surveying have been conducted by MnDOT to assess the financial and engineering risks of two relocation alternatives proposed through the Rouchleau Pit (Fig. 9), as well as a third alternative through the mine site to the southwest. This type of apparent conflict between surface infrastructure and mining has characterized development along the Mesabi Iron Range for more than 120 years. It was particularly acute during the shift in the “ore of choice” in the 1950’s and 1960’s from natural ores that typically occur in narrow and steep, structurally-controlled deposits; to taconite that occurs in more widely distributed layers. Entire cities have been moved and removed, as at Hibbing (*See Field Trip B in this guide book for example*).

3) Wind turbines on northern horizon

Minnesota Power’s Taconite Ridge Energy Center is visible to the north, located on U.S. Steel property in Mt. Iron. It consists of 10 wind turbines that generate 25-megawatts, capable of powering the equivalent of 8,000 homes annually.
(Reference=http://www.hometownfocus.us/news/2013-09-06/Mining_Features; accessed 2/2014)

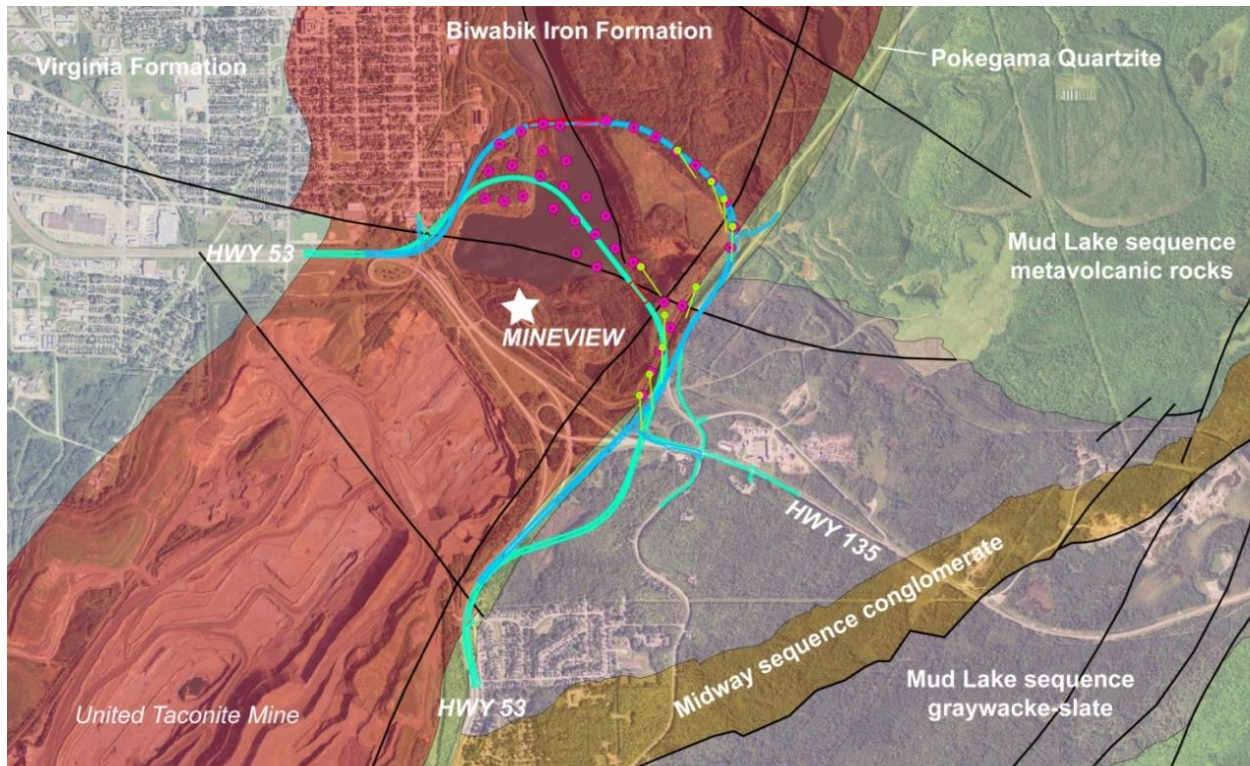


Figure 9—Map showing potential new routes for a portion of Highway 53 (blue and green lines through Rouchleau mine complex), and associated test drilling (magenta dots=holes to iron-formation; green dots=holes to Archean) to evaluate both the engineering and resource implications of proposed realignments. Colored overlay on air photo imagery shows geologic units and faults from Jirsa and others (2012). Image created using data from MnDOT.

STOP 9

Archean Giants Range batholith at “Confusion Hill”

Location: UTM: 534,337E/5,269,458N; outcrops at Laurentian Wayside, near Highway 169 south of its split with Highway 53.

Description:

Exposures at Confusion Hill are a small part of the Giants Range batholith, which forms the core bedrock of the Laurentian (drainage) divide. The batholith is a belt of intrusions that can be traced on geophysical maps and outcrop east to the Mesoproterozoic Duluth Complex, and west beyond the western border of Minnesota. It separates Archean supracrustal sequences in the Virginia Horn from those of the Vermilion District to the north—making stratigraphic correlation between the two districts speculative in the near absence of high-precision geochronologic data.

Exposed near this wayside and in road cuts on both sides of the highway is an array of variably layered intrusions having both tonalitic (white) and dioritic (black) compositions. A cursory look shows intrusive relationships that conclusively demonstrate that diorite was emplaced into tonalite at one locality; and at another, tonalite was emplaced into diorite. In detail, all compositions intermediate between the two end members are also present locally. Although the dioritic component is abundant here, the bulk of the mapped unit is tonalitic. Emplacement of this unit, now known as the Lookout Mountain tonalite, probably involved some degree of magma mingling. It may be equivalent to tonalitic gneiss exposed along strike to the east and having a somewhat imprecise U-Pb zircon age of 2718 ± 67 Ma (Southwick, 1994). Dikes of tonalite that cut the adjacent high-grade supracrustal rocks of the high-grade Minntac sequence contain metamorphic fabrics, yet little evidence of metamorphic origin can be

seen in the interior of the body, implying it is syntectonic to pre-tectonic with respect to D₂ deformation. U-Pb zircon dates (Boerboom and Zartman, 1993) of two components of the batholith exposed to the north bracket the age of D₂ deformation between about 2674 and 2682 Ma.

STOP 10

Archean diorite in Giants Range batholith

Location: UTM: 534,337E/5,269,458N; outcrops at Laurentian Wayside, near Highway 169 south of its split with Highway 53.

Description:

Rock type exposed in this now partially reclaimed quarry is a massive hornblende-pyroxene-biotite diorite. Currently, little is known about the intrusion, as no petrologic study, geochronologic analysis, or mapping of its contacts has been conducted by the authors. Nevertheless, it is similar to other small alkalic plutons in and adjacent to the Giants Range Batholith. These vary in composition—in some cases within a single intrusion—from syenite to monzodiorite to lamprophyre and pyroxenite (Boerboom, 1994). It is interesting to speculate that this intrusion may fall into the category of the late alkalic to calc-alkalic intrusions that are temporally, and likely geochemically, related to Timiskaming-type assemblages such as the Midway sequence.

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FRIDAY AFTERNOON FIELD TRIPS
MAY 16, 2014

FIELD TRIP A

Friday, May 16, 2014

STATE DRILL CORE LIBRARY—HIBBING MINNESOTA **Minnesota Department of Natural Resources—Division of Lands and Minerals**

LEADERS:

*Dave Dahl, (MnDNR), and
Dean Rossell (Kennecott)*

INTRODUCTION

The Minnesota Department of Natural Resources maintains a Drill Core Library in Hibbing, Minnesota. It serves as the single State of Minnesota repository for archiving bedrock and earthen material core samples collected during minerals exploration, engineering, and geoscience research programs across the state. The library attracts a worldwide audience of scientists who use core samples to develop new ideas about the capacity of the state's bedrock and glacial materials to host mineral resources, and to model the geologic forces and features that have shaped the state's foundation. This trip will provide opportunities for visitors to tour the facility and view some very old exploration data collections, century-old historical cores, and recent scientific and exploratory cores. The latter include cores taken from iron-formation, Midcontinent rift peridotite and gabbro, greenstone belt prospects, and sedimentary and glacial settings.



Figure 1. Interior of Building #3 showing stacked core boxes.

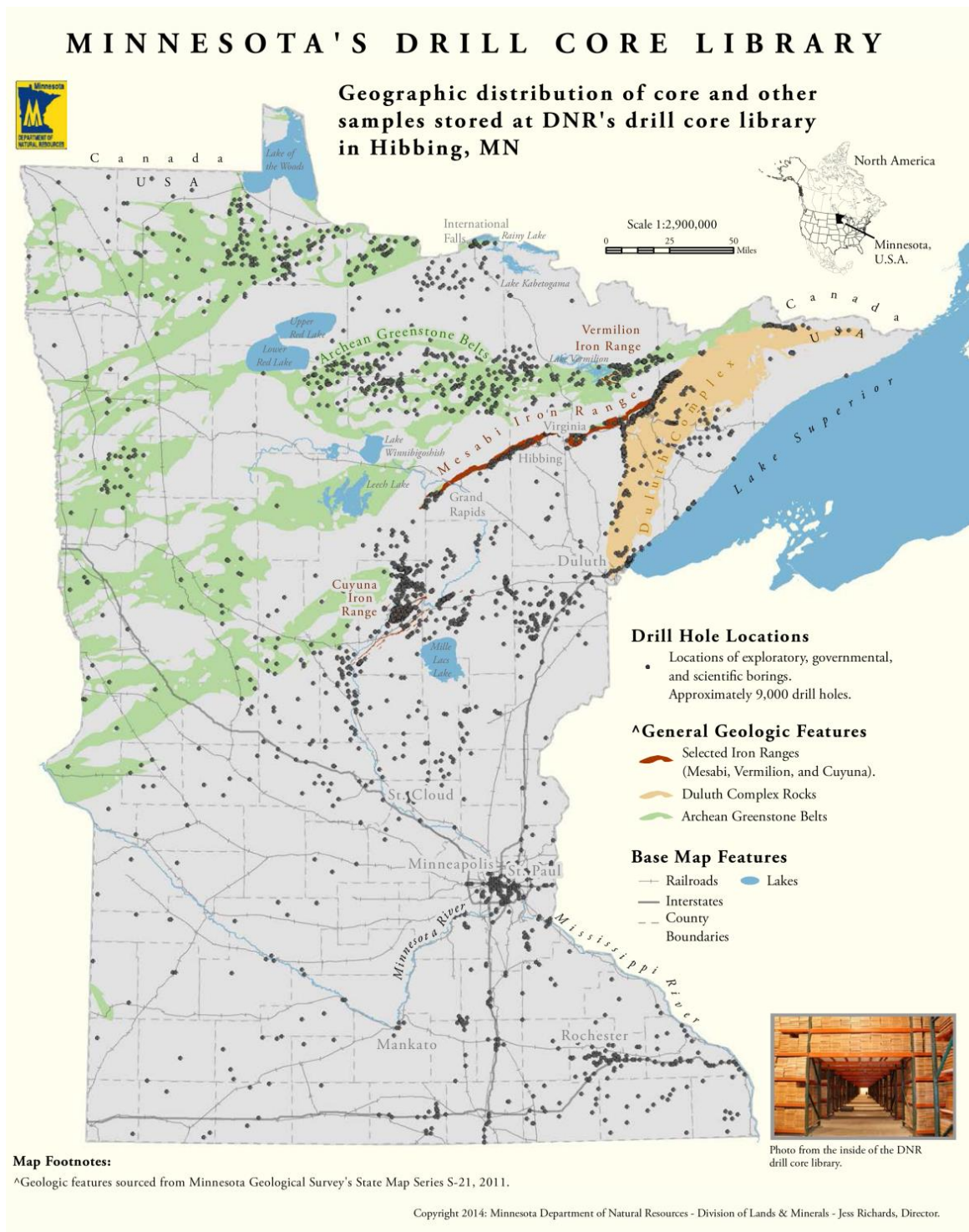


Figure 2. Map of Minnesota showing locations of drill holes from which cores stored at DNR were extracted.

The three buildings that comprise the core library facility house more than 3 million lineal feet of drilled core samples archived from approximately 9,000 exploratory and scientific borings. Some archived samples are well cuttings (depending on the drilling method employed during sampling). The archive collection contains approximately 7,000 mineral exploration cores, 1,500 roadway and bridge foundation cores, and 500 cores collected during scientific, governmental and academic research investigations (Fig. 2).

Building #1, built in 1972 has a storage capacity of 400,000 lineal feet of core. Building #2, constructed in 1979 has a 600,000 lineal foot storage capacity. Building #3 (Fig. 1), originally constructed in 1989 with a capacity of 800,000 lineal feet, has been expanded twice. In 1995 the building was doubled in size through an addition, and in 2009 the building was nearly doubled in size again through addition of a wing. In present configuration, the three buildings have capacity to store approximately 4 million lineal feet of NQ-sized core samples. Core samples are normally transferred to the facility in two-foot long boxes, 5 core segments per box, or 10 feet of core per box. Box storage capacities range from 7 segments for small diameter (A- and E-size) core to 2 segments for large diameter (PQ—size) core. Boxes are generally designed to hold 50 lbs weight or less.

Today, most exploratory boring samples are delivered to the library in fulfillment of statutory requirements (M.S. 103 I.601 and 103 I.605) which have been in effect since 1980. The library archives are augmented by substantial collections of historical (pre-1980) core samples that have been received via donation from mineral exploration companies or through consolidation of agency core collections. The earliest known collar date in the core collection is 1905, for exploratory cores taken along the Gunflint trail. Mineral exploration archive documents housed in Hibbing indicate that core samples were obtained in the Vermilion district as early as the 1870's.

Core samples are expensive to obtain and maintain, but that money is well spent. Samples acquired to meet one investigative objective commonly are “recycled and reused” several times in subsequent programs. They can be used to test new working models of geologic processes, and to provide new insights on the disposition and location of mineral resources. Analytical techniques (geochemical, geochronologic, and geophysical) are constantly evolving, and these cores provide ready materials for testing within well constrained geologic contexts. Archived cores have provided the basis for advancement in the evaluation of several copper-nickel, gold, titanium, and iron deposits and prospects (Tamarack, Birch Lake, Maturi, Spruce Road, Serpentine, Mud Creek, Lost Lake, Virginia Horn, Longnose, TiTac, Buckeye, Emily and others). New private investments to advance these properties, some of which include School Trust or other state-owned mineral lands, are on the order of \$200 million over the past decade. The DNR recently convened a working group to increase efficient delivery of core library services and to attract research and investment in the evaluation and understanding of Minnesota resources.

FIELD TRIP B

Friday, May 16, 2014

HIBBING'S IRON MINING AND CULTURAL HISTORY

LEADERS:

*Henry Djerlev,
Bob Kearney,
Erica Larson, and
Hibbing Historical Society Staff*

INTRODUCTION

The nearly 120 years of iron ore mining in the Hibbing area has certainly played a large part in shaping the history and culture of the towns and residents and the rest of the Mesabi Iron Range. Within the outline of what is now called the Hull-Rust-Mahoning open pit, more than 30 separate mines operated from 1895 to the present. The early miners emigrated from dozens of countries, and all of their various languages were spoken in the mines. The mix of cultures made for awkward communications in the mines, and often created ethnic neighborhoods in the mining locations and larger towns. Initial underground mining quickly changed to open pit due to the nature of the ore body and the introduction of large steam operated drills, shovels and trains. Very quickly, with eastern monies invested, some of our larger corporations emerged, such as United States Steel. Very small "mining locations" grew into formal towns like Hibbing founded by Frank Hibbing and A. J. Trimble. Open pit mining quickly progressed from small individual pits, into one large open pit that was nicknamed "The Grand Canyon of the North". This expansion made it necessary to physically move, from 1919 to 1921, what was called North Hibbing to the south in order to make room for increased mining. This gave Hibbing another nickname: "The Town that Moved". During this short tour of Hibbing, the National Historic Landmark of the Hull-Rust-Mahoning Mine will be visited and several of the historic buildings that were made possible by iron mining dollars. These include the spectacular Hibbing High School and the Hibbing Historical Society Museum. Stop locations given in latitude/longitude.

***Departure Point- Hibbing Park Hotel (47° 25' 38.71" N/92° 35' 26.22"W)**

On the short trip from the Hibbing Park Hotel to Stop 1 the bus will pass by several historic points in Hibbing such as the one time home of Andrew "Bus Andy" Anderson, built in 1920. Andy with his partner Carl Wickman initiated what was to become The Greyhound Bus Company.

Carl, after losing his job at the Alice mine started as a salesman for the Hupmobile company. In 1914, after seeing the failing sales of the seven passenger Hupmobile, he tried to show his clients what a great product the



Figure 1: Early photo of one of Mesabi Transportations Hupmobile Buses.

Hupmobile was by taking people for short rides. Wickman was soon giving miners rides to and from work for a cheap fifteen cents a ride. When they found out that giving the miners transportation was more profitable, Wickman and Anderson created the Mesaba Transportation Company.

Three years after starting the company, they were running 18 buses and were making \$40,000 a year. In 1922, he sold the company for \$60,000. In 1933, the company was formally named The Greyhound Corporation and was running nationally.

Other historic points along this leg of the tour will be the Sons of Italy Hall (1923), Mesabi Railway Company(1921)—now Zimmy's Restaurant, the Androy Hotel (1923), Bennett Park, and the Greyhound Bus Museum.



Source: The Town that Moved website

Figure 2 - Frank Hibbing

The small city got its start by a German immigrant named Frank Hibbing who founded the town in 1893. Originally named Frans Dietrich Von Ahlen, Frank took his mother's last name of "Hibbing" which comes from English descent. Frank decided to do this out of honor for his mother who passed away in his infancy and thought that this would be a good move for his exploration of the "New World."

Frank Hibbing originally settled in Beaver Dam, Wisconsin where he worked at a farm and shingle mill. Originally he had hopes of becoming a lawyer but after finding out the extreme differences between the German and English language, he decided to forego that dream and then became interested in the area's most abundant resource: timber.

In 1887, Frank Hibbing moved to Duluth and became a real estate salesman which eventually lead him further north into the Vermillion Range. It was not until 1892 when he and thirty men set out to cut a road from Mountain Iron to what was then called Section 22. While cutting this road, Frank Hibbing found iron ore on the ground and realized what that meant to the area's economy. Little did Frank Hibbing know at the time, this ore deposit would be one of the largest in the world!

In 1893, the city of Hibbing was laid out and named in honor of Frank Hibbing. The city even has a statue to the German who had the sense to notice the reddish soil and the value that it had. Artist Robert Mitchell, born in Alice location, created the statute which was dedicated on October 21, 1941. Robert was the son of a Hanna Company mining captain for whom Mitchell location was named.

Frank took so much pride in his new town that he used personal means to finance the first water plant, electrical plant, hotel, saw mill, and bank building. Frank Hibbing made Duluth his home for the last ten years of his life until his death from appendicitis on July 30, 1897. He did retain close communication with "his" town during that time. Frank Hibbing was only 40 years old.



Source: The Town that Moved website

Figure 3- Frank Hibbing Park

Stop 1- HULL-RUST MAHONING MINE (47° 26' 50.53" N/92° 56' 45.86"W)



"The Largest Open Pit Iron Ore Mine in the World" is more than three miles long, two miles wide and 600 feet deep. This man-made "Grand Canyon of the North" was the one of the first open pit mines on the Mesabi Iron Range. This amazing view continues to grow as the Hibbing Taconite Company mine expands its mining operations. Rotary drills, 33-cubic-yard shovels and 240-ton production trucks can be seen in action at this National Historic Site. Occasionally, you may witness a production mining blast of nearly 1 million tons used to clear bedrock away and break the taconite ore for processing in the plant.

Figure 4: Recent photo of the Hull-Rust-Mahoning Pit

Since 1895 more than 1.4 billion tons of earth have been removed on its 2,000 acres of land, and more than 800 million gross tons of iron ore have been shipped from the mine. At peak production in the 1940's, as much as one quarter of the ore mined in the United States came from the Hull Rust Mine. Currently Hibbing Taconite Company (Cliffs) produces approximately 8 Million tons of taconite pellets annually. Over 520 million tons of waste material and 690 million tons of iron ore have been removed.

A slide presentation in the observation building explains the colorful history of the mine and early mining. An observation building, mine exhibits, mine shovel bucket, mining truck, interpretive graphics and a walking trail complete the trip to the Hull Rust Mine View.

A miner poses near the edge of the pit. This area of the Mesabi Range was first explored in 1893, shortly after the Mountain Iron Mine was established in 1892. The early development was as an underground mine, but open pit mining soon proved to be a better choice because of the shallow nature of the ore deposits. The many smaller open pit mines developed in the area soon merged into one large mine.

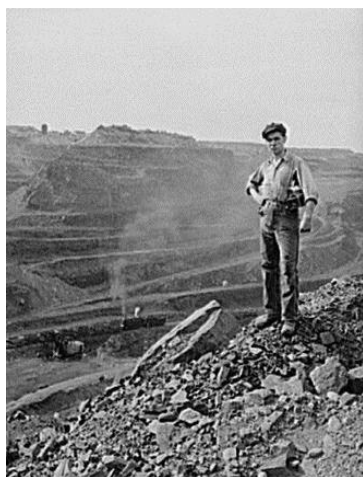


Figure 5- Early Miner along the edge of the Hull-Rust Pit

The growth of the mine even resulted in the town of Hibbing being relocated to accommodate expansion. The move started in 1919 and took two years to complete at a cost of \$16,000,000. A total of 185 houses and 20 businesses were moved, and some of the larger buildings had to be cut in half for the move. Only a portion of the network of city streets and foundations from old North Hibbing remain in the vicinity of the Mine Observation Overlook.



Figure 6- Street sign in North Hibbing

Other historic points along the second leg of the tour will be Frank Hibbing Park (1941), the Godfrey House (1920's) and the Mitchell-Tappan House (1897).

Stop 2- The Hibbing High School - 1922 (47° 25' 33.30" N/92° 55' 57.08"W)

One of the buildings that was built during the Oliver Mining Company's relocation of North Hibbing was the Hibbing High School. Built in 1921 by the Oliver Mining Company, the school building originally cost almost \$4 million dollars. In today's dollars that would be close to \$45 million dollars!



Figure 7- Hibbing High School under construction.



Figure 8- Present Day Hibbing High School

Why did it cost so much? Well it is simple, in order to lure prospective workers and miners to work in dangerous situations such as tunnels and around explosions, they had to provide a first class environment for their families and especially their children. So they went all out for the education.

The auditorium of the high school was modeled after the famous Capitol Theatre in New York City. The auditorium has cut glass chandeliers which were imported from Belgium which light the 1800 velvet seated venue. The chandeliers were originally priced at \$15,000 when built and today are insured for over \$250,000 each.

The auditorium has a rarity in it as well, which is a Barton pipe organ. Only two are in existence in North America. With 1800 pipes, it can synthesize any instrument excluding the violin.



Figure 9- Hibbing High School Auditorium

A few celebrities have attended this high school including basketball star and musician Robert Zimmerman aka "Bob Dylan". Robert Zimmerman may have been born in Duluth, Minnesota on May 24, 1941 but it is Hibbing, Minnesota where he grew up.

Robert grew up in Duluth, MN until he was six years old. It was when his father was sick with polio that his family moved back to his mother's hometown of Hibbing, MN. This is where he fell in love with music and formed many bands throughout his high school career including such names as "The Shadow Blaster" and "The Golden Chords."



Figure 10- Dylan Family Home



Figure 11- Bob Dylan, 1963

During the high school talent show, Danny and the Juniors, played so loud that the principal cut off the microphone.

Robert Zimmerman left Hibbing in 1959 to move to Minneapolis so he could enroll at the University of Minnesota. This is where Robert Zimmerman did two things; he fell in love with folk music and changed his name to Bob Dylan. The reason for the change was that Bob was very familiar with the poetry of Dylan Thomas. In a 2004 interview, Bob stated, "You're born, you know, the wrong names, wrong parents. I mean, that happens. You call yourself what you want to call yourself. This is the land of the free."

Stop 3- Hibbing Historical Society Museum (47° 25' 25.92" N/92° 56' 13.36"W)

The current focus of the Hibbing Historical Society is the documentation and presentation of the early "Mining Locations" that grew up in the vicinity of the surrounding mines. There were three types of mining locations, but all had the common factor that they placed on company-owned land. There were



Source: Hibbing Historical Society

Figure 12- Mahoning Location and Rail Shops

over 175 locations between 1892 and the 1920's on the Mesabi Range. Initially miners and sometimes their families just found an open piece of land near the mine they worked at where they built a very small home from whatever building materials they could locate. These were "squatter's locations" and often called 'chicken towns' as there were often farm animals in the yards or even in the living quarters themselves, during the winter months.

As time progressed the companies decided that if they organized a mining location and built modest homes for its miners, they had better success in keeping a stable and highly trained workforce. These were called "company locations" and were actually townsites that were surveyed in and could included paved streets with water and sewer utilities. Mahoning location is a fine local example of a company location in the vicinity of Hibbing. The homes in these Locations were built by the mining firms and then leased or rented to the employees. In some cases the residents were permitted to purchase their house, but not the land. Common monthly rental fees might be \$1.00 per \$100 invested by the mining company.

Very rarely a mining company, such as United States Steel, would design and build "Model Locations". These communities would consist of more elaborate homes with full services plus community buildings such as recreation halls, hospitals and fire departments. Stellar examples of a Model Location would be Morgan Park in Duluth and the community of Coleraine on the western end of the Mesabi Range. The company plan here was for a more attractive community with higher quality constructed homes that would show them in a better light.

More than 30 squatter's locations and company locations were located just to the north, east and west of Hibbing to service a like number of small early mines.

In 1915, the town of Hibbing had 20,000 people who all had to uproot their homes and families and move them south to the small village of Alice. Many of the buildings were actually lifted and rolled down to Alice. The Oliver Mining Company (later to become US Steel) was the brainchild behind this move and agreed that if the town relocated 2 miles south to Alice, they would develop the downtown buildings with low interest loans for the retailers.

At the Hibbing Historical Society Museum there is a large scale model of the Hibbing with excellent exhibits that describe the physical moving of that "North Hibbing" portion of the town.

The move started in 1919 after four years of careful planning and was completed in 1921. The buildings were all moved down what was called at the time, "the First Avenue Highway" which is still in existence today. In total, about 200 structures were moved to the new town while new structures were also built including the Hibbing High School, the Androy Hotel, the Rood Hospital, and the Village Hall. These buildings were created with mining company money to help ease the settlers' mood about having to move the entire town. Only one structure did not make it to the new town during the move. A hotel tumbled off the rollers and crashed into a million pieces. One eyewitness referred to it as "an enormous pile of kindling."



Figure 13- North Hibbing at crest of Mining.

The city of Alice was then renamed to Hibbing and annexed. The land size of the city of Hibbing is the largest in Minnesota, even surpassing both Minneapolis and St. Paul's city limits! A children's book chronicles the amazing story of the "Town that Moved".

On the very short final leg of the tour we'll pass-by the family home of Bob Dylan (Robert Zimmerman).

End of tour and return to the Hibbing Park Hotel:

Sources:

- City of Hibbing website- (http://www.hibbing.mn.us/index.asp?Type=B_LIST&SEC={0BA3C178-6F53-4831-B739-0315936323C6})
- Hibbing Historical Society
- Hibbing High School website- (<http://www.hibbing.k12.mn.us/>)
- "Hibbing Historical Walking Tour" pamphlet - Hibbing Daily Tribune & Roger Saccoman Architecture
- Hibbing Chamber of Commerce website- (<http://www.hibbing.org/pages/History/>)
- Hibbing: The Town That Moved website- (<http://minnesotaghosts.com/index.php/library/mnhistory/81-hibbing-the-town-that-moved>)
- "The "Locations" - Company Communities on the Minnesota's Iron Ranges", 1982, by Arnold R. Alanen (Minnesota Historical Society).

FIELD TRIP C

Friday, May 16, 2014

MINNESOTA DISCOVERY CENTER

LEADERS:

Discovery Center Staff



Figure 1. Miners statue near MDC entrance

The Minnesota Discovery Center museum and research library in Chisholm (a few miles north of Hibbing) houses artifacts, examines mining methods, explores regional geology, and hosts traveling exhibits that highlight the story of the predominantly European immigrants who migrated to this region at the turn of the 20th century to find work in the burgeoning iron ore industry. Their stories document the development of the Mesabi Iron Range, a region that became the nation's largest producer of iron ore. The museum, formerly known as "Ironworld," is perched at the edge of a lake-filled gorge that represents the collective footprint of many open-pit and underground mine properties. This field trip includes a guided tour of the museum, and a trolley ride across a portion of the mine.

FIELD TRIP D

Friday, May 16, 2014

COLERAINE MINERALS RESEARCH LABORATORY
Natural Resources Research Institute
University of Minnesota-Duluth

LEADERS:

Dick Kiesel (Director CMRL)

Dave Hendrickson (Director Strategic Planning)

Matt Mlinar (Program Coordinator Mineral Processing)

Basak Anameric (Program Coordinator High Temperature Process)

This trip will tour the Coleraine Minerals Research Laboratory (CMRL) in Coleraine, about 25 miles SW of Hibbing. The CMRL conducts applied research that supports technology-based economic development for iron ore mining, non-ferrous minerals, industrial minerals, environmental remediation, alternative iron making, and the use of taconite mining products for various value-added aggregate applications. The facility consists of an analytical laboratory, mineral processing and pyrometallurgical processing capabilities from bench to pilot scale for applied research and development projects. Geographic proximity to the nation's largest iron mining district (the Mesabi Iron Range) has meant that the CMRL has historically conducted minerals development research, and contributed to the training and development of a substantial number of iron mining and minerals industry professionals. Demand varies from solving short term problems, identifying unique market niches, to providing medium to long range technical innovation and developing products and processes for the future.



FIELD TRIP E

Friday, May 16, 2014

MINEVIEW FROM A CANOE

LEADER:

*Mark Jirsa (Minnesota Geological Survey);
with assistance from*

*Daniel Jordan (Iron Range Resources and Rehabilitation Board), and
Dale Cartwright (Minnesota Department of Natural Resources, Division of Lands and Minerals)*

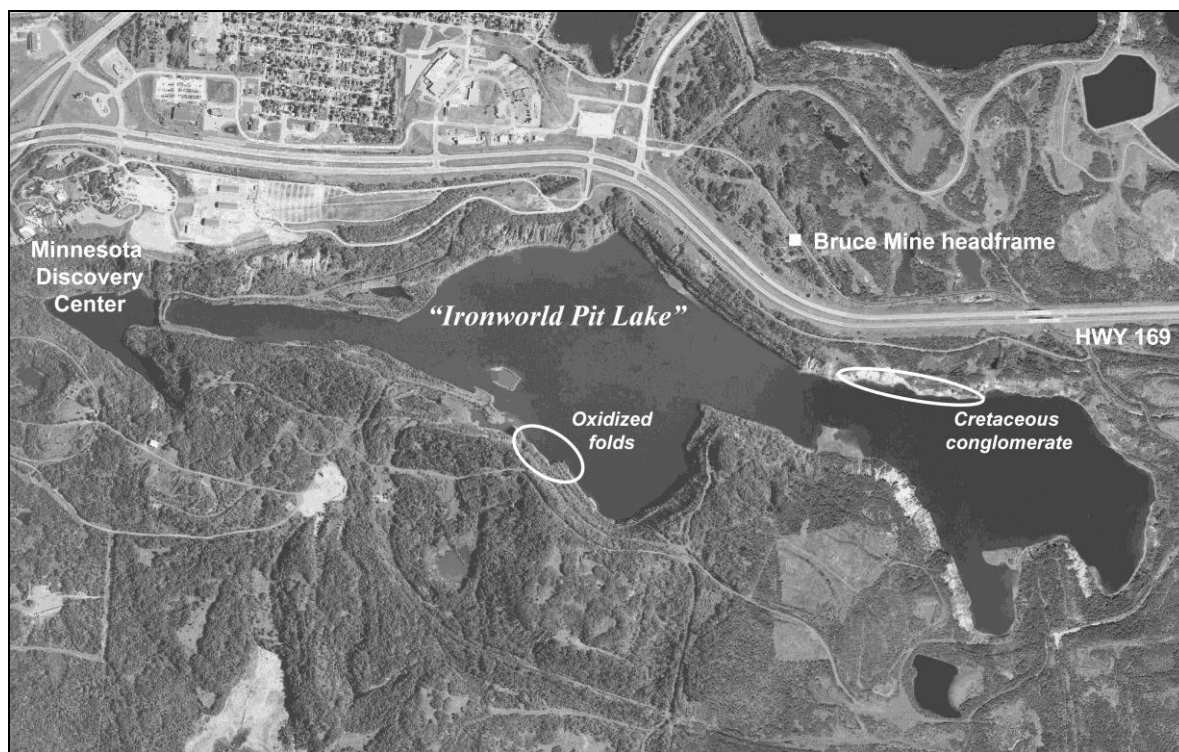


Figure 1. Airphoto image of the collection of inactive natural (hematite-goethite) ore mines that form what is referred to here as “Ironworld Pit Lake,” just south of Chisholm. Image shows general locations of 2 main geologic features (ovals) that will be viewed from the gunnells, and other local landmarks. Width of photo ~ 2.5 miles.

This trip offers a unique duck’s-eye view of the geology at “Ironworld Pit Lake” in Chisholm, a few miles north of Hibbing. The lake occupies the abandoned footprints of as many as 15 separate natural (hematite-goethite) ore mine properties. These include the Pillsbury (operating years 1898-1969), Glen (1902-1957), Leonard-Burt (1909-1974), Leonard (1903-1974), Clark (1900-1925), Monroe-Tener (1905-1981), South Tener (1928-1981), Bruce Annex (1929-1937), Dunwoody (1917-1977), Douglas (1942-1977), Neville R (1947-78), Duncan (1914-1970), Pillsbury-Brown (1951-1978), Chisholm (1901-1967), and Godfrey (1926-1963) mines.

One wall of the pit exposes a 30-foot thick slab of what is inferred to be Cretaceous iron-rich conglomerate that was glaciotectonically dislodged and thrust over till (Fig. 1). In the early days of mining (1892-1950’s), these hematite-pebble conglomerates were prized as extremely high-grade ore.

Another wall portrays fold and fault structures that likely were genetically related to the formation of natural ores by oxidation and leaching of various layers of Biwabik Iron Formation. The central part of mine pit appears to follow major NW-trending fault/fold structures and subparallel joints (Fig.2), and thus obliquely crosses the strike of iron-formation. As a result, a comparatively thick section of strata is exposed along pit walls—perhaps including parts of the Lower slaty, Upper cherty, and Upper slaty members, depending on water level (*SEE Field Trip 1, this guidebook for vernacular*).

Intuitively, most of the natural ore was exhausted from this site; however, exposures of oxidized (near-ore) can be seen locally on pit walls, depending on water level. The various types of natural ore can generally be color-correlated with the inferred protore (via Gruner, 1946). For example, the precursor of “blue ore,” composed of semi-massive martite (magnetite pseudomorphed by hematite), may have been cherty magnetite-rich layers. Yellow-colored ores that contain primarily goethite and limonite (a generic term for undifferentiated, hydrated iron oxides; typically hydrated goethite) formed from layers of thinly bedded to laminated (slaty) iron-silicates. Brown-colored ores consist of mixtures of goethite, limonite, hematite, and martite, and likely were derived from intimately interbedded cherty and slaty layers. The processes of oxidation and localized leaching of silica and iron-carbonate results in considerable volume loss (as much as 50%), and some of the slump structures visible on pit walls are a product of collapse related to this alteration. Some are also undoubtedly related to collapse into historic underground workings. In general, it is difficult, and in some cases impossible, to assign specific episodes of deformation to individual structures (*SEE discussion of iron-formation structures in Field Trip 1, this guidebook*).

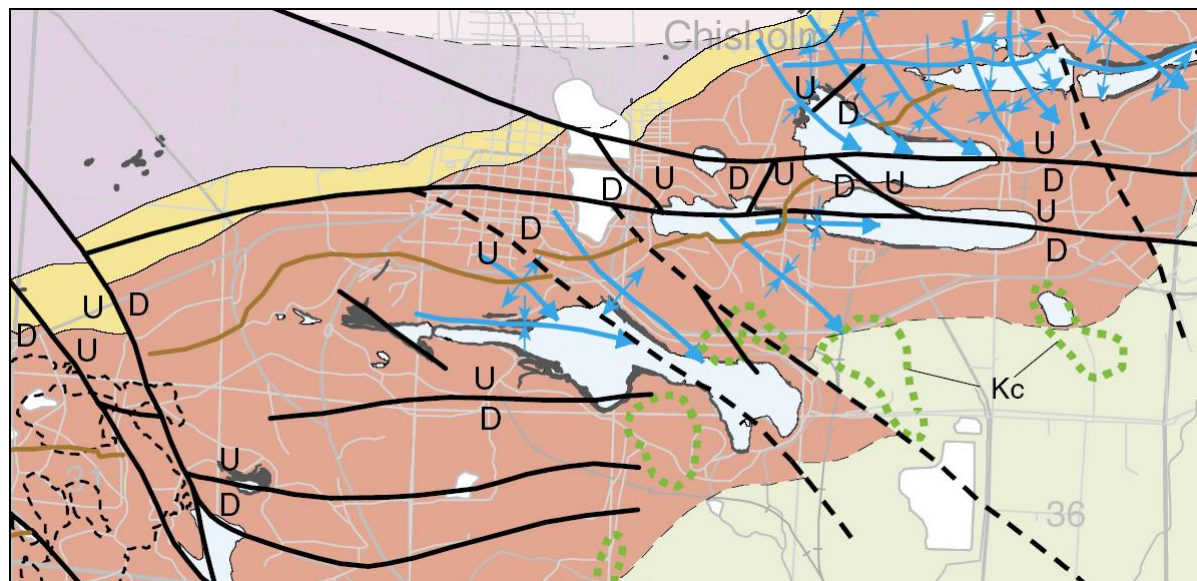


Figure 2. Bedrock geologic map of the Chisholm area showing mine pit lakes (pale blue), faults (solid and dashed thick black lines with letters denoting inferred fault movement; Up, Down), fold axes (thick blue lines), and subsurface extent of shallowly southeast-dipping Biwabik Iron Formation (reddish). Brown line in iron-formation approximates the surface trajectory of the Intermediate slate unit, which marks the stratigraphic top of the Lower cherty member. Dashed green lines represent known extent of Cretaceous strata. Width of photo ~ 6 miles (gray section lines). Map is clipped from 1:100,000-scale (Jirsa and others, 2005).

Historically, mining geologists and engineers on the Mesabi Iron Range classified natural ore bodies into 3 main types: trough, fissure, and flat-lying (Wolff, 1917). Trough ore bodies are as large 3000 feet long, 1000 feet wide, and 200-400 feet deep. They formed typically along permeable faults or joint sets

by selective leaching and oxidation. Consequent collapse into linear zones of reduced volume produced the synclinal or trough shapes. Fissure ore bodies are similar, but smaller ($\leq 200' \times 2.5' \times 50'$)—having formed along lesser joint structures and typically involving only minor collapse. Flat-lying ore bodies represent oxidation and leaching along select stratigraphic intervals. They are irregular in shape, commonly follow bedding planes outboard of vertical faults or joints, and therefore persist over considerable distances. The Godfrey Mine that lies just south of Ironworld Pit Lake was developed in such an ore body. Ore was mined there from a 20-30 foot-thick silicate horizon that lies at the stratigraphic top of the Lower cherty member, just beneath what's known as the Intermediate Slate or Paint Rock horizon. The Godfrey Mine's underground workings extend for nearly a mile, and produced more than 12 million tons of ore.

The ores were extracted from this area utilizing first underground mining, followed by open-pit methods. The underground workings were recently digitized by the Lands and Minerals Division of the Department of Natural Resources (DNR) using historic paper records. The resulting 3D imagery reveals an extensive network of underground workings at various depths (Fig. 3). The deepest of the mines shown here was the Monroe-Tener, at ~ 220 feet below surface. Much of the current lake basin was created by subsequent open-pit mining.

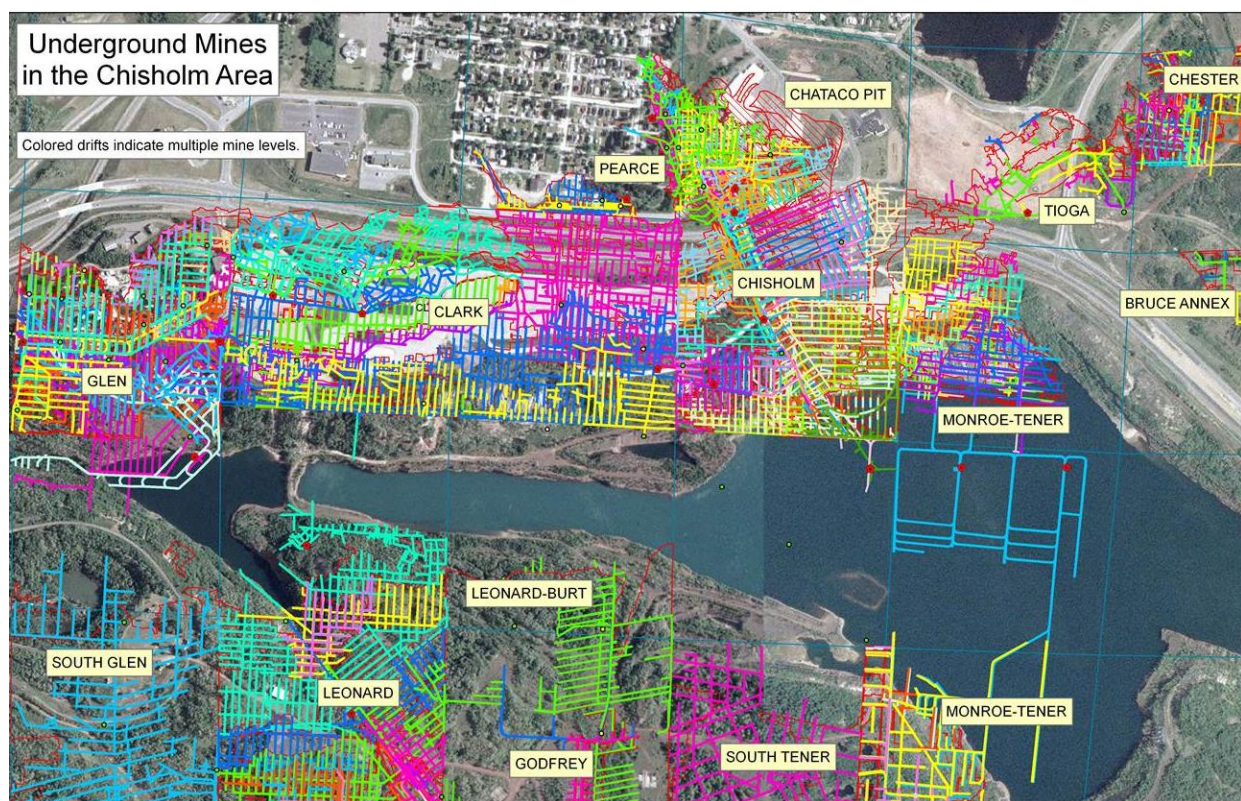


Figure 3. Airphoto image of the Chisholm area showing extensive underground workings digitized in 3D. Color coding for drifts and shafts differs for each mine, but represents various depths of workings. From Cartwright and others, 2011. Width of photo ~ 1.5 miles. See website http://www.dnr.state.mn.us/lands_minerals/underground/index.html for more details.

The work by DNR and associated geologic mapping by the Minnesota Geological Survey (Jirsa and Meyer, 2007; Jirsa and others, 2002, 2005; Jennings and Reynolds, 2005) was undertaken in large part to evaluate connectivity of ground and surface waters between historic and active mines on the Mesabi Iron Range. Obviously, the underground workings have significant influence on water movement—at least in

the upper several hundred feet—and they present engineering challenges for potential mining of taconite in the future. On-going surface subsidence into these historic underground workings (via sink holes) continues to damage local infrastructure. Despite extensive underground operations on parts of the Mesabi Iron Range between 1892 and 1961, the only remaining head frame is that for the Bruce Mine just north of “Ironworld Pit Lake” (Fig. 4).

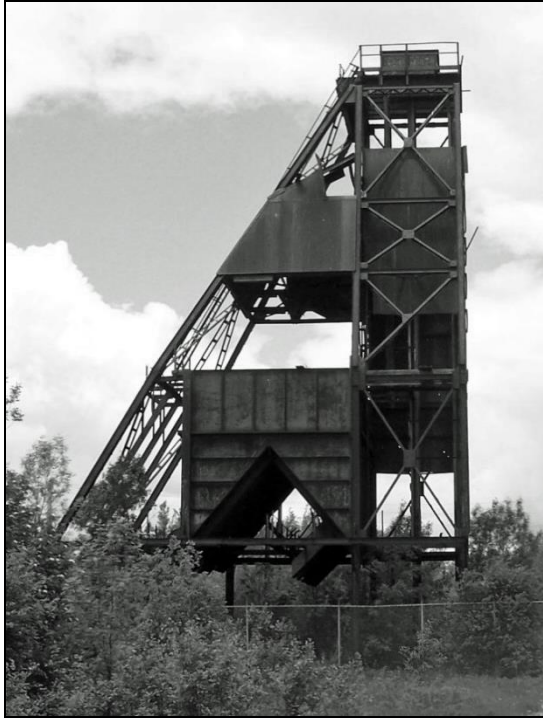


Figure 4. Head-frame from underground mining at the Bruce Mine; the last of its kind on the Mesabi Iron Range.

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