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Cover Photos: Upper left – Dick Ojakangas, Peter Jongewaard and John Arola at the United Taconite mine, MN (2004); Upper right – Hal James and trip participants near Champion, MI (1964); Bottom – Carl Dutton, Bill Cannon and Matt Walton, Northern MI (1975).

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FIELD TRIP 1

VOLCANIC STRATIGRAPHY, HYDROTHERMAL ALTERATION, AND VMS POTENTIAL OF THE LOWER ELY GREENSTONE, FIVEMILE LAKE TO SIXMILE LAKE AREA

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INTRODUCTION

The Vermilion district of northeastern Minnesota contains one of the classic granite-greenstone terranes in the United States. This district comprises the south-central part of the Wawa subprovince of the Superior Province of the Canadian Shield, and has been broadly correlated with the Saganagons Assemblage of the Wawa subprovince in northwestern Ontario (Peterson et al., 2001; Peterson and Patelke, 2003). In Canada, the Wawa subprovince hosts numerous lode gold (e.g. the Hemlo and Renabie districts) and volcanic-hosted massive sulfide (VMS) orebodies (e.g. the Winston Lake, Willroy, Big Nama Creek, Willecho, and Geco deposits; Fyon et al., 1992). The Vermilion district is well known for its numerous, previously mined, massive hematitic iron ore deposits. These iron deposits were discovered in the early 1880s, and virtually all subsequent exploration efforts in the region were targeted on similar iron-formation hosted hematite deposits. However, the discovery of world-class ore deposits in Ontario (the Kidd Creek VMS deposit in 1964 and the Hemlo gold deposit in 1980) led to short periods of both base-metal and gold mineral exploration in the Vermilion district. To date, no lode gold and/or VMS orebodies have been discovered in the Vermilion district to date. This field trip will explore recent research efforts to better understand the potential for VMS mineralization within the Lower member of the Ely Greenstone (Lower Ely). Peterson and Patelke (2004, this volume) explore new research regarding the potential for lode gold mineralization within the Newton Belt of the Vermilion district.

Since the mid-1990's, the Minnesota Department of Natural Resources, the Minerals Diversification Plan of the Minnesota Legislature through the Minerals Coordinating Committee, the Permanent University Trust Fund of the University of Minnesota, the Natural Resources Research Institute (University of Minnesota Duluth), and the University of Wisconsin Oshkosh have provided funding to support a series of systematic studies to better understand the VMS potential of the Lower Ely. Early studies (Lawler and Riihilouma (1997); Hudak and Morton, 1999; Peterson and Jirsa, 1999a; Peterson, 2001) showed that numerous base metal anomalies suggestive of localized VMS mineralization exist in the soils, rocks, and lake sediments within this region.

Beginning in 2000, research efforts regarding the VMS potential of the Lower Ely have been focused in an approximately two mile wide corridor extending from Skeleton Lake northward to Sixmile Lake (Newkirk et al., 2001a, 2001b; Odette et al., 2001a, 2001b; Hudak et al., 2002a, 2002b; Hocker et al., 2003; Hudak et al., in prep. a). This area was chosen for several reasons: 1) there are historical base metal prospects in the area; 2) lake sediment samples in the area contain the highest copper (Sixmile Lake) and zinc (Fivemile Lake) anomalies within the state of Minnesota (Peterson, 2001); 3) the area contains several untested geophysical anomalies that have orientations roughly parallel to the strike of the strata (Peterson, 2001); 4) the area contains regionally extensive quartz-epidote alteration zones (Peterson and Jirsa, 1999a, 1999b; Peterson, 2001) that may be related to ancient, potentially VMS deposit-producing hydrothermal systems; and 5) the bulk of the area comprises state land, and is easily accessible. These research efforts have been multidisciplinary, and have included several months of detailed field

mapping (1:50 to 1:5000 scale) for stratigraphy and metamorphosed hydrothermal alteration mineral assemblages, petrographic studies (600+ thin sections), whole rock lithogeochemical analyses (~200), x-ray diffraction studies and electron microprobe investigations.

REGIONAL GEOLOGICAL SETTING

Supracrustal rocks in the Vermilion district consist of volcanic-dominated stratigraphic sequences of the Wawa subprovince of the Superior Province of the Canadian Shield. Rocks of the Wawa subprovince 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 designated informally as the Leech Lake structural discontinuity (Jirsa et al., 1992). In the region west and north of the Soudan Mine, 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). The Soudan belt 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 volcaniclastic 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 relationship between stratigraphic units within each belt is 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. A simplified regional geological map of the Neo-Archean terranes of northeastern Minnesota and adjacent Ontario is presented in Figure 1-1.

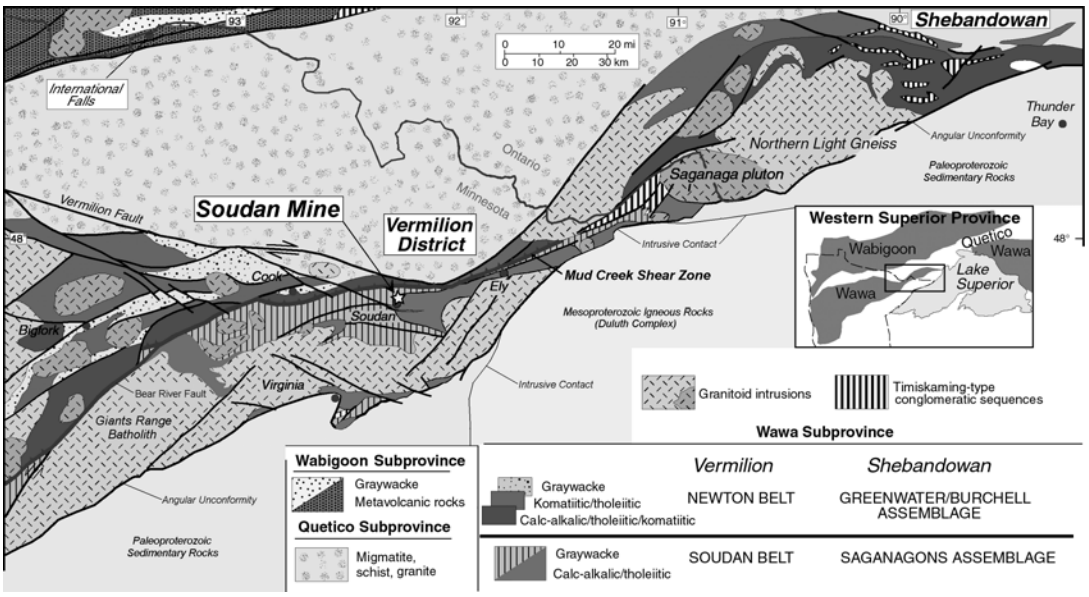


Figure 1-1. Simplified correlation map of Neoarchean assemblages across the U.S. - Canada border (modified from Peterson et al., 2001). Inset illustrates major subprovinces of the southwestern Superior Province.

Lithostratigraphic units in the western Vermilion district include: (1) the Lower member, Soudan Iron Formation member, and Upper member (Upper Ely) of the Ely Greenstone, 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, 1999a) 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 (Fig. 1-2). Contacts between the different units are typically conformable, although considerable overlap in time and space is documented between volcanic and sedimentary sequences (Southwick, 1993). Rock types associated with the lithostratigraphic units in the area are presented in Table 1-1.

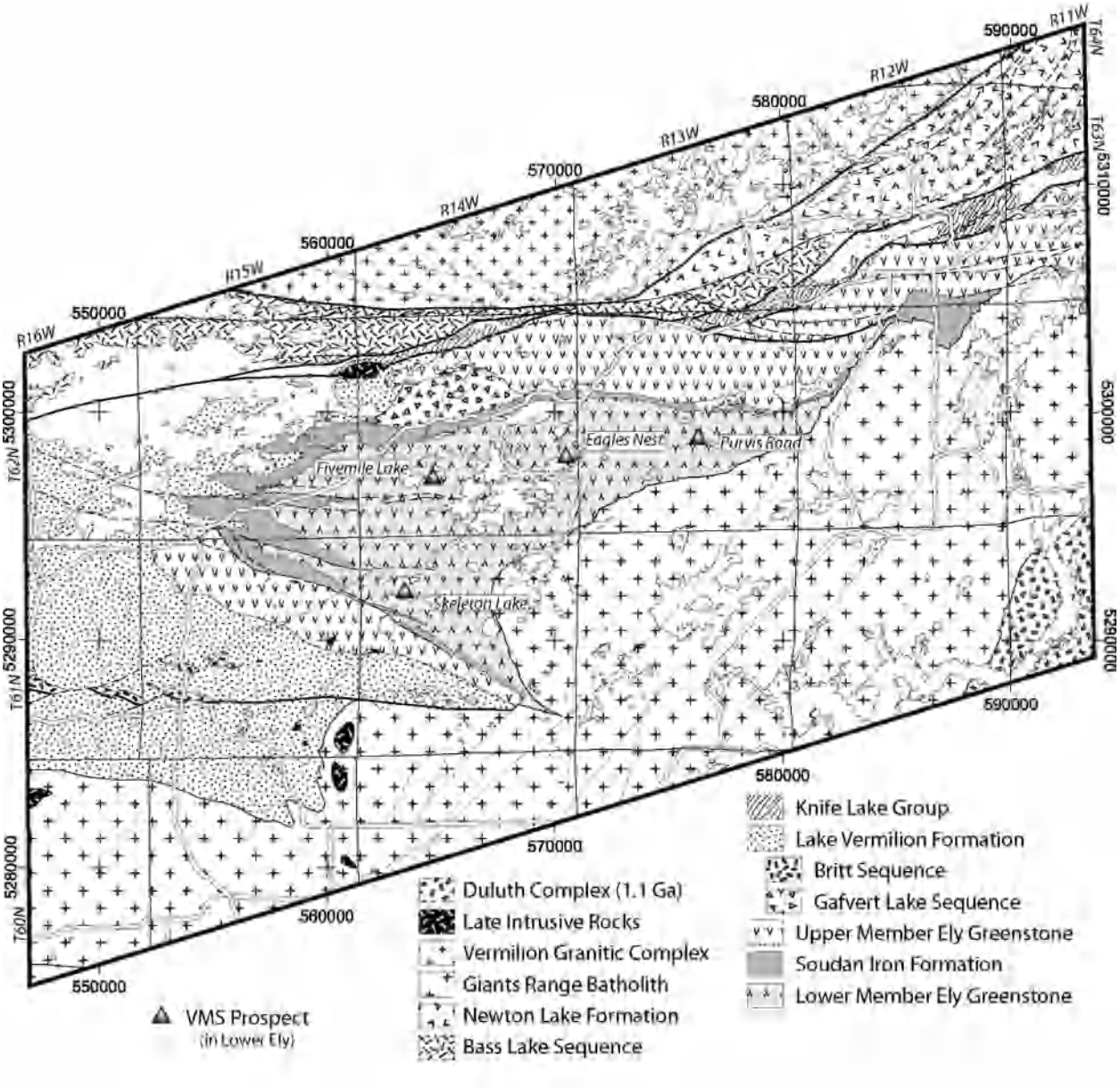


Figure 1-2. Simplified lithostratigraphic map of the western Vermilion district with the location of VMS prospects in the Lower Member of the Ely Greenstone (from Peterson and Jirsa, 1998a). UTM Nad 83 coordinates in meters.

Table 1-1. Lithostratigraphic units within the western Vermilion district (from Peterson and Jirsa, 1998a).

Intrusive Rocks	
Late Intrusions	Plutons and stocks of syenite, monzonite, diorite, and lamprophyre
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 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, & sheared equivalents
Lake Vermilion Formation	Graywacke, slate, dacitic tuff, and minor conglomerate
Gafvert Lake Sequence	Dacitic to trachyandesitic lava flows, tuffs, and intrusions
Britt Sequence	Tholeiitic basalt lava flows
Upper Ely Greenstone	Tholeiitic basalt lava flows and iron-formation
Soudan Iron Formation	Layered cherty iron-formation, epiclastic rocks, and tuff
Lower Ely Greenstone	Calc-alkalic & tholeiitic basalt-rhyolite lava flows, tuffs, epiclastic rocks and minor iron-formations

The Lower Ely and Soudan Iron Formation (Morey et al., 1970) comprise the strata we will be investigating during this field trip. The Lower Ely consists dominantly of volcanic arc-associated, pillowed and massive basalt and andesite flows of calc-alkalic and tholeiitic composition (Fig. 1-3). Hypabyssal diabase, gabbro, and diorite sills, isolated dacite, rhyodacite, and rhyolite lava flows, domes, tuffs, lapilli tuffs, and tuff-breccias (Fig. 1-4), and local chemical, volcaniclastic and epiclastic sedimentary strata occur throughout the sequence. Pillows in the lowermost sections of the Lower Ely (named the Fivemile Lake sequence by Peterson and Patelke, 2003) typically are irregularly shaped and well-vesiculated, indicating moderate to shallow water depths of formation (Schulz, 1982; Peterson, 2001; Hudak et al., 2002a). Rare massive to bedded scoria deposits interstratified with the pillowed flows also indicate shallow subaqueous, and perhaps locally, subaerial volcanism (Peterson, 2001; Hudak et al., 2002a). Pillowed basaltic rocks in the uppermost parts of the Lower Ely (named the Central Basalt Sequence by Peterson and Patelke, 2003) typically are sparsely-vesiculated and ovoid, and scoria-rich primary and reworked volcaniclastic rocks are not prevalent (Hudak et al., 2002b; Hudak et al., in prep. a; Peterson and Patelke, 2003). These features suggest a general subsidence of the volcanic pile to deep-water conditions.

Stratigraphically overlying the volcanic rocks of the Lower Ely is the Soudan Iron Formation, which consists dominantly of laminated Algoma-type iron-formation, with interstratified basalt lava flows and volcaniclastic and epiclastic strata derived from basaltic to dacitic volcanic strata. A gradational contact over several tens to hundreds of meters occurs between the underlying Lower Ely and the overlying Soudan Iron Formation member of the Ely Greenstone (Hudak et al., 2002b; Peterson and Patelke, 2003). In general, the exhalative nature of many of the rocks within the Soudan Iron Formation member represent deep-water chemical deposition throughout a period of quiescence, which began during the latest stages of volcanism associated with the Lower Ely. The stratigraphic thickness of the Soudan Iron Formation member varies from 50 to 3,000 meters, and averages approximately 700 meters. The thickest sections occur in the nose of the Tower-Soudan anticline (Fig. 1-2), and probably represent stratigraphy that is repeated by shearing and thickened by folding.

The time period over which the Soudan Iron Formation was deposited is poorly constrained, as geochronological data is very limited within the Ely Greenstone. Nevertheless, the size (thickness and strike length) of the Soudan Iron Formation is much larger than typical Algoma-type iron-formations, and

its formation probably occurred during a profound break in effusive and pyroclastic volcanism that was accompanied by a long-lived period of hydrothermal activity. The upper contact of the Soudan Iron Formation is more diverse; overlying stratigraphic units along strike include the Upper Ely, the Gafvert Lake Sequence, and the Lake Vermilion Formation.

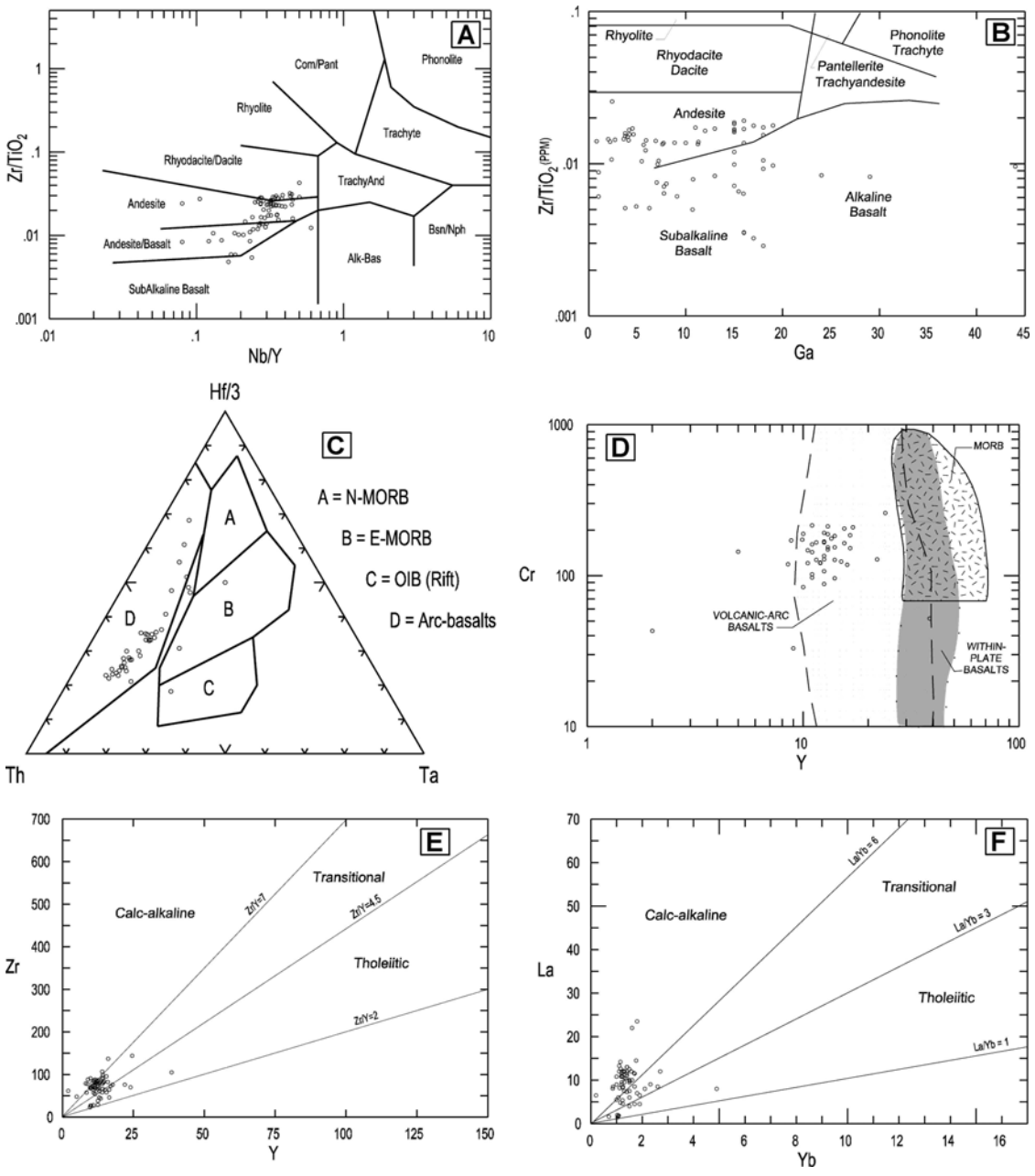


Figure 1-3. Lithogeochemical characteristics of mafic and intermediate volcanic and volcaniclastic rocks of the Lower member of the Ely Greenstone in the vicinities of Fivemile, Needleboy, and Sixmile Lakes (after Hudak et al., 2002a; Hudak et al., in prep. a). Figures 1-4a and 1-4b modified from Winchester and Floyd (1977), Figure 1-4c modified from Wood (1980), Figure 1-4d modified from Winchester and Floyd (1976), Figures 1-4e and 1-4f modified from Barrett and MacLean (1999).

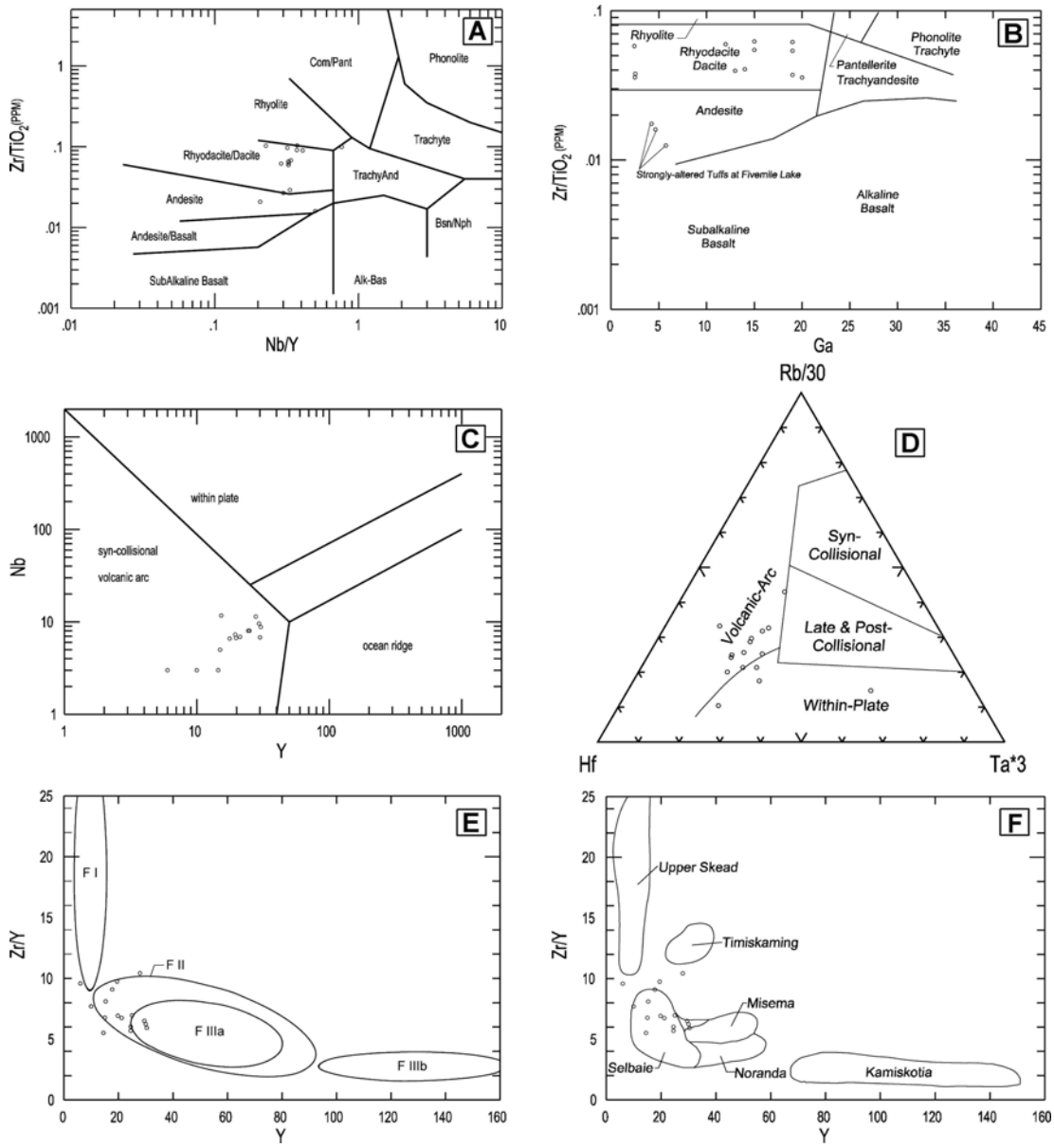


Figure 1-4. Lithogeochemical characteristics of felsic volcanic and volcaniclastic rocks of the Lower member of the Ely Greenstone in the vicinities of Fivemile, Needleboy, and Sixmile Lakes (after Hudak et al., 2002a; Hudak et al., in prep. a). Figures 1-5a and 1-5b modified from Winchester and Floyd (1977), Figure 1-5c modified from Pearce et al. (1984), Figure 1-5d modified from Harris et al. (1986), Figures 1-5e and 1-5f modified from Piercey et al. (2001) and Barrie et al. (1993).

STRUCTURAL GEOLOGY

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 (D_1) produced broad, locally recumbent folds within the Soudan belt and major fault zones throughout the region. In the

Newton belt, D_1 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 D_1 deformation (Jirsa, 2000). A large, map-scale structure related to D_1 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 (S_1) locally.

A second deformation event (D_2) associated with synchronous regional metamorphism resulted in foliation development and structures having largely dextral asymmetry. D_2 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 D_2 deformation affected all of the supracrustal rocks in the area and is reasonably constrained by geochronology, the regional foliation (S_2) can be used in the field to temporally relate other structural, intrusive, and deformation events. The relationship between S_2 fabric and shear structures indicates that most shearing occurred relatively late in the D_2 event. Major shearing that produced the Mud Creek and related shear zones (e.g., the Murray shear zone, Fig. 1-1) is attributed to the late stages of D_2 dextral transpression.

Structures related to the third deformation event (D_3) include abundant NE- and NW-trending faults that dissect the stratigraphic assemblages. Named structures related to D_3 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 north of the Soudan Mine (Fig. 1-1).

VMS PROSPECTS AND EXPLORATION HISTORY

Since the mid-1860's, numerous mineral exploration programs have been conducted in the Vermilion district. Most of these exploration programs focused on identifying minable deposits of massive hematitic iron ores, such as those mined between 1883 and 1962 in the Soudan Iron Formation member at the Soudan Mine. During the 1980s and early 1990s, subeconomic lode gold mineralization was discovered in close proximity to the east-west-trending Murray shear zone, which dissects the Lower Ely, and in close proximity to the Mud Creek shear zone, which separates the Soudan belt from the Newton belt to the north (Fig. 1-1).

Four VMS prospects (Fig. 1-2) occur within the Lower Ely, and occur in close proximity up-section from a semiconformable quartz-epidote alteration zone that extends for at least 19km along strike in the north limb of the Tower-Soudan Anticline (Peterson, 2001). These include the Skeleton Lake prospect (drilled by Exxon, 1972), the Eagles Nest prospect (drilled by Newmont, 1988), the Fivemile Lake prospect (drilled by Teck, 1994), and the Purvis Road prospect (drilled by Rendrag, 1999). The stratigraphy, hydrothermal alteration, and mineralization associated with the Skeleton Lake, Eagles Nest, and Fivemile Lake prospects have been recently re-evaluated (Hudak and Morton, 1999; Hudak et al., 2002a; Hudak et al., 2002b). Detailed studies of the Fivemile Lake, Eagles Nest, and Skeleton Lake prospects indicate that copper-zinc sulfide mineralization occurs proximal to relatively narrow (10s of meters wide) disconformable zones of chlorite and/or sericite alteration within felsic volcanic and volcaniclastic rocks (Hudak and Morton, 1999; Hudak et al., 2003). Hudak et al. (2003) and Hudak et al., (in prep. a) have noted that these disconformable alteration zones and associated mineralization commonly occur near high concentrations of synvolcanic mafic dikes, suggesting that these disconformable hydrothermal alteration zones occurred within synvolcanic fault zones (c.f. Gibson et al., 1999).

VMS-ASSOCIATED VOLCANIC AND HYDROTHERMAL ALTERATION PROCESSES

The composition and distribution of hydrothermal alteration mineral assemblages in the Lower Ely (Table 1-2) is similar to that described in major lava-flow dominated VMS mining districts worldwide (e.g. the Noranda Camp, Quebec; Morton and Franklin, 1987; Franklin, 1996; Gibson et al., 1999; Hudak and Morton, 1999; Peterson, 2001; Hudak et al., 2002; Hudak et al., in prep. a). Results of recent studies (Peterson, 2001; Odette et al., 2001a, 2001b; Hudak et al., 2002; Hudak et al., in prep. a) indicate that not only are the compositions and geometries of the regional alteration mineral assemblages identical to those present in many lava flow dominated massive sulfide mining districts, but that detailed alteration mineral chemistries (Hocker et al., 2003) are also consistent with those associated with the VMS ore deposits in these mining camps. These two observations suggest that the processes that formed the alteration mineral assemblages in the Lower Ely were similar to those that formed equivalent alteration zones in well-established VMS mining camps. Thus, one of the most perplexing questions plaguing economic geologists is why economically significant VMS deposits have not yet been discovered in the Vermilion district.

A general genetic model for the formation of VMS deposits and associated hydrothermal alteration zone, as recently presented by Franklin et al. (1998), requires convective metalliferous hydrothermal fluid generation in the sub-seafloor environment via heating of down-welling seawater and leaching of metals from the enclosing volcanic and sedimentary strata (Fig. 1-5). The size of a convective hydrothermal system is a function of the abundance of heat in the upper two kilometers of the sub-seafloor crust (Franklin, 1996; Franklin et al., 1998). The intrusion of hypabyssal synvolcanic dikes and/or sills into the shallow sub-seafloor may vigorously enhance the dynamics of convective hydrothermal cells (Campbell et al., 1981). On reaching a critical reaction temperature of ~ 350°C, sustained acid pH in the hydrothermal fluid (evolved fluid) is achieved, and metals are leached from the rocks into the evolved fluid via primary mineral breakdown by calcium metasomatism, silicification, and hydrolysis reactions (Seyfried et al., 1999). In basalt-dominated systems (such as that in the Lower Ely), leaching-related alteration of mafic "source" zones (lower semi-conformable alteration) forms a mineral assemblage composed of albite-epidote-zoisite/clinozoisite-actinolite-quartz. These zones are variably metal-depleted, and are characterized by patchy silicification and epidotization associated with areas metasomatically enriched in silica and calcium.

In lava flow-dominated stratigraphic sequences, regionally confined discordant “pipe-like”, and more regionally extensive “semiconformable” alteration zones are present (Morton and Franklin, 1987). The “pipe-like” semi-conformable alteration zones are closely associated with zones of cross-stratal permeability (e.g. synvolcanic fault zones), and are characterized by well-defined vertically extensive alteration zones containing anomalous abundances of sericite, chlorite (both Fe- and Mg-rich varieties), actinolite/ferroactinolite, quartz, pyrite, and locally, chalcopyrite and/or pyrrhotite. Semiconformable alteration zones extend for several kilometers to tens of kilometers in the rocks stratigraphically beneath and adjacent to VMS mineralized horizons (Santaguida et al., 2002a; Santaguida et al., 2002b). In mafic-dominated volcanic environments, such alteration typically is associated with regional zones of spilitization (an alteration assemblage composed of albite + quartz + Mg-rich chlorite ± sericite), silicification (quartz ± albite), and epidote-quartz alteration (epidote + quartz ± actinolite ± carbonate) (Morton and Franklin, 1987; Gibson et al., 1999; Santaguida et al.. 2002a, Santaguida et al., 2002b). Regional semiconformable alteration zones in felsic rocks in VMS producing camps such as Noranda (Quebec) or Sturgeon Lake (Ontario), typically comprise extensive zones of spilitization, silicification, and sericitization (sericite + quartz ± Mg-rich chlorite) (Morton and Franklin, 1987; Gibson et al., 1999).

Table 1-2. Greenschist facies metamorphosed hydrothermal alteration mineral assemblages recognized in the Lower Ely (Hudak et al., 2002a; Hudak et al., in prep. a).

Alteration Mineral Assemblage	Geometry
No apparent alteration (unaltered)	Not applicable
Least altered	Semiconformable
Silicified	Semiconformable
Quartz + Chlorite (chlorite > 15%)	Semiconformable
Epidote + Quartz ± Chlorite ± Albite (Epidote >10%, Actinolite <10%, Fe-chlorite <20%)	Semiconformable
Epidote + Quartz ± Chlorite ± Albite (Epidote >10%, Actinolite <10%, Fe-chlorite >20%)	Semiconformable
Albite + Quartz ± Epidote ± Actinolite ± Chlorite (Epidote <10%, Actionlite <10%)	Semiconformable
Secondary Plagioclase Feldspar ± Chlorite ± Sericite	Semiconformable
Biotite ± Fe-/Mg-chlorite	Semiconformable
Iron Oxide / Hematite	Semiconformable
Epidote + Quartz + Actinolite ± Chlorite ± Albite (Epidote>10%, Actinolite > 10%)	Semi/Disconformable
Mottled Epidote +Quartz ± Actinolite ± Chlorite ± Albite ("Patchy Epidosite")	Semi/Disconformable
Actinolite ± Fe-chlorite ± Epidote ± Albite (Actinolite >10%, Epidote <10%)	Disconformable
Fe-chlorite + Sericite (Fe-chlorite >10%, Sericite >5%)	Disconformable
Fe-chlorite + Sericite (Fe-chlorite >10%, Sericite<10%, Fe-chlorite + Sericite >25%)	Disconformable
Fe-chlorite + Fe-carbonate ± Sericite ± Actinolite ± Epidote (Fe-chlorite >25%, Fe-carbonate > 5%, Actinolite, Sericite, Epidote < 5%)	Disconformable
Sericite ± Pyrophyllite ± Fe-chlorite (Sericite/Pyrophyllite >5%, Fe-chlorite <10%)	Disconformable
Sericite ± Carbonate (Sericite > 5%, calcite/dolomite/ankerite present, chlorite absent)	Disconformable
Mg-chlorite ± sericite	Disconformable
Chlorite + Fe-carbonate ± Sericite	Shear Zones
Sericite + Green Mica ± Carbonate	Shear Zones
Tourmaline + quartz	Veins

Both discordant and semiconformable alteration zones have been discovered in the Vermilion district (Table 1-2), and have been described by Hudak and Morton (1999), Peterson (2001), Odette et al. (2001), and Hudak et al. (2002a). Semiconformable alteration zones in the Lower Ely are dominated by mineral assemblages containing various proportions of quartz, epidote, zoisite/clinozoisite, Fe-chlorite, Mg-chlorite, actinolite, ferroactinolite, sericite/pyrophyllite, and albite. Odette et al. (2001a, 2001b) and Hudak et al. (2002a) have shown via mass balance analysis that semiconformable quartz + epidote ± actinolite ± albite ± chlorite alteration mineral assemblages in the Fivemile Lake area are metasomatically enriched in calcium and silica, and depleted in base metals (copper and zinc) by 50-90%. Pipe-like, northeast-trending disconformable alteration zones in the Lower Ely are largely composed of Fe-rich chlorite, sericite/pyrophyllite, actinolite and/or ferroactinolite. Pipe-like alteration zones that have been mapped up-section have, to date, not led to the discovery of economically significant VMS deposits, but have been instrumental in locating potential base metal sulfide-bearing stratigraphic horizons and localized chemical exhalites. The mineralogy (Hudak et al., 2002a) and chemical compositions of the alteration minerals (Hocker et al., 2003) in these disconformable zones are consistent with those associated with economic VMS deposits in both Canada and Scandinavia (c.f. Hannington et al., 2002).

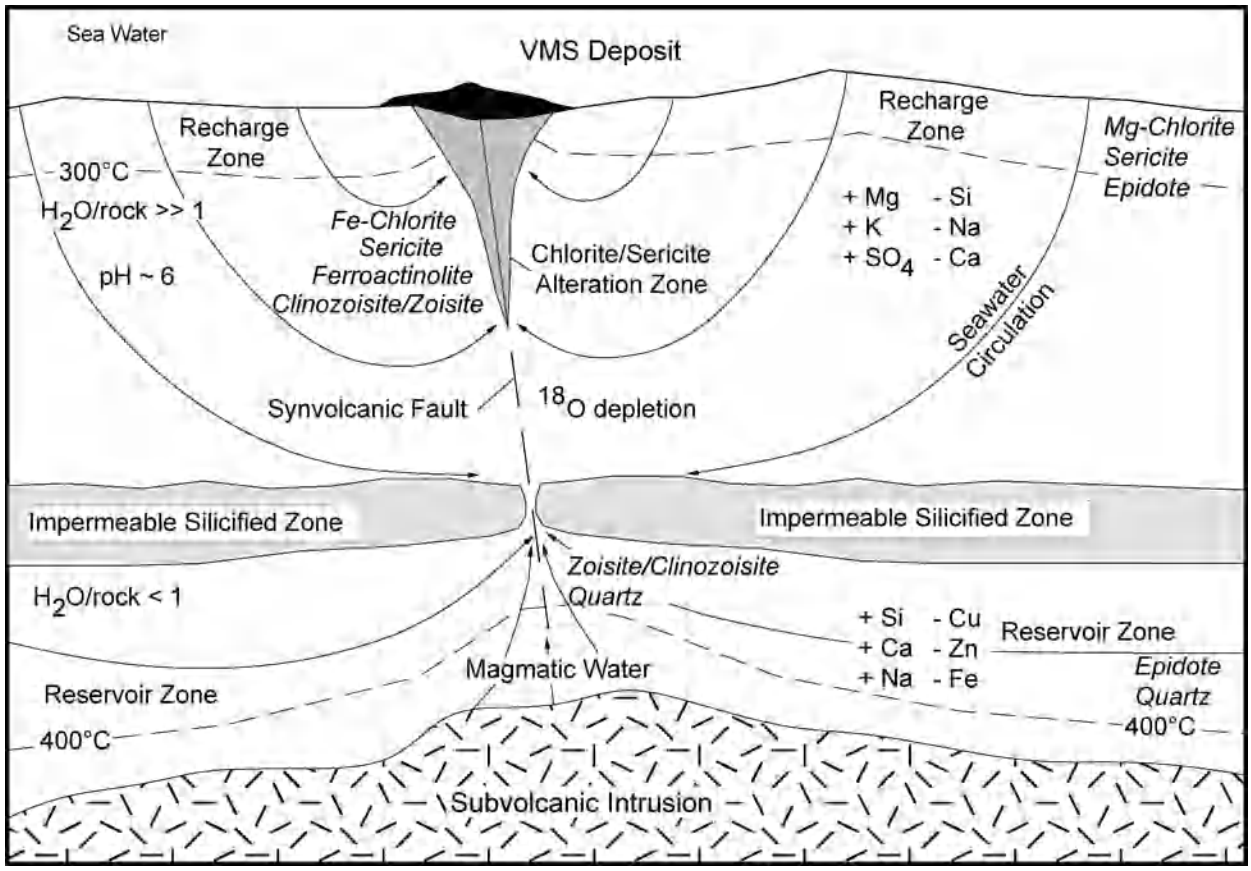


Figure 1-5. Simplified schematic model of a convective hydrothermal system associated with the formation of Noranda–type (Morton and Franklin, 1987) or lava-flow dominated-type (Gibson et al., 1999) VMS deposits (modified from Franklin , 1996).

Water depth controls aspects of associated VMS deposit composition (metal ratios) and alteration assemblages (Franklin, 1986; Franklin, 1993). Field evidence from the recent studies in the Lower Ely (Peterson, 2001; Hudak et al., in press) indicates that both shallow and deep subaqueous environments were present in the Vermilion district, as represented by the Fivemile Lake and Central Basalt Sequences. Such environments are not only associated with base-metal VMS deposits, but may also contain precious metal-rich (e.g. gold, silver) VMS deposits (Sillitoe et al., 1996; Hannington et al., 1999). Thus, the Vermilion district, and in particular, the rocks within a few kilometers of the area we are investigating on this field trip, may contain VMS deposits with a variety of base- and precious metal compositions.

FIELD TRIP STOPS – DAY 1 (MAY 4)

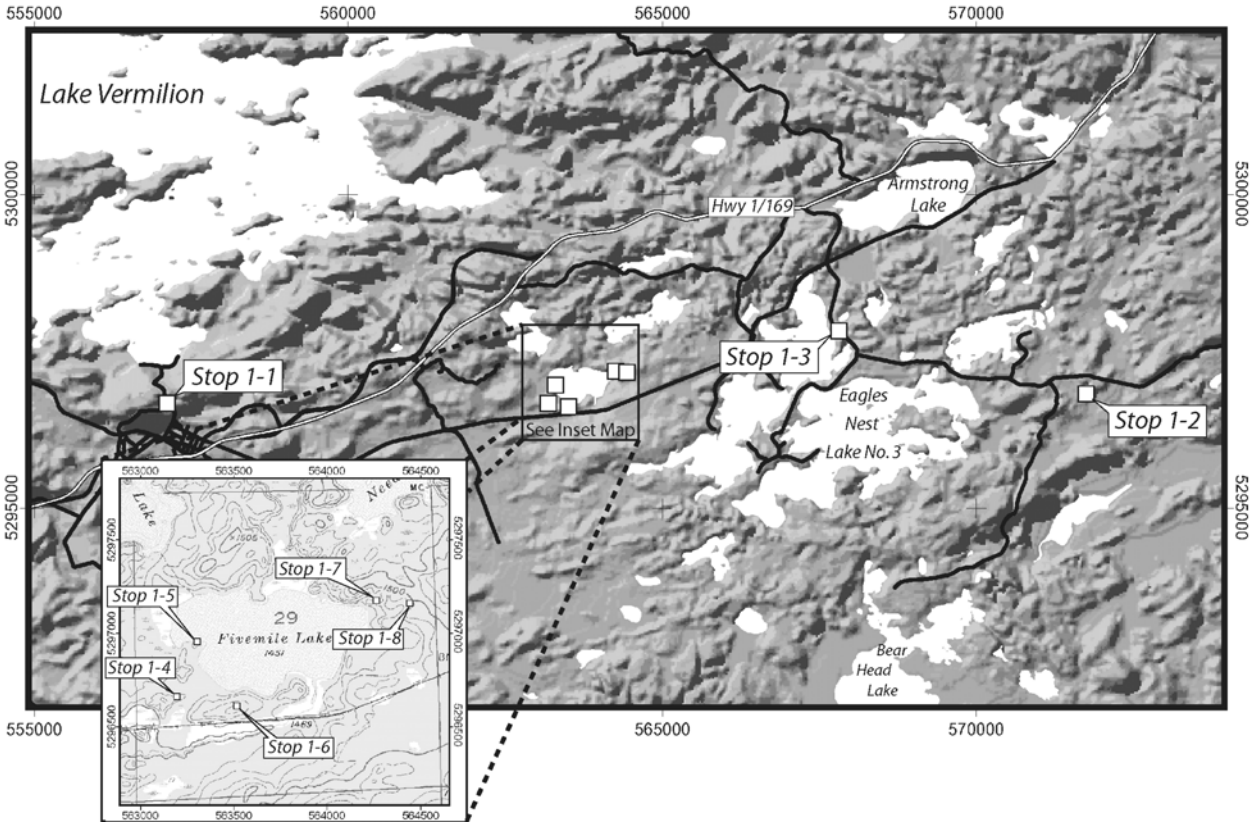


Figure 1-6. Field trip stop locations for Day 1 of the Vermilion district field trip.

STOP 1-1: **ARCHEAN SOUDAN IRON FORMATION** ***No hammering please!***

Location: (NE, NE, S. 27, T.62N., R.15W., NAD83 UTM 557120E, 5296660N)

Description: This classic exposure of the Soudan Iron Formation (Fig. 1-6) lies on the north limb of the Tower-Soudan anticline, and at the stratigraphic top of the volcanic sequences known collectively as the Lower member of the Ely Greenstone—the focus of this field trip. In a general way, the upper kilometer or so of strata assigned to the Lower Ely displays an upward-stratigraphic progression marked by increasing numbers and thicknesses of iron-formation-rich units, and corresponding decreasing thicknesses of the interleaved basaltic flows. This represents a general cessation of volcanism over some time. It is interesting to observe the rhythmic microlaminations (1 mm or so thick) in various cherty beds exposed here and speculate about the paleoenvironment—that is, whether these represent daily heating/cooling cycles, tidal, climatic, annual, or some other repetitive influence in the depositional environment. What is known about iron-formation units in the Ely Greenstone is that their deposition occurred during periods of relative volcanic and tectonic quiescence by the slow, subaqueous "rain" of chemical precipitates. The Soudan marks an abrupt transition from a period of waning basaltic volcanism in comparatively deep water, to one of mixed, locally emergent volcanic and associated sedimentary deposition represented by exposures to the north and west. The latter include dacitic rocks of what is

known as the Gafvert Lake sequence, and tuffaceous graywacke of the Lake Vermilion Formation. Whether the transition at Soudan represents a significant unconformity remains enigmatic.

This exposure also represents a microcosm of deformation in the western Vermilion district. It displays two generations of close folding in delicate laminations of chert (creamy white), chert-hematite jasper (red), and magnetite-chert (black to silver-colored). The second generation of folds (F_2) is tectonic in origin, having subvertical axial surfaces that trend eastward, and steeply plunging axes. Most display Z-asymmetry. The earlier folds (F_{0-1}) appear to have been sharply refolded to produce complex interference patterns. Lundy (1985) studied folding at this locality and concluded that some of the apparent interference structures are the product of early-formed sheath folds that did not involve refolding by D_2. The F_1 structures are predominantly intrafolial, and exhibit a great variety of style and orientation; implying they formed by layer-parallel, soft-sediment slumping. Work by Hudleston (1976) and Jirsa et al. (1992) extends these types of observations to regional tectonic interpretations of an early (D_1) folding that included the development of the Tower-Soudan anticline in rigid volcanic complexes, and associated large nappe structures in the more ductile marginal sedimentary strata, followed by D_2 metamorphism and transpressive deformation.

STOP 1-2: XENOLITHIC HORNBLENDE DIORITE, PURVIS PLUTON

Location: (NE, SW, S. 30, T.62N., R.13W., NAD 83 UTM 571755E, 5296815N)

Description: The Purvis pluton is an east-west trending, moderate-sized (~3km^3), sill-like multiphase dioritic to tonalitic intrusion with a strike length of 5.7 km and a thickness that ranges from 100-1200 meters (Peterson, 2001). This intrusion occurs in the lower stratigraphic section of the north limb of the Tower-Soudan anticline (Peterson and Jirsa, 1999a; Jirsa et al., 2001). Recent work by Drexler and Hudak (in press) indicate that the intrusion has several phases, including 1) xenolithic hornblende diorite; 2) xenolithic hornblende tonalite; 3) xenolithic leucotonalite; 4) leucotonalite and trondhjemite; and 5) leucotonalite dikes (Fig. 1-7).

Detailed field mapping by Peterson (2001), Hovis (2001) and Drexler and Hudak (in press) suggest that the Purvis pluton is a synvolcanic intrusion based on the following characteristics: 1) it lacks a contact metamorphic aureole; 2) its uppermost contact is proximally associated with intense, semiconformable quartz + epidote alteration zones; 3) D_2 deformation fabrics occur in both the intrusion and the surrounding country rocks; and 4) early xenolithic diorite phases are cross-cut by thin, commonly D_2-deformed dikes of younger tonalite and trondhjemite phases. Galley (2002) and Galley (2003) have indicated that these characteristics are key features of synvolcanic intrusions temporally associated with the genesis of many Precambrian VMS deposits. Peterson (2001) has suggested that the Purvis pluton may have been the heat source that drove hydrothermal systems that produced the Eagles Nest and Purvis Road VMS prospects. A sample of the leucotonalite phase has been submitted to Dr. Daniel Holm (Kent State University) for geochronological analysis, but this age is not yet available. Based on the presence of the D_2 fabric in the pluton, U/Pb zircon dates of a D_2-deformed felsic lava dome in the Lower Ely in the Fivemile Lake area, and a post-D_2 quartz feldspar porphyry in the Newton Belt (Peterson et al., 2001), we anticipate an age between 2683 Ma and 2722 Ma for the Purvis pluton.

This locale offers an opportunity to investigate the xenolithic hornblende diorite phase of the Purvis pluton. The outcrop adjacent to the road predominately contains four types of xenoliths: 1) dark green xenoliths of amygdaloidal (5-8%) basalt-andesite pillow lavas which locally have preserved selvedges and interpillow hyaloclastite, and are locally contact metamorphosed along their margins; 2) pale green epidote + quartz-altered basalt-andesite lava xenoliths that are up to 15 cm in diameter; 3) rare coarse-grained gabbro/diorite xenoliths up to 3 cm in diameter; and 4) rare <1-2 cm diameter subangular

chert xenoliths. Large iron-formation xenoliths up to several meters in diameter, and amphibolite xenoliths up to several centimeters in diameter, may be observed in xenolithic hornblende tonalite outcrops located east of this location (NAD 83 UTM coordinates 0574920E, 5297600N).

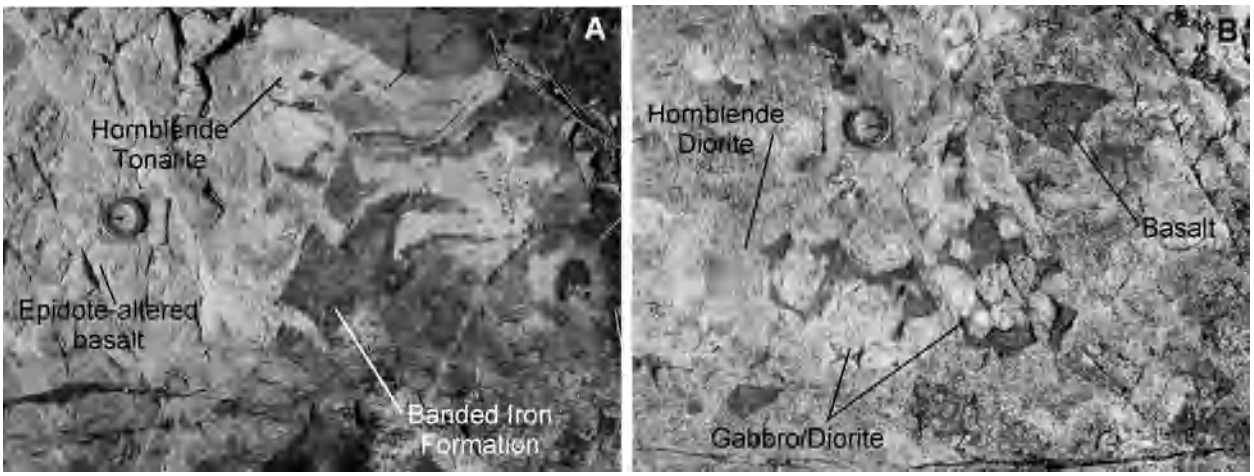


Figure 1-7. Xenolithic hornblende tonalite (A) and xenolithic hornblende diorite (B) phases of the Purvis pluton..

Iron-formation xenoliths present in outcrops east of here were likely derived from iron-formation horizons that occur immediately southwest of Purvis Lake. Basalt and altered basalt fragments also were derived from the surrounding Lower Ely. The presence of epidote-quartz altered mafic xenoliths suggests that this phase of the Purvis pluton stoped its way upward into an earlier-formed proximal zone of quartz-epidote alteration formed from high temperature seawater-rock interaction (e.g. Galley, 2003). Amphibolite xenoliths are believed to represent contact metamorphosed basalt fragments based on petrographic similarities (Drexler and Hudak, in press). Drexler and Hudak (in press) have shown that coarse-grained gabbro/diorite fragments likely represent xenoliths of the earliest phases of the pluton.

A short hike up to the top of the ridge allows investigation of a newly exposed (and still lichen-free) outcrop displaying the complicated textures between medium-grained diorite and xenoliths in this phase of the Purvis pluton. Collectively, six types of xenoliths comprise 55-60% of the intrusion at this location. These xenolith types include: 1) pale to dark green, subangular to subround basalt-andesite clasts which range from 1-30cm in diameter and make up 15-20% of the rock; 2) subround to subangular, coarse-grained gabbro/diorite xenoliths that are up to 20cm in diameter; 3) subround epidote + quartz-altered basalt-andesite xenoliths that are up to 20cm in diameter; 4) subangular, dark grayish-green fine-grained diorite xenoliths which are up to 10cm in diameter; 5) dark green, coarse-grained, actinolite-rich amphibolite fragments up to 12 cm in diameter; and 6) rare pillow selvedge / interpillow hyaloclastite xenoliths which locally make up 1% of the unit.

Moving east along the ridge, one can observe additional outcrops of xenolithic diorite. Locally (UTM NAD 83 coordinate 571798E, 5296828N) ENE-trending 10-20cm wide leucotonalite dikes cut through the xenolithic diorite phase, indicating the age relationships between these two units. Farther east along the ridge, the complicated contact between the Purvis pluton and the overlying basalt-andesite pillow lavas can be observed. These lavas contain patchy epidote + quartz alteration, and have locally been metamorphosed to hornfels immediately adjacent to the intrusion.

Studies of ancient VMS deposits has documented that the deposits commonly occur in depressions on the paleo-seafloor (3rd-order basins) while modern deposits on the seafloor are found on high-standing structures, such as ridges. These differences are probably more apparent than real, in that

the modern deposits are generally confined to the axial graben, or depression, of what are otherwise are high-standing structures. In addition, both ancient and modern deposits occur in areas of anomalously high heat flow, generally linked to synvolcanic intrusions beneath the hydrothermal systems. The recent mapping in the Purvis Road area has shown the presence of all of the attributes of typical VMS-forming hydrothermal systems. These attributes include a synvolcanic intrusive heat source (the Purvis pluton), a paleotopographic high-standing structure, VMS-style alteration mineral assemblages, and the presence of Cu and Zn-rich massive sulfide (recent logging in this area has exposed numerous angular boulders of massive sulfide in the basal till). The VMS attributes of the Purvis road area are presented in Figure 1-8.

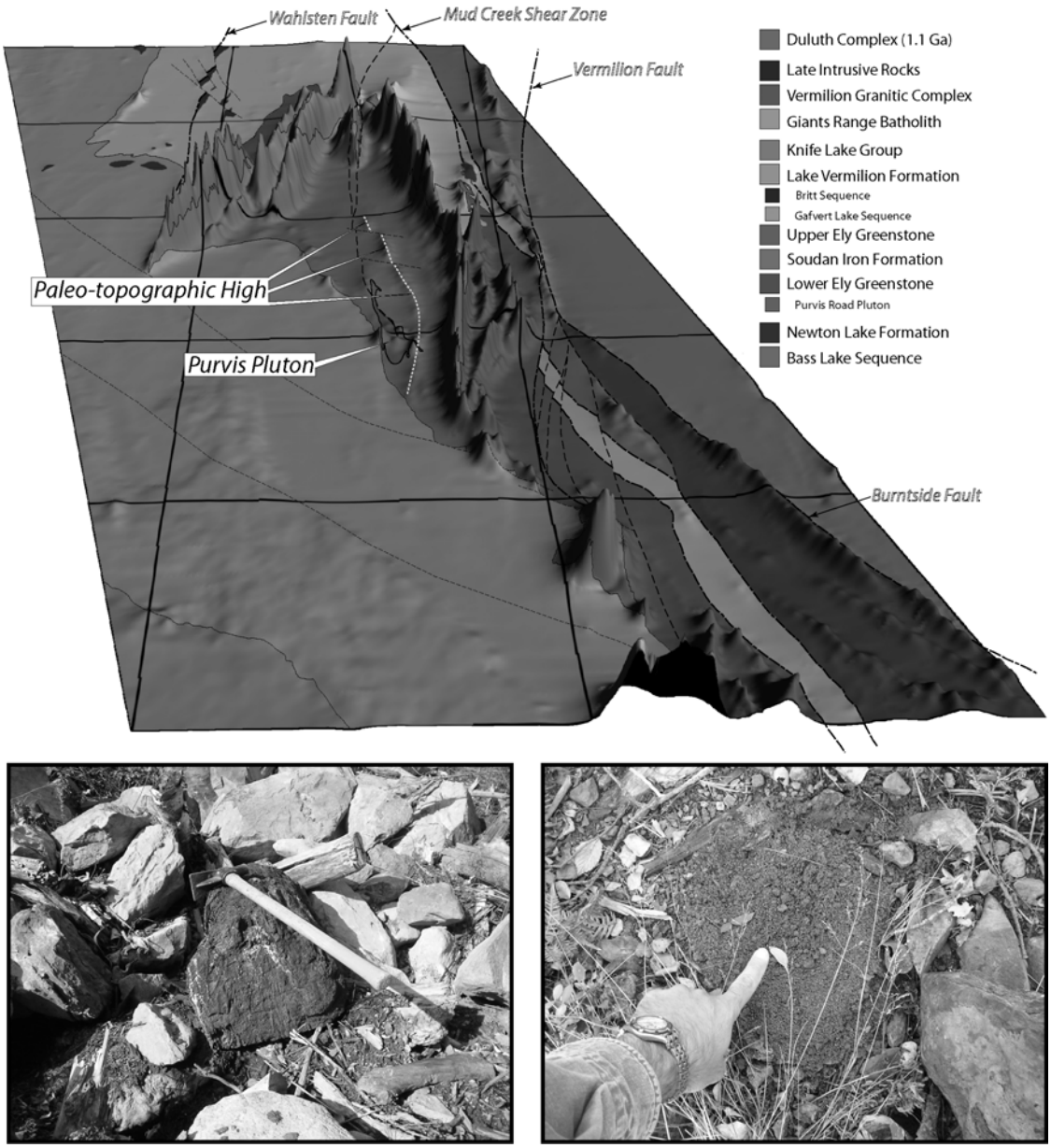


Figure 1-8. Simplified regional lithostratigraphic sequences draped on a 3D view (looking due west) of a residual field aeromagnetic anomaly map of the western Vermilion district (same map area as Fig. 1-2). The lower two photographs show massive sulfide boulders exposed north of the Purvis pluton.

STOP 1-3. SILICIFIED FIVEMILE LAKE SEQUENCE PILLOW LAVAS / REGIONAL SEMICONFORMABLE ALTERATION

Location: (SE, SE, S. 22, T.62N., R.14W., NAD83 UTM 567818E, 5297818N)

Description: Geological mapping by Peterson (2001) has indicated the presence of a regional semiconformable quartz-epidote alteration zone extending for at least 19 km along strike within the Lower Ely along the north limb of the Tower-Soudan anticline. This type of alteration is a common feature in many Archean VMS camps (e.g. Noranda (Gibson, 1989) and Snow Lake (Skirrow and Franklin, 1994)), and is attributed to silica- and calcium-dumping that occurs in the deep, sub-seafloor as downwelling hydrothermal fluids are heated to temperatures in excess of 350°C (Franklin, 1986; Franklin, 1993). Semiconformable alteration zones associated with VMS systems are generally much larger in area than their associated mineralization, and therefore provide exploration geologists regional areas in which to concentrate more detailed, follow-up field mapping, geochemical studies, and geophysical surveys for identifying VMS targets.

Semiconformable quartz-epidote alteration zones are commonly metal-depleted. Peterson (2001), Odette et al. (2001a, 2001b), and Hudak et al. (2002a) have shown that that quartz-epidote altered basalts within this regional semiconformable alteration zone are commonly depleted in base metals by as much as 50-90%. Peterson (2001) completed simplified mass balance calculations of Cu and Zn depletion associated with the regionally extensive quartz-epidote alteration zone within the Lower Ely, and this work reveals the importance of this alteration to the VMS mineral potential of this area. The surface expression of the quartz-epidote alteration is approximately 19 kilometers long, and averages approximately 1.5 kilometers in width. The area of this zone is on the order of 30.5 km^2. Using Southwick et al. (1998) values for Cu and Zn of 59 and 115 ppm, and a mean density of the rocks equal to 2.8g/cm^3, the total contained copper and zinc for various volumes of rock from the Lower Ely are calculated, and presented in the top of Table 1-3. These volumes include a 1.0 kilometer thick slab (volume = 30.5 km^3), a 2.0 kilometer thick slab (volume = 61 km^3), a 3.0 kilometer thick slab (volume = 91.5 km^3), and a circular disk (volume = 591 km^3). Percent leaching calculations for copper and zinc are presented in the lower portions of Table 1-3, with values of metal leached (in pounds) from the four calculated volumes of rock. The leaching models include 5%, 10%, 25%, and 50% metal leached from the rock. Given the patchy nature of this alteration, an estimate of somewhere between 10 and 25% metal leaching seems reasonable, as does the 91.5 km^3 volume. Therefore, one can estimate that many billions of pounds of copper and zinc have been leached out of the volcanic rocks in the north limb of the Lower Ely Greenstone. The concentration of these base metals into specific sites by a VMS style hydrothermal system would have created significant tonnages of massive sulfide. This data clearly reflects the potential for VMS mineralization up-section from this zone.

At this location we can observe part of the regionally extensive quartz-epidote alteration zone. The outcrop contains relatively undeformed bun- and mattress-shaped pillows. Interpillow hyaloclastite zones are generally pale to dark green in color, and are chlorite and/or actinolite-rich. Minor red-brown staining locally occurs in these zones, and is indicative of the presence of trace amounts of pyrite and/or chalcopyrite. Pillow selvedges commonly contain up to 10% round to oval, pipe-like quartz-epidote and/or actinolite chlorite amygdules. The cores of the pillows are typically pale green gray in color due to nearly wholesale replacement of the original igneous minerals by quartz and epidote.

Table 1-3. Copper and zinc leaching calculations for the regional semiconformable quartz-epidote alteration zone present in the Lower Ely (Peterson, 2001).

	Volume (Km^3)			
	30.5	**61**	**91.5**	**591**
Total Contained Cu (lbs)	11,084,920,000	22,169,840,000	33,254,760,000	214,793,040,000
Total Contained Zn (lbs)	21,606,200,000	43,212,400,000	64,818,600,000	418,664,400,000
5% Cu leached (lbs)	554,246,000	1,108,492,000	1,662,738,000	10,739,652,000
10% Cu leached (lbs)	1,108,492,000	2,216,984,000	3,325,476,000	21,479,304,000
25% Cu leached (lbs)	2,771,230,000	5,542,460,000	8,313,690,000	53,698,260,000
50% Cu leached (lbs)	5,542,460,000	11,084,920,000	16,627,380,000	107,396,520,000
5% Zn leached (lbs)	1,080,310,000	2,160,620,000	3,240,930,000	20,933,220,000
10% Zn leached (lbs)	2,160,620,000	4,321,240,000	6,481,860,000	41,866,440,000
25% Zn leached (lbs)	5,401,550,000	10,803,100,000	16,204,650,000	104,666,100,000
50% Zn leached (lbs)	10,803,100,000	21,606,200,000	32,409,300,000	209,322,200,000

STOP 1-4: WELL-PRESERVED PILLOW LAVAS, SOUTH OF FIVEMILE LAKE

Location: (NE, SW, S. 29, T.62N., R.14W., NAD83 UTM 563195E, 5296660N)

Description: The Fivemile Lake Prospect comprises an extremely well preserved, and locally intensely hydrothermally altered, bimodal package of basalt-andesite pillows lavas and coherant and volcaniclastic facies felsic metavolcanic strata which have locally been intruded by diabase and/or diorite synvolcanic sills and dikes. Recent geological studies at the Fivemile Lake Prospect have included detailed field mapping, diamond drill core relogging, petrographic, lithogeochemical, x-ray diffraction, and electron microprobe investigations (Hudak and Morton, 1999; Peterson, 2001; Newkirk et al., 200a, 2001b; Odette et al., 2001a, 2001b; Hudak et al., 2002a; Hocker et al. 2003). These studies were undertaken to a) better understand the geology (in particular, the physical volcanology) of these regions; b) identify and evaluate the composition and distribution of the metamorphosed hydrothermal alteration mineral assemblages that occur at each of these prospects; c) evaluate and compare the chemical compositions of alteration mineral phases to those found at VMS deposits of various ages in various parts of the Canadian Shield, the western United States, and Tasmania; and d) evaluate the potential for VMS and/or lode gold mineralization in these areas.

Field mapping and diamond drill core relogging indicate that the Fivemile Lake Prospect comprises thirteen distinct types of strata that can be subdivided into six major lithological classes (Hudak et al., 2002a). These include: a) calc-alkaline and tholeiitic, mafic to intermediate volcanic and volcaniclastic rocks (basalt-andesite pillow lavas, pillow breccias, hyaloclastite deposits, tuff, lapilli tuffs and monogenic volcanic breccias); b) chemical sedimentary rocks (semi-massive sulfide exhalite); c) calc-alkaline to tholeiitic, intermediate to felsic volcanic and volcaniclastic rocks (massive quartz-phyric rhyodacite lava flows, and aphyric to quartz-phyric rhyodacite to rhyolite tuff, lapilli tuff, and tuff breccia deposits); d) synvolcanic intrusive rocks (diorite sill-dike complexes, diabase dikes and sills, and massive gabbro); e) post-volcanic intrusive rocks (feldspar porphyry dikes and sills and quartz-feldspar porphyry dikes); and f) post-volcanic schists (quartz-sericite-ankerite schists, chlorite schists, and quartz-sericite-

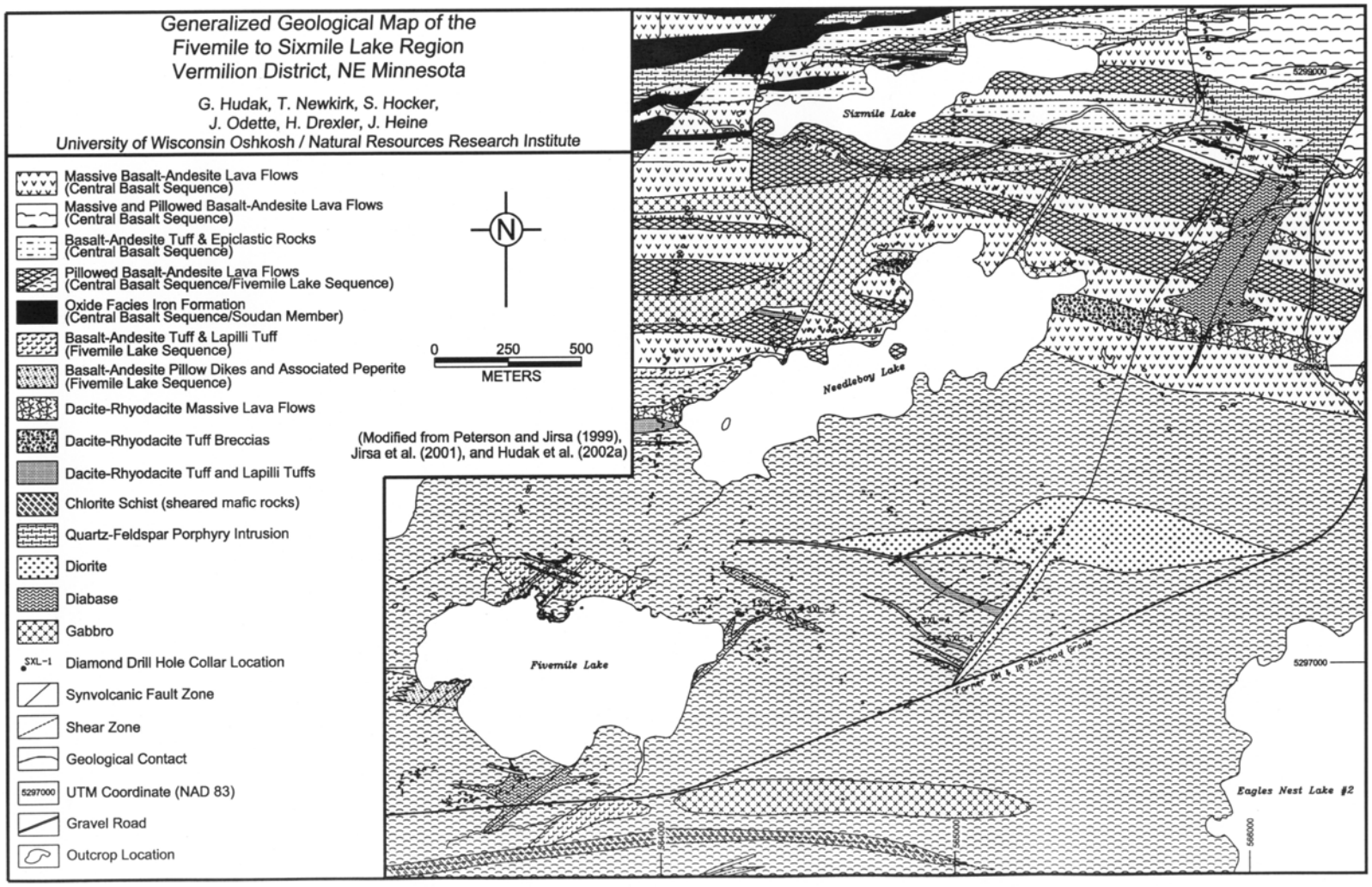


Figure 1.9. Generalized geological map of the Fivemile Lake – Needleboy Lake – Sixmile Lake area (modified after Peterson and Jirsa (1999a), Jirsa et al. (2001), and Hudak et al. (2002a).

green mica schists). Supracrustal strata at the prospect vary from east-west to southeast-northwest striking, dip steeply to the north, and are north-facing. Because of this geometry, the geological map of the area (Figure 1-9) essentially illustrates a cross-section through the Lower Ely crust. Volcanic and volcaniclastic rocks have chemical characteristics indicative of formation with an ancient volcanic arc setting (Hudak and Morton, 1999; Hudak et al., 2002a).

At this stop, we will investigate two outcrops that display exceptionally well-preserved pillow lavas and associated interpillow hyaloclastite. Detailed analysis (Newkirk et al., 2001a, 2001b; Hudak et al., 2002a) indicates that the pillow lavas at Fivemile Lake are morphologically similarly to undeformed Cenozoic pillow lavas from New Zealand and Tasmania. The nomenclature used for describing pillow lavas is illustrated in Figure 1-10).

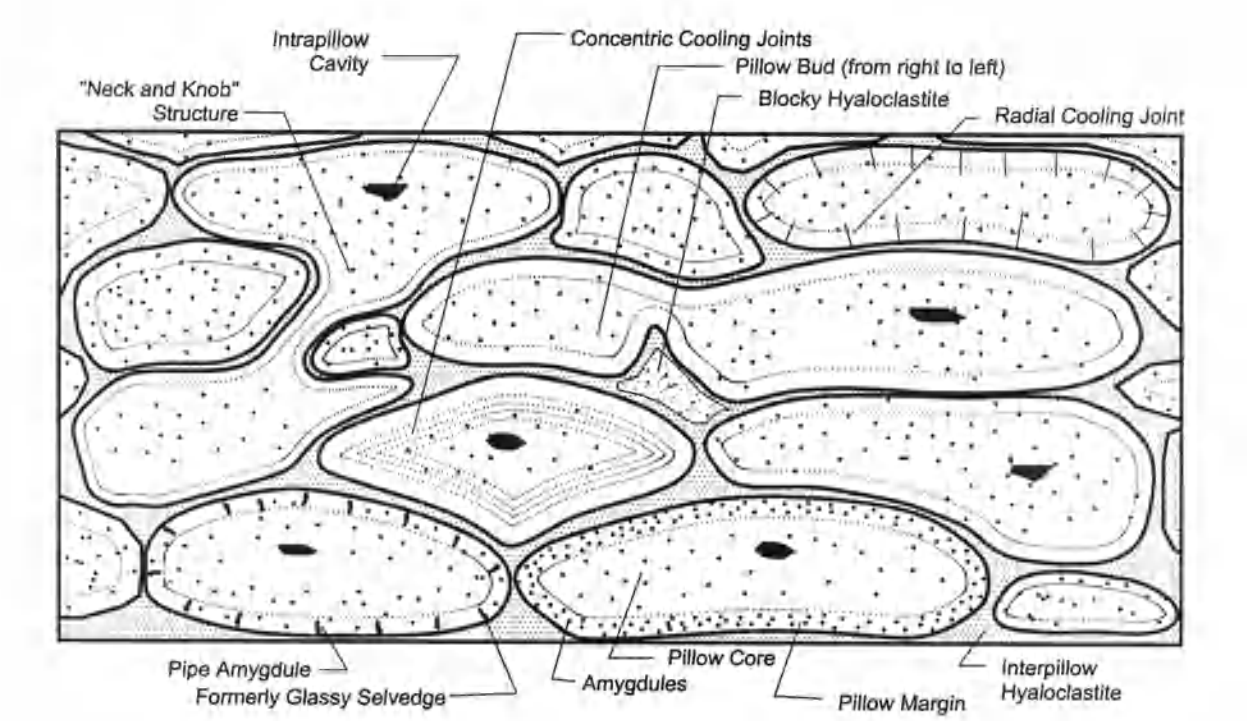


Figure 1-10. Nomenclature used for describing pillow lavas and associated hyaloclastite (after Cas and Wright (1987) and Gibson (1989).

At the first outcrop (NAD 83 UTM coordinates 563195E, 5296660N) we can observe bun- to lens-shaped, quartz-epidote-actinolite-chlorite-altered basalt-andesite pillow lavas. The general strike of the pillow lavas at this location is approximately 085°. The pillows vary from 0.4-1.0 meters high and 0.5-2.0 meters long, and contain zones of chlorite-rich interpillow hyaloclastite that ranges from 0.5-3.0 cm wide. Local brownish-red sulfide staining may be observed in the interpillow hyaloclastite zones. The pillow selvedges vary from 1-6 cm wide, are commonly silicified and/or quartz-epidote altered, and contain approximately 15% 1-5mm oval to round, quartz ± carbonate filled amygdules. Multiple pillow rinds, separated by 4-8 cm, may be locally observed. Reentrant pillow crusts and apparent pillow buds at this location suggest flow from west to east. The cores of the pillows at this location contain up to 5% 2-5 cm rounded quartz amygdules, and are faintly plagioclase feldspar-phyric (3-5% <1mm tabular phenocrysts). Faint radial cooling joints are also locally preserved in several of the pillows on this outcrop.

A second spectacular outcrop of well-preserved, east-west striking pillow lavas occurs approximately 30 meters to the east (NAD 83 UTM coordinates 563225E, 5296660N). Pillow morphology at this outcrop is quite variable, ranging from bun- and mattress-shaped to amoeboid. Topping directions based on pillow morphology are consistently north. Pillows here are typically 50-80 cm in height and 1-1.5 meters long, although pillows as small as 30 cm in diameter and up to 2 meters long may be locally observed. Pillows selvedges vary from 1-7 cm wide (average 2-3 cm) and contain 15-20%, 3-7 mm diameter, round quartz amygdules. Pillow cores generally contain 5-10% 3-10mm round to oval quartz amygdules. Several pillows on the western and north-central part of the outcrop contain multiple pillow rinds. A pillow displaying exceptionally well-preserved "neck and knob" structure may be observed on the east central part of the outcrop.

Vesicle size and abundance in pillow lavas (McPhie et al., 1993 and references therein) may be used with other textural and facies characteristics to interpret pillow lavas erupted in "relatively shallow" (<1-2 km water depth) versus "relatively deep" (>2 km water depth) submarine environments. Multiple pillow rind structures are common features in pillow lavas erupted in "relatively shallow" water environments, and are formed by implosion and rupture of the pillow skin resulting from condensation of exsolved gases concentrated in the upper part of the pillow (Kawachi and Pringle, 1988). The relatively high vesicularity of the pillows, coupled with the presence of multiple rind structures, indicate that the pillow lavas in this part of the Lower Ely were formed in a relatively shallow submarine environment.

STOP 1-5: <u>PILLOW DIKES, PEPERITES, AND SYNVOLCANIC FAULTS</u>, SW SHORE, FIVEMILE LAKE ***No hammering please!***

Location: (SW, NW, S29, T. 62N, R. 14W; NAD 83 UTM Coordinates 0563305E, 5296955N)

Description: Peperite is defined as a breccia-like volcaniclastic deposit formed from the in-situ disintegration and mixing of magma or lava with wet, poorly consolidated sediment (Batiza and White, 2000; Schmidt and Schminke, 2000). Peperite is a common rock type in submarine arc-related volcano sedimentary sequences, and is associated with synvolcanic intrusions. It often occurs in phreatomagmatic vent filling deposits and along contacts between synvolcanic intrusions and poorly consolidated, wet, submarine sediments. Key components that aid in the identification of peperites include evidence of partial fluidization of the sediment, local contact metamorphism, and in some rare instances, vesiculation within the sediment (Skilling et al., 2002).

In general, the end-member compositions in a peperite are the igneous component (a synvolcanic dike or sill, or a lava flow) and the sedimentary component. Igneous component domains in peperites vary from tabular to lobe-like to pillow-like, and can commonly have irregular, interconnected, fold like, and pod-like shapes. Mixing that results from quenching, hydromagmatic explosions, magma-sediment density contrasts, and/or mechanical stresses resulting from inflation or movement of magma, causes the igneous component of the peperite to be broken-up and dispersed within the wet sediment to form peperite (McPhie and Orth, 1999). Factors such as magma and/or sediment composition, magma injection velocities, volatile content of the magma, relative volumes of sediment and magma involved, volume of pore water heated, presence or absence of shock waves, and the confining pressure all influence the morphology of peperites. Juvenile clast morphologies in peperite are extremely variable, and may be irregular, blocky (commonly jigsaw-fit), pumiceous or scoriaceous, irregular (amoeboid), elongate, wispy, platy or ragged. It is not uncommon for mixed juvenile clast populations to be present within a single peperite deposit (Skilling et al., 2002).

This stop encompasses a series of outcrops that occur on a small peninsula along the southwest shoreline of Fivemile Lake (informally and affectionately known as "Pike Point" based on the exceptionally large pike that can be caught from this shoreline). A generalized geological map of this

area, which points out key features to observe at this stop, is illustrated in Figure 1-11. The area was originally mapped as a series of pillow lavas and felsic debris flow deposits (Peterson and Jirsa, 1999a; Peterson, 2001); however, if the pillow lavas represent a stratigraphic unit, they are oriented approximately 40°-45° to the strike of the supracrustal strata in this area, suggesting that either a) deposition occurred upon a steep paleotopographic surface (which is not supported by the pillow morphology); or b) a fold may be locally present (evidence for this is not currently supported by regional mapping).

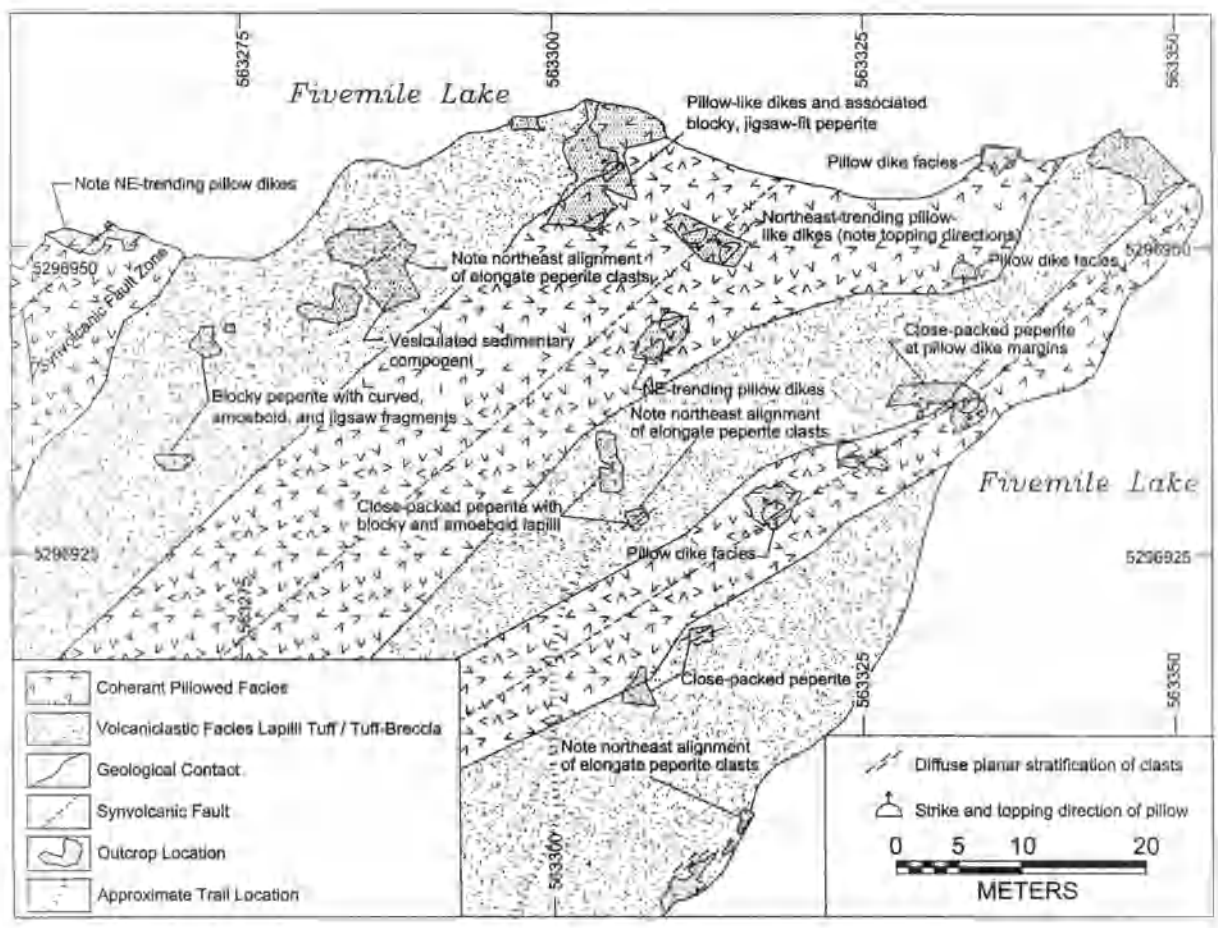


Figure 1-11. Geological map of the southwestern shoreline ("Pike Point") area of Fivemile Lake (after Hudak et al., in prep. b).

Detailed facies mapping at scales ranging from 1:50 to 1:100 by Hudak et al. (2002a) and Hudak et al. (in press) now suggests that "Pike Point" comprises a series of northeast-trending andesite-dacite pillowed dikes that intruded a deposit of wet, unconsolidated basalt-andesite tuff to form blocky to amoeboid peperites. Pillows may be formed within water-saturated hyaloclastite or sediments within the sub-seafloor provided that the adjacent rocks are poorly consolidated and saturated with water (McPhie et al., 1993 and references therein). The interpretation that the pillowed igneous domains represent dikes rather than pillowed lava flows is based on several characteristics of the dikes: 1) the strike of the pillows is northeast, whereas the strike of the volcanic strata in this region is general more or less east-west; 2) topping directions measured in the pillows are variable and inconsistent; 3) on the most northwestern of the outcrops, apophyses of the dikes can clearly be seen propagating to the north-northwest; 4) the

chemical compositions of the pillow domains and the blocky to amoeboid fragments in the volcaniclastic deposits is identical; and 5) the trend of the pillows, when followed to the north shore of Fivemile Lake, leads to a series of chemically identical, northeast-trending dikes and associated peperite deposits that occur within the vent facies of a 700 meter (strike length) by 200 meter (height) tuff cone volcano mapped by Hudak et al. (in prep. a, in prep. b; we will see these deposits tomorrow at field trip Stop 1-9). The interpretation that the volcaniclastic deposits represent peperites is based on the following characteristics of the deposits: 1) at several locations along the point, blocky jigsaw puzzle-fit fragments indicate in-situ fragmentation of northeast-trending pillow dikes; 2) fragment shapes in the volcaniclastic deposits vary from amoeboid to blocky and jigsaw puzzle-fit, to curviplanar, all indicative of fragmentation by external water; 3) vesiculation is common within the matrix of the volcaniclastic deposits; and 4) the general trend of the long directions of the fragments is consistent with the strikes of the pillow dikes. Several textures that may be observed at this location are illustrated in Figure 1-12.

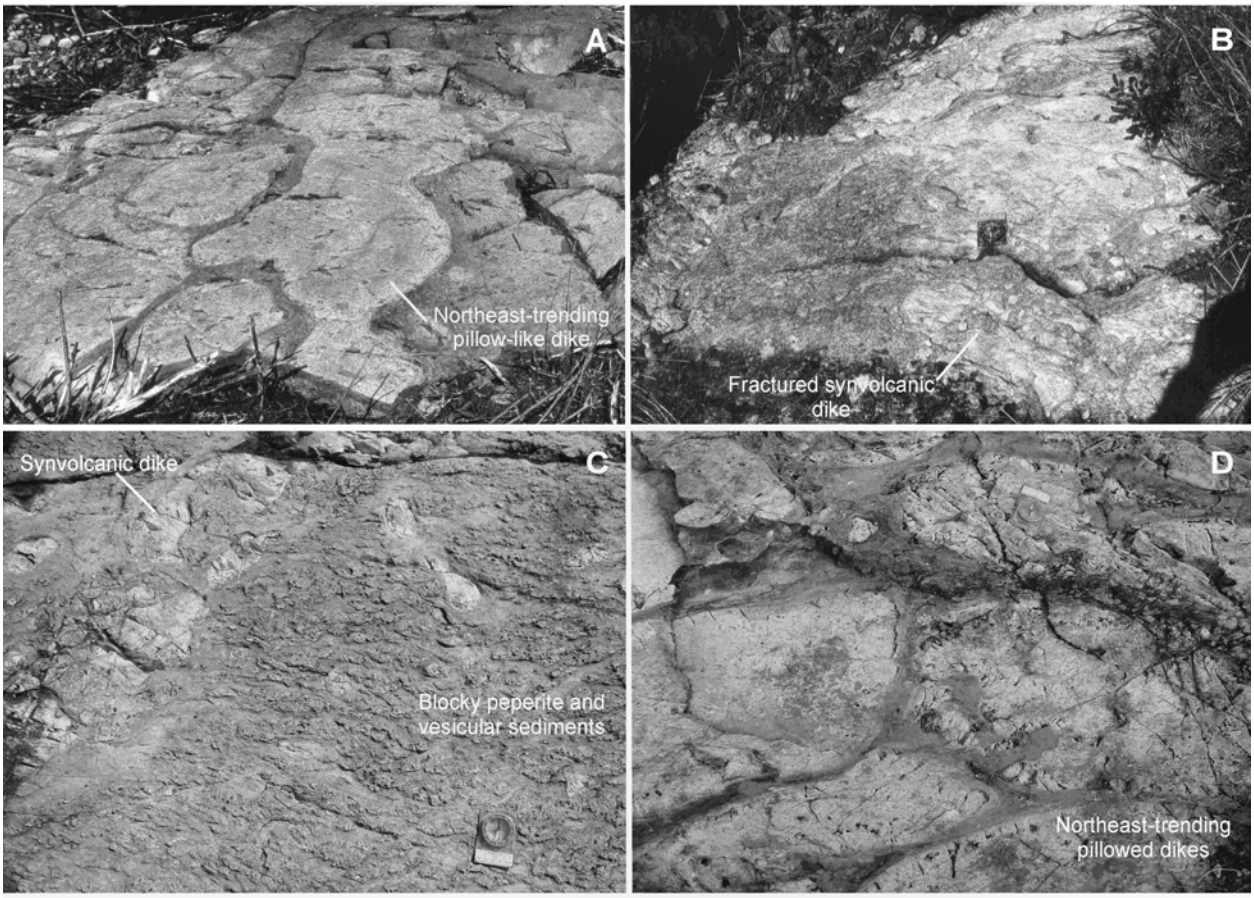


Figure 1-12. Textures within peperites that occur along the southwestern shoreline of Fivemile Lake: 1-12a) northeast-trending pillow dikes on the northwest side of the point; 1-12b) fractured synvolcanic dike, west-central part of point; 1-12c) synvolcanic dike and blocky peperite with vesiculated sediments, south-central part of point; 1-12d) northeast-trending pillowed dikes, east-central part of point.

STOP 1-6: EPIDOTE – QUARTZ ALTERED DIABASE SILLS AND DIKES, SOUTH OF FIVEMILE LAKE

Location: (NE, SW, S.29, T.62N., R.14W., NAD 83 UTM 563515E, 5296610N)

Description: This stop affords us the opportunity to investigate both synvolcanic and post-volcanic hydrothermal alteration in the Fivemile Lake area. Field, petrographic, x-ray diffraction, electron microprobe, and lithogeochemical studies have identified twelve distinctive metamorphosed hydrothermal alteration mineral assemblages at the Fivemile Lake prospect (Odette et al., 2001a, 2001b; Hudak et al., 2002a; Hocker et al., 2003). Synvolcanic, semi-conformable alteration assemblages have been identified on the basis of the modal percentages of the minerals epidote, actinolite, and chlorite, and include: a) epidote + quartz ± chlorite ± albite; b) epidote + quartz + chlorite ± albite; c) epidote + quartz + actinolite ± chlorite ± albite; and d) albite + quartz ± epidote ± actinolite. These quartz + epidote-rich alteration assemblages are interpreted to represent "seafloor metasomatism" associated with hydrothermal fluids that moved through the subaqueous rocks. Detailed mapping and lithogeochemical trends suggest that early, relatively low temperature hydrothermal alteration mineral assemblages (e.g., albite + quartz ± epidote ± actinolite) were later overprinted by higher temperature alteration mineral assemblages (e.g., epidote + quartz ± chlorite ± albite, epidote + quartz + chlorite ± albite, epidote + quartz + actinolite ± chlorite ± albite) as the volcanic pile was progressively buried by volcanism deeper below the seafloor. It is believed that the early, near seafloor environment, which was locally base metal enriched by VMS-style mineralization, eventually became depleted in base metals from later, deep-seated hydrothermal circulation. Synvolcanic, disconformable alteration mineral assemblages are commonly proximal to synvolcanic fault zones and VMS-style mineralization at the prospect, and include: a) actinolite + quartz ± iron-rich chlorite ± epidote ± albite; b) iron-rich chlorite + sericite; c) iron-rich chlorite ± sericite ± iron-carbonate ± actinolite ± epidote; d) sericite ± iron-rich chlorite; and e) mottled epidote + quartz ± actinolite. These alteration zones likely represent areas of high temperature metasomatism associated with VMS mineralization at, or up-section from, the Fivemile Lake prospect. Post-volcanic hydrothermal alteration mineral assemblages at the prospect include: a) carbonate (ankerite, dolomite, or calcite); b) sericite + carbonate (ankerite, dolomite, or calcite); and c) sericite + green mica ± carbonate (ankerite, dolomite or calcite). The close association of structurally deformed rocks and these alteration zones, and the orientations of these alteration zones, suggest that these alteration mineral assemblages resulted from syntectonic hydrothermal processes associated with the D_2 Murray Shear Zone deformational event. A hydrothermal alteration mineral assemblage map for the Fivemile Lake prospect is illustrated in Figure 1-13.

The mottled epidote + quartz + actinolite alteration zone present in this outcrop is confined to a diabase dike-sill complex located immediately south of Fivemile Lake. This 400 meter by 500 meter disconformable to semiconformable alteration zone locally cross-cuts semiconformable alteration zones comprising the epidote + quartz ± chlorite ± secondary feldspar assemblage, the epidote + quartz + iron-chlorite ± secondary feldspar assemblage, and the secondary feldspar + quartz ± epidote ± actinolite alteration assemblage.

The mottled epidote + quartz + actinolite alteration zone is characterized by a dark green, actinolite and epidote-rich groundmass that contains discrete 0.1-2.0 meter diameter, round to lens-shaped masses containing anhedral granular epidote and zoisite intergrown with polygonal quartz. These rounded epidote-quartz-rich masses can easily be mistaken for individual pillow lavas; however, they can be distinguished from pillows because they lack abundant amygdules, and have no discrete pillow selvedges or interpillow hyaloclastite zones. Petrographic, x-ray, and electron microprobe analyses indicate the presence of iron-rich chlorite, iron-rich actinolite, pistacite and zoisite/clinozoisite within this assemblage (Odette et al., 2001a, 2001b; Hocker et al., 2003).

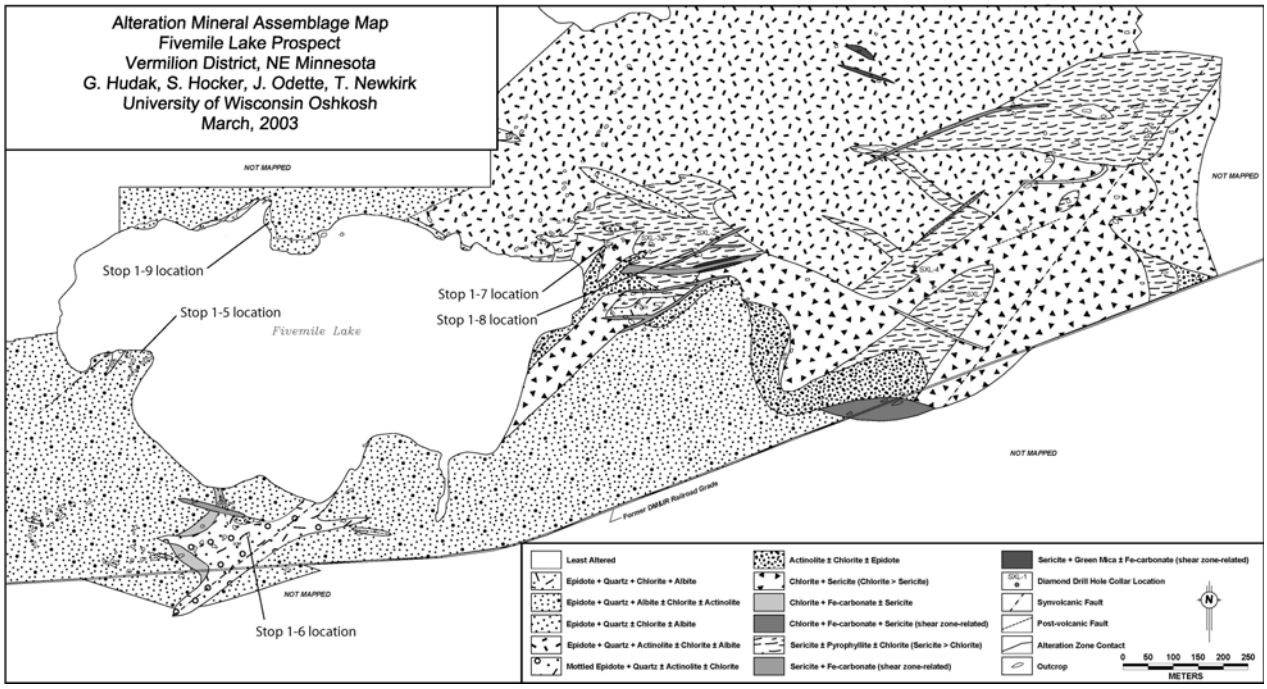


Figure 1-13. Hydrothermal alteration map of the Fivemile Lake prospect (after Hudak et al., 2002a).

The mineralogy, mineral compositions, and pronounced calcium enrichment and copper depletion that characterize the mottled epidote + quartz + actinolite alteration assemblage (Odette et al, 2001a, 2001b) are consistent with formation within a synvolcanic, semiconformable to disconformable alteration zone formed by high temperature, high fluid to rock ratio interactions deep within a sub-seafloor hydrothermal cell (c. f. Franklin, 1996; Seyfried et al., 1999). Similar clinozoisite/zoisite bearing alteration mineral assemblages described by Hannington et al. (2002) that occur down-section from the world-class VMS deposits in the Noranda mining camp in Quebec are believed to have formed from anomalous fluid flow at high water to rock ratios and high temperatures. Large zones of base metal-depleted epidosite in the Troodos ophiolite complex have been interpreted by Richardson et al. (1987) to represent deep-seated fossil reaction zones where metaliferous "black smoker" fluids are produced. The confinement of this alteration assemblage to the base of the northeast-trending diabase dike / sill complex, indicates that this alteration assemblage occurs in the deep subsurface of a northeast-trending synvolcanic structure that focused magma, as well as high temperature, perhaps base metal-rich hydrothermal fluids, up-section. Base metal exhalite mineralization occurs approximately 1 km northeast of this location (Stops 1-7 and 1-8, Fig. 1-13).

An east-southeast – west-northwest-trending post-volcanic shear zone is present in the north-central part of the outcrop. The orientation of the schistosity adjacent to the shear zone indicates a dextral sense of shear. The shear zone is one of several east-southeast – west-northwest trending shears that occur at the Fivemile Lake Prospect. Similarly oriented dextral shear zones mapped by Peterson and Patelke (2003) between the north edge of the Murray Shear Zone and the Mine Shear zone in Sections 25 and 26, T. 62N, R. 15 W (approximately 2-3 km west of this location) have been interpreted as Reidel R structures.

STOP 1-7: <u>ALTERED PILLOW LAVAS, BASE METAL EXHALITE, SCORIA LAPILLI TUFF</u>, NE OF FIVEMILE LAKE

Location: (SE, NE, S.29, T.62N., R.14W., NAD83 UTM 564260E, 5297175N)

Description: A 40-meter long, east-northeast striking, <1-2 meter thick chert-pyrite-chlorite ± chalcopyrite exhalite horizon crops out at this location approximately 50 meters northeast of Fivemile Lake (Fig. 1-13). Although this unit is volumetrically insignificant within the Fivemile Lake area, its presence is extremely important in terms of understanding potentially ore forming processes in this part of the Lower Ely. A lithogeochemical analysis of the exhalite (Hudak et al., 2002a) indicates slightly elevated Cu (470-509 ppm) and Zn (66-135 ppm). D. M. Peterson (personal communication, 1996) first recognized this unit as being similar in morphology and mineralogy to chemical exhalites associated with, and locally hosting, volcanic-associated massive sulfide mineralization in the Archean basalts and andesites within the Noranda mining camp of Quebec, Canada. This exhalite horizon appears to be on strike with the northernmost of four untested EM conductors that have been identified within Fivemile Lake (Peterson, 2001; Peterson and Jirsa, 1999a).

Extremely detailed mapping (1:120 scale) has been completed (Hudak et al., 2002a) to further evaluate the stratigraphic and structural controls on this exhalite deposit (Fig. 1-14). The exhalite is relatively easy to distinguish from its surrounding rocks based on its deep red-brown, gossan-like, sulfide-stained weathered surface. When broken, the exhalite is composed of vaguely banded chert and chlorite that contains 5-10% disseminated to banded pyrite. Locally, anhedral disseminated chalcopyrite and locally, extremely fine-grained, finely-disseminated, honey- to dark brown sphalerite grains are present in trace amounts.

Detailed mapping indicates that the exhalite unit separates epidote + quartz + actinolite ± chlorite-altered basalt-andesite pillow lavas from overlying iron-chlorite- and sericite-altered basalt-andesite tuffs and lapilli tuffs. The exhalite unit stratigraphically and conformably overlies the pillow lavas, and its sharp basal contact illustrates how the chemical sediment draped over the pillow-dominated paleotopography of the seafloor at the time of deposition. Locally, exhalite deposits cut through the pillow lavas, replacing the formerly glassy interpillow hyaloclastite. This relationship suggests that the interpillow hyaloclastite zones represented areas of high permeability, and these regions were the locus of hydrothermal fluid migration upward to, and possibly within the immediate subsurface, of the paleoseafloor. These features are consistent with the interpretation that this represents a region of vent-proximal hydrothermal activity.

The contact between the exhalite and the overlying lapilli tuff and tuff deposits is also sharp. It is important to note that fragments of the exhalite have not been observed in the basal parts of these overlying volcaniclastic deposits. In addition, no soft sediment deformation has been observed that could be attributed to syndepositional tractional deformation associated with the sedimentation of the overlying volcaniclastic rocks. This suggests that the exhalite deposit was relatively competent at the time the overlying tuffs were deposited. Such field evidence, combined with the lack of intense quartz-epidote-actinolite ± chlorite alteration (which is so prevalent in the footwall pillow basalts) in the overlying tuffs, also suggests that the hydrothermal activity associated with the genesis of the exhalite deposit was completed by the time that the overlying volcaniclastic rocks were deposited.

Three other features suggest that exhalite deposition was closely associated with a volcanic vent-proximal setting. First, property-scale mapping at Fivemile Lake (Hudak et al., 2002a) clearly illustrates that a synvolcanic diabase intrusion (observed in the previous stop) cross-cuts the stratigraphy in the immediate vicinity of the exhalite deposit (Fig. 1-9). Second, detailed mapping clearly indicates that the exhalite unit is abruptly terminated at its western margin by north to northwest striking pillow dikes that are similar in morphology to the pillow dikes that occur in close association with the peperite deposits

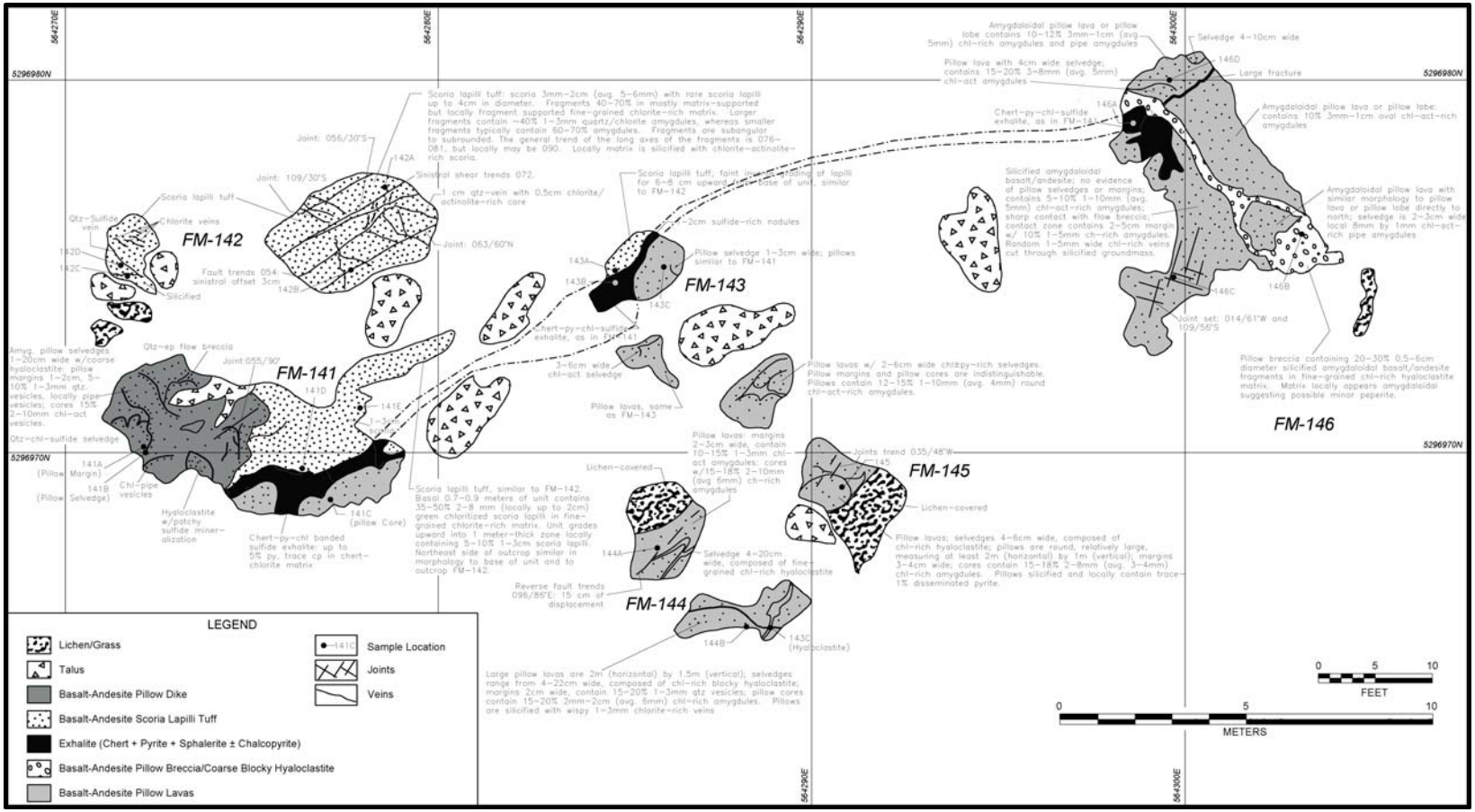


Figure 1-14. Detailed geological map of exhalite mineralization northeast of Fivemile Lake (from Hudak et al., 2002a). Note the stratigraphic positioning of the chemical exhalite horizon between basalt-andesite pillow lava and basalt-andesite lapilli tuff, suggesting a period of volcanic quiescence which was accompanied by hydrothermal activity prior to the deposition of the lapilli tuff unit. Also note that the western part of the exhalite is cut by pillowed basalt-andesite dikes, once again suggesting the presence of a synvolcanic structure. Note that all UTM coordinates on the diagram are in NAD 27 coordinates.

along the southwestern shoreline of Fivemile Lake. The location of both the diabase dikes and the pillow dikes clearly indicates the local presence of a major synvolcanic structure. Thick interpillow hyaloclastite deposits in the vicinity of the exhalite also suggest a vent-proximal environment (Newkirk et al., 2001a, 2001b; Hudak et al., 2002a).

In summary, semi-massive sulfide exhalite associated mineralization clearly has formed within a volcanic (and associated hydrothermal) vent-proximal environment at Fivemile Lake. Such environments are commonly associated with economic massive sulfide mineralization in other lava-flow dominated Archean volcanic settings, such as at Noranda, Quebec (Gibson, 1989; Gibson et al., 1999; Santaguida, 1999).

STOP 1-8: <u>RHYOLITE LAVA DOME</u>, NE OF FIVEMILE LAKE

Location: (SE, NE, S.29, T.62N., R.14W., NAD 83 UTM 564440E, 5297160N)

Description: The first day of our field trip concludes with an investigation of a 200-meter long, 15-45 meter thick unit of quartz-phyric, rhyodacitic to rhyolite lava that occurs approximately 200 meters northeast of Fivemile Lake (Figs. 1-9 and 1-13). The U/Pb zircon age for this unit is 2722.6 ± 0.9 Ma (Peterson et al., 2001). This biscuit shaped, relatively thick mass of rhyodacitic to rhyolitic is morphologically similar to felsic lava domes and cryptodomes (Cas and Wright, 1987). The presence of quartz-phyric lava fragments in epiclastic deposits approximately 400 meters southeast of this location, as well as the presence of the overlying subaqueously deposited volcaniclastic sediments and intermediate to mafic lava flows, suggests that this unit was at least partially emergent as a subaqueous lava dome.

The uppermost contact of this unit is not exposed in outcrop. Where exposed in drill core (diamond drill hole SXL-2), the massive quartz-phyric lava is interstratified with, and overlain by, 1-2 meters of felsic tuff, which in turn, is overlain by pillowed basalt-andesite lava flows. The basal contact of the unit is exposed in diamond drill holes SXL-2 and SXL-3. In SXL-2, the basal contact is represented by a 3-meter thick shear zone composed of quartz, carbonate, sericite and green mica (fuchsite or mariposite?). In SXL-3, the base of the unit is immediately underlain by a 2-meter thick quartz-feldspar porphyry dike that, in turn, is underlain by a shear zone composed of quartz-carbonate-sericite-chlorite schist. This shear zone locally contains 1-3%, 1-8mm wide veins and bands composed of red to red-brown sphalerite. The presence of this shear zone conservatively precludes using the pillowed basalt-andesite lava flows as reliable bounding facies; however, the lack of evidence for extensive structural transposition of the felsic lavas, the extremely thick sequence of pillow lavas stratigraphically below the felsic lavas, and the presence to the southeast of submarine epiclastic strata with felsic lava fragments identical in composition to the dome, suggest that the felsic lavas were formed in a subaqueous environment.

Drill core intersections of these lava flows are characterized by a massive to faintly amygdaloidal (up to 1% 5mm rounded gray quartz amygdules), pale gray (least altered) to pale yellow green (sericitized) aphanitic groundmass which contains 1-4% grayish blue to colorless euhedral square to subhedral, locally broken, <1-2.5mm quartz phenocrysts (Fig. 1-15a). Pale brown to honey-colored sphalerite veins are locally present in the drill core and range from <1-8cm in width (Figure 1-15b). Two types of veins are present: a) sphalerite veins which are associated with quartz and/or chlorite, which display no indication of structural deformation (in-situ sphalerite veins); and b) sphalerite veins which are associated with local domains of quartz-sericite-ankerite schist, and appear to have been formed by local remobilization of sphalerite along post-volcanic shear zones (structurally remobilized sphalerite veins). The presence of in-situ sphalerite veins, and the abundance of structurally remobilized sphalerite veins in the quartz-phyric lava flows relative to other units in the Fivemile Lake area, suggests that the original

sphalerite mineralization in the Fivemile Lake area was synvolcanic and associated with the massive quartz-phyric lava flows.

As seen in this exposure, the quartz-phyric lava flows contain an aphanitic, pale-gray to gray groundmass. These lavas contain 5-8% 1-3mm diameter subhedral to square euhedral quartz phenocrysts that are locally broken. Rare, 2-4mm subhedral to euhedral quartz glomerocrysts are locally present. Faint, pale gray <1-2mm tabular feldspar phenocrysts are also present, and comprise up to 5% of the rock. Petrographic analysis indicates that lavas are characterized by a very fine-grained quartzo-feldspathic groundmass and subhedral to euhedral quartz phenocrysts that are locally resorbed. Alteration minerals that occur in the groundmass include sericite, epidote, clinozoisite/zoisite, iron-carbonate and iron-rich chlorite. Also present on the southwest side of this outcrop is a northeast-trending, post-volcanic quartz-feldspar porphyry dike.

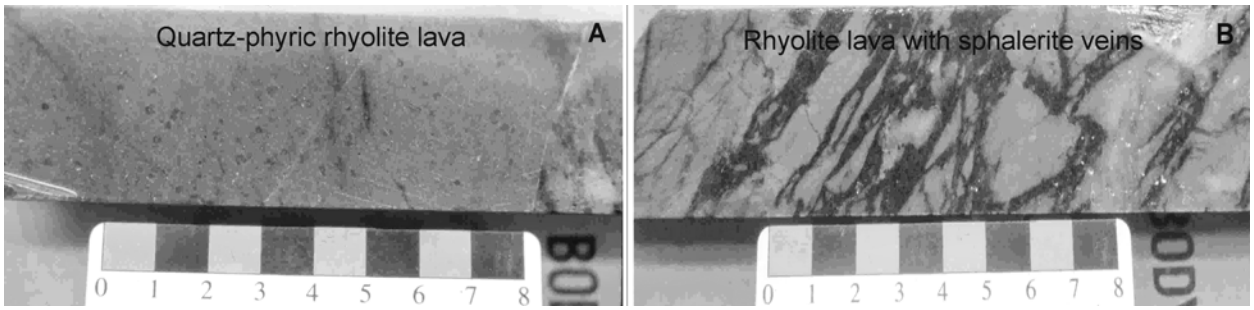


Figure 1-15. Drill core appearance of unmineralized, sericite-altered rhyolite dome (A, SXL-2-173') and silicified rhyolite dome cut by red-brown "honey-colored" sphalerite veins (B, SXL-2-280'). Modified from Hudak et al. (2002a). Scale bar units are centimeters.

FIELD TRIP STOPS – DAY 2 (MAY 5)

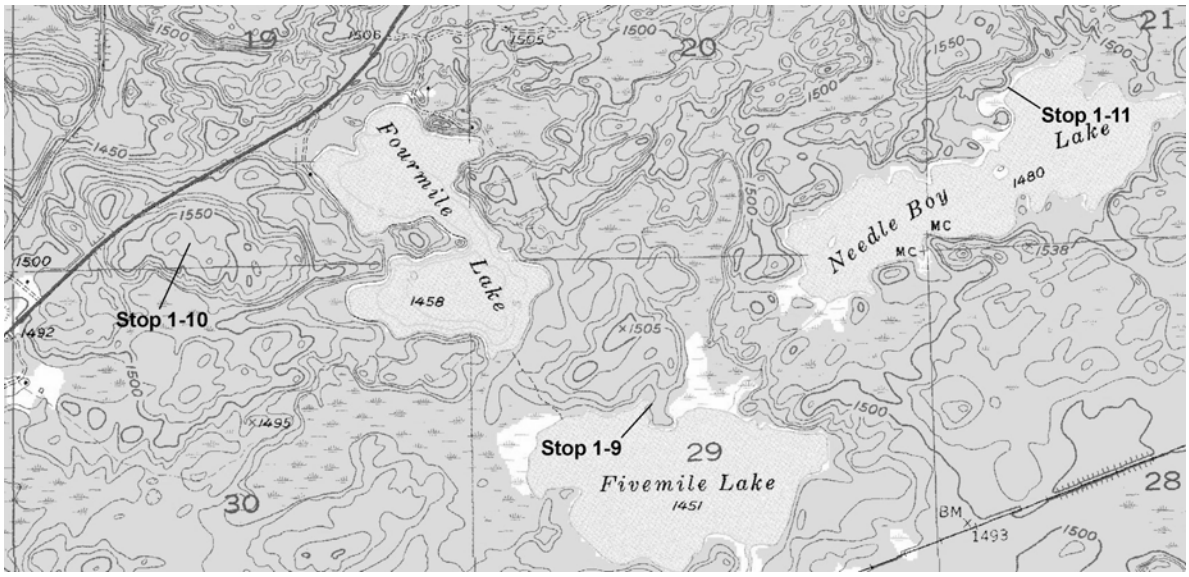


Figure 1-16. Locations of field trip stops 1-9, 1-10, and 1-11.

STOP 1-9: **<u>FLUIDAL PEPERITES AND VOLCANIC VENT FACIES</u>, NORTH SHORELINE OF FIVEMILE LAKE** ***No hammering please!***

Location: (SE, NW, S.29, T.62N., R.14W., NAD 83 UTM 563590E, 5297215N). Note: We will be accessing the north side of Fivemile Lake via a private road. Obtain permission to use this road from people in the first house on the right as you turn onto this private drive.

Description: Detailed (1:100 to 1:2000 scale) mapping by Hudak et al. (in prep. a) and Hudak (in press) has identified the paleo-vent of a relative small basalt-andesite tuff cone volcano immediately north of this location (Figs. 1-9 and 1-16). The vent facies here is well exposed in a series of outcrops along the north shoreline of Fivemile Lake, and is extremely complicated (Fig. 1-17a). At this location, we can observe steeply north dipping, east-west striking basalt-andesite pillow lavas that occur stratigraphically beneath the tuff cone, and the contact between the pillowed flows and the vent fill deposits, which are composed of exceptionally well preserved blocky to amoeboid peperites (Figs. 1-17 b and 1-17c).

The contact between the pillowed flows and the vent fill deposits is interpreted to be a synvolcanic fault zone. Several jigsaw puzzle-fit, more or less in-situ pillow lava blocks can be observed within the vent fill deposits near this contact. The basalt-andesite tuff comprising the vent fill deposits have been intruded by a series of andesite-dacite dikes that are compositionally identical to the pillowed dikes observed southwest of this location at "Pike Point". Interactions between the wet, unconsolidated vent fill tuffs and the dikes have produced an absolutely spectacular deposit of amoeboid peperite. The peperite displays several exceptional textures, including: 1) jigsaw puzzle-fit igneous component lapilli and blocks indicative of in-situ fragmentation; 2) local contact metamorphosed zones ranging from a few millimeters to a few centimeters in width immediately adjacent to the lapilli and blocks within the peperite; 3) local chilled margins on the peperite fragments indicating rapid quenching and fragmentation of the igneous component from interactions with relatively cool, wet, unconsolidated vent fill tuffs; 4) complicated amoeboid peperite, especially on the northeastern part of the outcrop; and 5) the presence of 1-2mm diameter, commonly quartz-filled vesicles within the vent fill tuffs. Also note the presence of 1-2 cm wide, several centimeter-long parallel bands containing high abundances (20-50%) of 1-2mm vesicles. These bands are generally more or less parallel with the margins of adjacent igneous component peperite fragments, but locally, these bands show no preferred directional relationship with the margins of nearby clasts.

Petrographic observations (Hudak et al., in press; Hudak et al., in prep. b) indicate that these zones are composed of parallel bands of highly vesicular, locally plagioclase-phyric fragments that are compositionally identical to the coherent igneous component clasts in the peperite. The origin of the bands may be related to collapse of thin vapor films along the margin of the hot igneous component as occurs in laboratory simulations of peperite formation (Zimanowski and Buttner, 2002). The parallel bands in this outcrop may have formed when the margins of the igneous component shattered due to explosive interaction with external water as the vapor film collapsed. The parallel bands observed on the outcrop indicate that this process occurred periodically and continually. Such a process is consistent with the pulsating phreatic and phreatomagmatic eruptions that occur in shallow submarine environments.

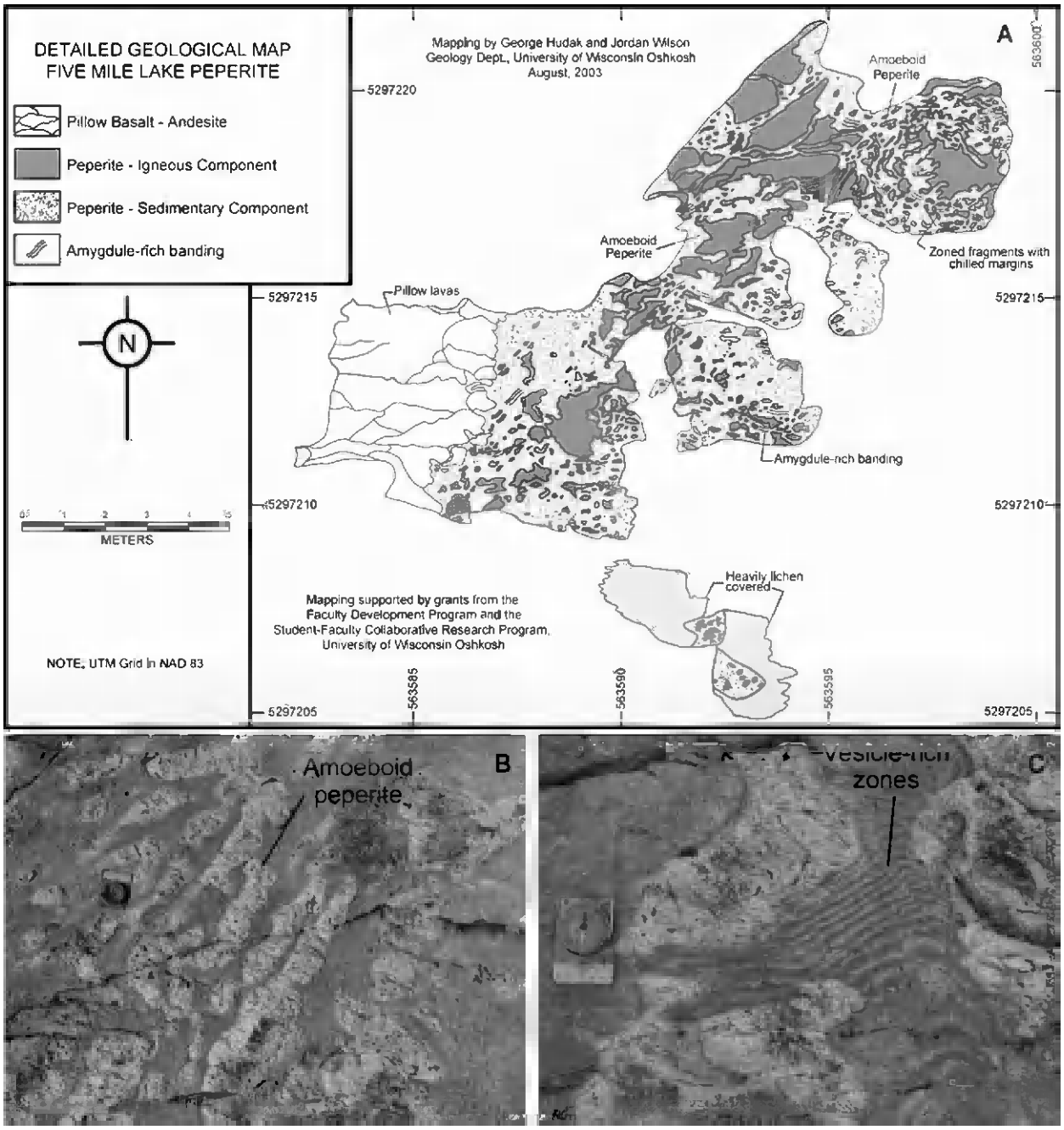


Figure 1-17. Detailed geological map (A) of the vent facies of a basalt-andesite tuff cone volcano located north of Fivemile Lake. Photographs A and B illustrate amoeboid peperite and parallel, vesicle-rich zones present in the outcrop.

STOP 1-10: CENTRAL BASALT SEQUENCE SHEET FLOWS, PILLOW LAVAS, AND PERLITIC HYALOCLASTITE, HILLTOP S OF HIGHWAY 169

Location: (SE, SW, S.19, T.62N., R.14W., NAD 83 UTM 56200E, 5297800N)

Description: 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 composition that are believed to be correlative with the tholeiitic Armstrong Lake volcanic sequence mapped by Jirsa et al. (2001) in the Eagles Nest quadrangle. Relative to massive and pillowed basalt and andesite flows in the Fivemile Lake sequence, Central Basalt sequence lavas 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.

This field trip stop (Fig. 1-16) displays exceptional preservation of fine-scale volcanic textures that are characteristic of the Central Basalt sequence. 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. 1-18). 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 (Hudak et al., in prep. b) 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 associate 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. Irregular, amoeboid apophyses of coherent basalt, ranging from 10-30 cm thick and up to 1.5 meters long, locally intrude into the hyaloclastite. These apophyses have pale green cores and dark green, very fine-grained margins which grade outward into in-situ, jigsaw-fit perlitic hyaloclastite. 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 pillows 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. A NNE-trending D_3 normal fault with a 10 cm displacement occurs in this unit on the southeastern part of the outcrop. 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 D_3 joints are well developed in this unit, as are lens-shaped psuedo-pillows that are up to 50 cm in diameter.

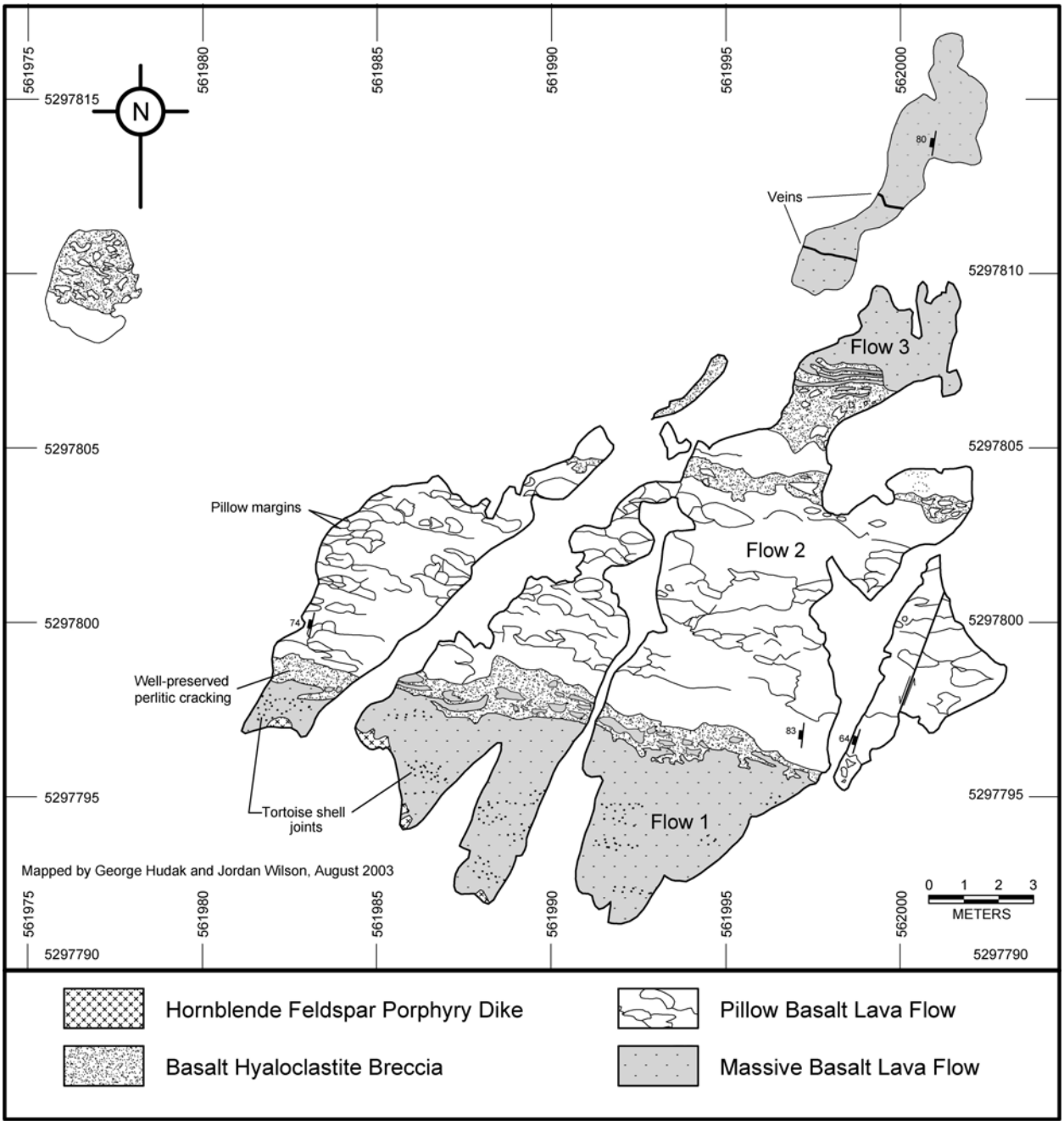


Figure 1-18. Detailed geological map of sheet flows, pillow lavas, hyaloclastite, and associated "self-peperite" at field trip stop 1-10.

STOP 1-11: <u>RHYODACITE BLOCK AND ASH FLOW DEPOSITS</u>, NORTH SHORELINE OF NEEDLEBOY LAKE

Location: (Optional Stop, NW, SW, S.21, T.62N., R.14W., NAD 83 UTM 564855E, 5298360N) This is an optional stop, as it requires a moderately strenuous 20-30 minute hike along a poorly maintained, easy to trip upon trail. *Be extremely careful and watch for protruding boulders and cobbles on the trail.*

Description: A 50 meter-thick unit of quartz-phyric rhyolite tuff and lapilli tuff crops out in a series of several closely-spaced outcrops along the northwestern shoreline of Needleboy Lake (Fig. 1-16). The pale gray to green-gray matrix contains 1-3% subhedral to angular quartz phenocrysts and crystal shards, 10-15% 1-5 cm subangular to subround pumice lapilli, and locally, up to 15% 2-11 cm subround to subangular quartz-phyric rhyolite lava lapilli and bombs which locally display a crude jigsaw puzzle-fit. Both the pumice and the felsic lava clasts are locally crudely stratified. Pumice lapilli altered by epidote commonly occur as pale green fragments within the deposits. The central part of the outcrop contains a well-developed, east-west trending, steeply north-dipping shear zone. Boudinage locally occur in the rhyolite deposits within this shear zone. A 30-35 meter thick gabbro sill occurs immediately north of this shear zone. This unit is characterized by an equigranular medium gray-green to brownish-green color, with pale green epidote altered feldspar and deep green actinolite present. Petrographic observations (Hudak et al., in prep. a) indicate that the gabbro has a sub-ophitic texture. Locally, steeply-dipping north-south-trending D_3 joints and east-west-trending quartz epidote veins may be observed on the northeastern side of this series of outcrops. A large east-west-trending outcrop occurs approximately 40 meters northwest of the northern-most outcrop near the shoreline. The southernmost 5 meters of the outcrop comprises a mafic lapilli tuff that contains up to 5% subangular silicified mafic lapilli. This tuff is progressively deformed into dark green chlorite schist with a well-developed, east-west-trending, steeply south-dipping S_2 foliation. The chlorite schist is in sharp contact with a massive, 3-12 meter-thick quartz- and feldspar-phyric rhyodacite sill. The sill is overlain by a mafic lapilli tuff containing up to 25% subangular to subround, locally jigsaw puzzle-fit amygdaloidal basalt and scoria clasts.

Felsic volcaniclastic deposits at the base of the outcrop are interpreted to represent a sequence of variably-altered rhyolite block and ash flow deposits. Block and ash flow deposits are unsorted deposits with an ash-matrix containing both monolithologic, low-vesicularity lava lapilli and blocks and pumice lapilli and blocks that may display normal or reverse coarse-tail grading (Cas and Wright, 1987; Freundt et al., 2000). The deposits are formed during the explosive collapse of lava domes, and can occur in either the subaerial or submarine environment (Hudak, 1996; Gibson et al., 1999). These deposits are believed to have been initially overlain by a series of basalt-andesite tuffs and lapilli tuffs; however, a synvolcanic gabbro sill intruded the contact between and subsequently separated these two units. Deformation that is concentrated along the contacts between this sill and the adjacent supracrustal strata is attributed to strain partitioning due to different rheological properties during the D_2 deformation event.

STOP 1-12: <u>SYNVOLCANIC FAULT ZONE WITH "EPIDOSITES"</u>, SE OF SIXMILE LAKE

Location: (NE, SE, S.21, T.62N., R.14W., NAD 83 UTM 565860E, 5298535N) *During the hike through the swampy area, be extremely careful to walk on areas with vegetation and tree roots.*

Description: The next several stops (1-12 through 1-17) represent a north-south section through the uppermost parts of the Lower Ely (Fig. 1-19). Basalt-andesite lava flows in this region vary from massive to pillowed, and are generally less amygdaloidal than the basalt-andesite lava flows in the southern part of the section near Fivemile Lake. The purpose of this stop is to investigate a northeast-trending,

disconformable zone of intensely quartz + epidote ± actinolite altered pillow lavas (epidosites), and to observe one of the post-volcanic quartz-feldspar porphyry dikes that commonly occur within the volcanic sequence.

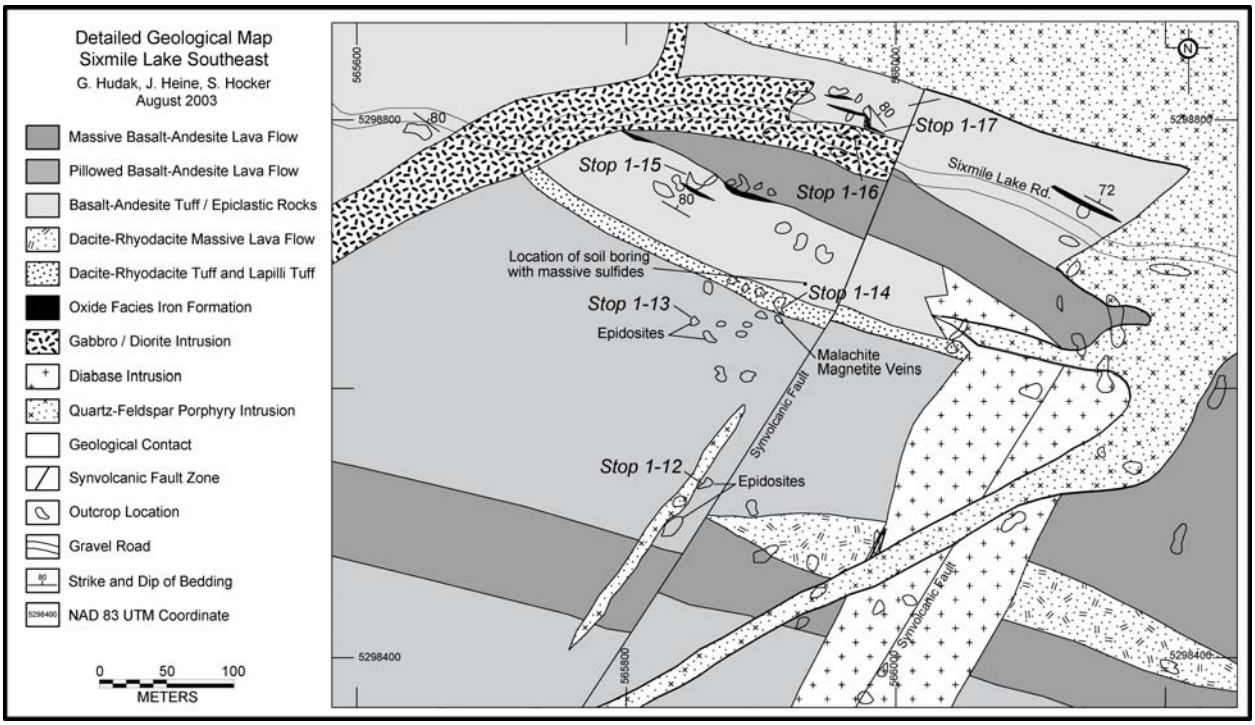


Figure 1-19. Detailed geological map of the area southeast of Sixmile Lake. Field trip stop locations 1-12 through 1-17 are also illustrated on the map.

The first outcrop we will investigate is composed of intensely epidote + quartz-altered basalt-andesite pillows and lobes. The pillows may be difficult to see due to the intense synvolcanic hydrothermal alteration, but may be located by identifying actinolite ± chlorite-altered zones of interpillow hyaloclastite, as well as concentrated zones of 0.1-1.0 cm round to oval actinolite ± chlorite-filled amygdules which occur within the pillow selvedges. Dark gray veins of magnetite with minor amounts of pyrite, chalcopyrite, and malachite up to 1 cm in width occur primarily within interpillow hyaloclastite zones, but also locally cross-cut the interiors of the pillows and lobes. The patchy pistachio-green and pink coloration on the outcrop is due to the presence of pistacite epidote (green) and zoisite-clinozoisite (pale pink). Santaguida et al. (2002a, 2002b) indicate that epidote compositions within epidote-quartz altered rocks in the Noranda Volcanic Complex of Quebec are sensitive to small-scale variations in fluid:rock ratios within individual alteration zones due to primary permeability differences in the strata being altered. Hannington et al. (2002) note that within the Proterozoic Kristineberg VMS district of Sweden, zoisite-clinozoisite formation occurs in regions altered at relatively high fluid:rock ratios and high temperatures by acidic hydrothermal fluids.

The second outcrop to observe occurs approximately 20 meters to the southwest of the first outcrop, and is located near the top of a prominent east-northeast trending ridge. Compositionally, the rock is identical to the first outcrop. Texturally, the rock is intensely brecciated, containing lapilli- to block-sized subangular fragments of intensely epidote + quartz-altered amygdaloidal basalt. Disconformable zones containing jigsaw-fit epidosite occur proximal to synvolcanic fault zones in the Josephine Ophiolite (Harper, 1999). Based on detailed mapping, the northeast trending, disconformable

zone of epidosite comprising these outcrops has been interpreted by Hudak et al. (2002b) as being proximal to a synvolcanic fault zone.

The third outcrop occurs approximately due north of the second outcrop visited. Exposed at this location is a relatively fresh quartz-feldspar porphyry dike. Similar quartz-feldspar porphyry dikes commonly occur throughout this region of the Vermilion district.

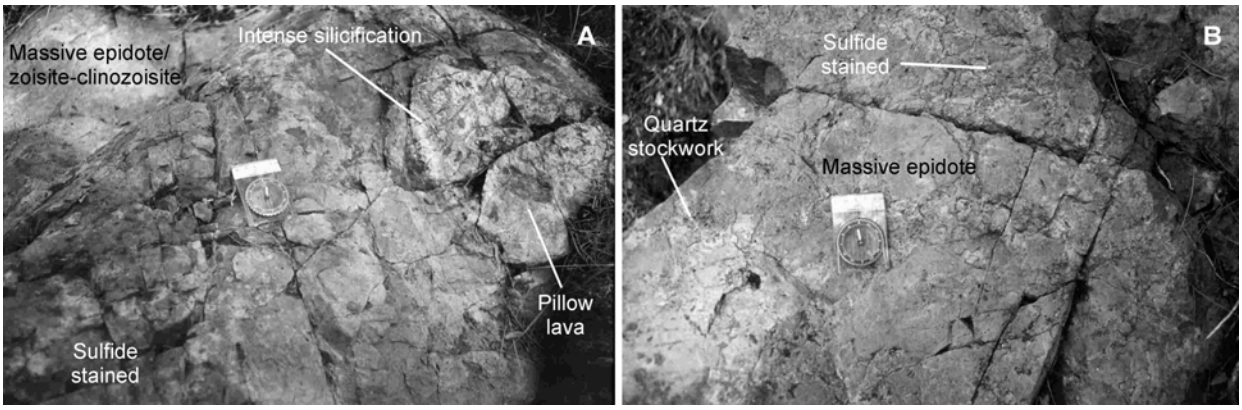


Figure 1-20. Photographs of intensely epidote-altered basalt-andesite lava flows from the area southeast of Sixmile Lake. Photo A shows brecciated, intensely epidote–quartz altered pillow lavas. Photo B shows typical relationships between intense epidote alteration, stockwork quartz veins, magnetite stringers, and sulfide staining.

STOP 1-13: <u>SYNVOLCANIC FAULT ZONE WITH EPIDOSITES</u>, SE OF SIXMILE LAKE

Location: (SE, NE, S.21, T.62N., R.14W., NAD 83 UTM 565850E, 5298560N)

Description: Several relatively small outcrops of intensely epidote + quartz ± actinolite ± chlorite-altered basalt-andesite pillow lavas occur along the southwestern part of an east-west trending ridge at this location. These rocks are mineralogically similar to those investigated at the previous outcrop; however, despite their intense alteration, they display extremely well preserved primary volcanic textures. The pillows at this location are primarily bun-shaped, and range from approximately 1-1.5 meters in diameter. Interpillow hyaloclastite zones range from <1-2 cm in width, and are pale yellow green in color due to the presence of epidote, actinolite, and chlorite. Pillow selvedges and cores once again vary in color from pale green to pinkish green due to variable percentages of pistacite and clinozoisite/zoisite. The pillow selvedges and cores contain up to 5% 0.1-1.0 cm diameter round to oval quartz ± actinolite ± chlorite-filled amygdules.

STOP 1-14: <u>STRINGER COPPER MINERALIZATION IN BASALT–ANDESITE, AND SERICITE-ALTERED RHYODACITE TUFFS</u>, SE OF SIXMILE LAKE

Location: (SE, NE, S.21, T.62N., R.14W., NAD 83 UTM 565910E, 5298655N)

Description: Another small group of outcrops occurs approximately 50 meters northeast of the previous outcrops investigated (Fig. 1-19). Stratigraphically, we are now positioned within the uppermost 100 to 200 meters of the Lower Ely. At this location, we can observe locally brecciated, strongly epidote + quartz-altered, and locally actinolite- and chlorite-altered basalt-andesite pillow lavas. Pillows at this outcrop are significantly smaller (generally approximately 0.5 meters in diameter) than those observed in the previous outcrop, suggesting that we are positioned near the top of a pillowed flow unit (Dimroth et

al. (1978) and Staudigal and Schminke (1984) have shown that pillows generally decrease in size near the top of a pillowed unit, and attribute this to change in pillow size due to decreasing effusion rates as an individual eruption comes to an end). The outcrop also locally contains up to 5% dark gray stringer-like veins of magnetite with associated pyrite, chalcopyrite, malachite, and associated deep-red sulfide burn (Fig. 1-21a). These veinlets are commonly parallel to a series of quartz stockwork veins that trend approximately 080° and dip steeply to the north. Both the magnetite-sulfide veins and quartz stockwork veins are consistent with cooling of a high temperature, iron-, base metal-, and silica-rich hydrothermal fluid in the shallow seafloor, and may represent stringer mineralization that occurs immediately down-section from massive sulfide mineralization.

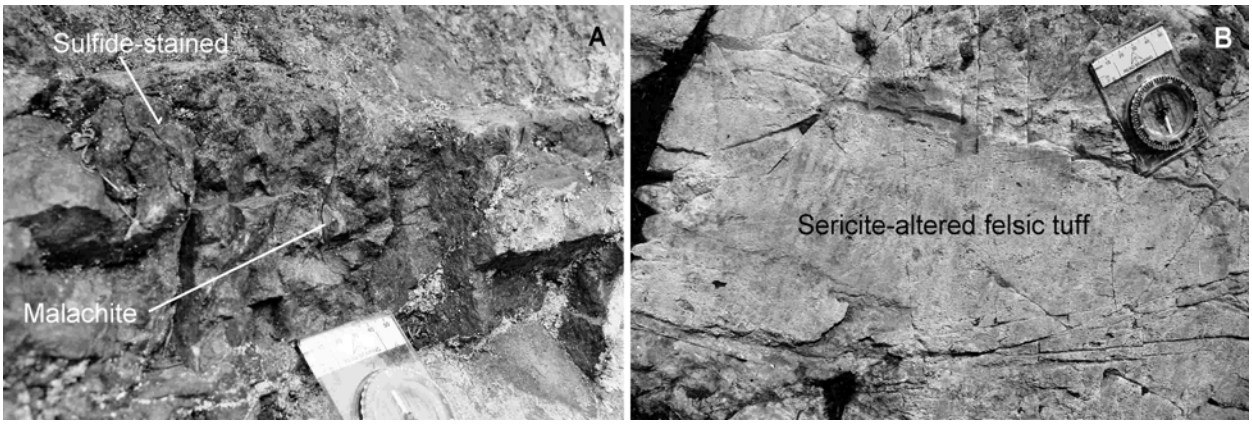


Figure 1-21. Mineralization and alteration present at field trip stop 1-14. Photograph A illustrates malachite mineralization associated with magnetite-chalcopyrite veins within altered basalt pillow lavas. B shows sericite-altered felsic tuff deposits located immediately up-section from the altered mafic lavas in photograph A.

The pillow lavas are immediately overlain by a 10-20 meter thick sequence of sparsely quartz and plagioclase-phyric rhyodacite tuffs (Fig. 21b). These volcaniclastic deposits typically are brownish-gray in color due to the presence of abundant sericite. Close-up observation of the unit reveals the presence of approximately 1% sub-1mm anhedral to subhedral quartz crystal chips, and 2-3% 1 mm diameter subhedral to euhedral tabular plagioclase feldspar phenocrysts. Rare pumice lapilli can be observed with careful observation; they range from 1-2 cm in diameter, are generally rounded to oval in shape, and contain up to 50% sub-1mm round vesicles (originally carbonate vesicles which have now been weathered-out based on petrographic observations by Hudak et al. (in prep. a)). The pumice lapilli in this outcrop increase in abundance from the base toward the top of the unit. Such inverse grading of pumice may be the result of subaerially erupted air fall deposited into the ocean, as large, cold pumice generally take longer to saturate and sink than small, cold pumice (Whitham and Sparks, 1986). This tuff horizon may represent the distal equivalent to fragmental felsic rocks mapped by Peterson and Patelke (2003) that occur stratigraphically below the Ely Greenstone – Soudan Member and the Soudan Mine. Sericite alteration is common in submarine felsic volcaniclastic rocks (Gibson et al., 1989), and may be associated with either vertically extensive discordant pipe-like alteration zones associated with hydrothermal upflow zones, or laterally extensive semiconformable alteration zones formed by extensive shallow seafloor seawater metasomatism of permeable submarine strata. More detailed mapping will be necessary to determine with certainty the geometry of the sericite alteration zone observed in this outcrop.

The exposures of felsic tuff dive abruptly to the north into a swamp. Hand auger boring in this swamp (NAD 83 UTM location 0565938E, 5298671N) yielded dark black, organic-rich soil containing abundant, angular, variably weathered clasts of massive pyrite. Stringer copper mineralization, sericite-altered felsic tuffs, the presence of massive sulfide in the subsurface beneath the swamp, and the presence

of rare sulfide lapilli in volcaniclastic deposits north of the swamp (seen in the next outcrop) suggests that the swamp may be underlain by a massive sulfide horizon (Hudak et al., 2002b).

STOP 1-15: CHERT, BANDED IRON-FORMATION AND ASSOCIATED CLASTIC SEDIMENTARY STRATA, SE OF SIXMILE LAKE

Location: (SE, NE, S.21, T.62 N., R.14W., NAD 83 UTM 565840E, 5298760N) *Be extremely careful when walking on these outcrop surfaces and descending into the small valley immediately south of these outcrops, as they are commonly very slippery.*

Description: North of the swamp, a southeast-trending outcrop ridge comprises a sequence of volcaniclastic and chemical sedimentary strata (Fig. 1-19). The purpose of this stop is to observe the variety of interflow sedimentary rocks that occur in this upper part of the Lower Ely, to observe well-preserved sedimentary textures, and to see further evidence for the presence of ancient hydrothermal activity southeast of Sixmile Lake.

A series of small outcrops occurring along the northwestern part of the outcrop ridge comprises west-northwest striking, steeply north dipping 10-30 cm thick chemical exhalites (chert horizons and banded magnetite + chert ± jasper banded iron-formation horizons) are that are interbedded with 10-60 cm thick, well-stratified horizons of mudstones, wackes, and matrix-supported polymict diamictites. The base of the stratigraphic sequence occurs at the southern end of the outcrop, where a 20 cm thick magnetite-rich horizon of banded iron-formation occurs (Fig. 1-22a). This chemical sedimentary horizon is overlain by a several meter-thick sequence of interbedded, laminated to bedded, commonly normally graded, mudstones and polymict diamictites interpreted to have formed from submarine debris flows and turbidity currents. The polymict diamictites range from 10-60 cm thick, and contain up to 5% 1-2.5 cm lens-shaped chert clasts, 10-40% silicified or chlorite-altered mafic volcanic lapilli, and rare (<1%) <1-2cm sulfide-rich fragments. Locally, clasts within the diamictites are vaguely imbricated. Interbedded mudstone horizons are up to 40 cm thick, and locally display cross beds that indicate stratigraphic topping direction is north.

Iron-formation horizons at this location indicate that low temperature exhalative hydrothermal activity was occurring in this part of the Lower Ely. It is interesting to note that these iron-formation horizons occur in the area that would be immediately up-section from the possible sulfide horizon within the swamp to the south. Several VMS horizons in similar volcanic sequences within Canada are capped by iron-formations (Franklin et al., 1981). Debris flows and/or turbidity currents that deposited the polymict diamictites and associated mudstones periodically interrupted periods of relatively low-temperature hydrothermal activity that formed the iron-formation horizons. Clasts within the diamictites are similar to the volcanic and chemical sedimentary strata that occur in the immediate vicinity of the deposits based on detailed mapping (Hudak et al., in prep. a). Chert clasts within the diamictites were likely derived from the iron-formation horizons. Rare sulfide-rich fragments locally contained within the diamictites also suggest that VMS mineralization occurred on or near the paleoseafloor in the vicinity of the area that is now Sixmile Lake, supporting the interpretation that a massive sulfide horizon may be present beneath the swamp which separates the outcrops at this stop from those at the previous stop.

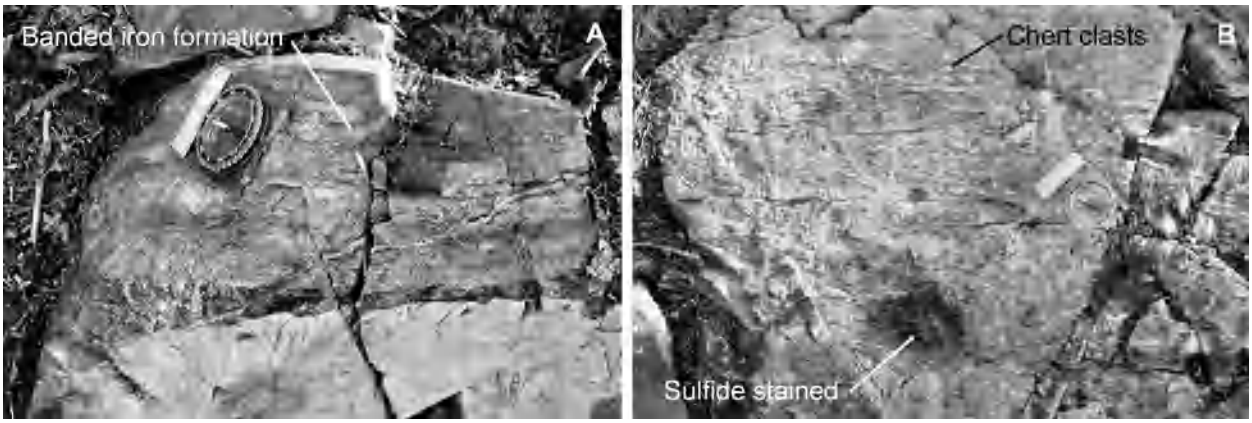


Figure 1-22. Sedimentary strata in the uppermost parts of the Lower Ely. Figure 1-22a shows oxide-facies banded iron-formation and chert exhalites located up-section from altered tuffs seen at field trip stop 1-15. Figure 1-22b shows diamictites that contain abundant chert clasts and rare sulfide clasts. These deposits are also located stratigraphically up-section from the felsic tuffs seen at the previous stop.

STOP 1-16: EPIDOTE-QUARTZ ALTERATION ON MARGIN OF DIABASE/GABBRO SILL, SE OF SIXMILE LAKE

Location: (SE, NE, S.21, T.62N., R.14., NAD 83 UTM 565960E, 5298790N) *The outcrop at this stop occurs on the Sixmile Lake Road. Be attentive for automobiles, trucks, and four-wheelers when observing this outcrop.*

Description: This relatively small (3 meters by 5 meters) outcrop exposes the northern margin of the east-west-trending gabbro-diorite sill that extends from at least southwest of Sixmile Lake to at least 500 meters east of this location (Fig. 1-19; Hudak et al., in prep.). The southern two-thirds of the outcrop comprises pale green to pinkish green, epidote- and clinozoisite/zoisite-altered subophitic gabbro which contains up to 1% 1-5mm quartz- and/or actinolite-filled amygdules. Numerous north-northeast, steeply west-dipping joints were developed during the regional D_3 deformation. The uppermost one-third of the outcrop contains an in-situ hyaloclastite breccia comprising pale-to dark green angular, locally jigsaw puzzle-fit, iron chlorite + actinolite + zoisite/clinozoisite-altered gabbro lapilli in a pinkish-tan to tan matrix of sericite, epidote, and zoisite/clinozoisite (Fig. 1-23a).

This outcrop illustrates, in part, the difficulty in determining hydrothermally altered flow top breccias from hyaloclastite breccias and/or peperites that form from intrusion of sills into wet, relatively unconsolidated sediments that occur within the shallow seafloor. The intense hydrothermal alteration that occurs at the northern margin of the outcrop indicates that high water:rock ratio, high temperature (>300°C) hydrothermal alteration was focused in the permeable, sill-marginal, hyaloclastite or peperite (?) breccia. Synvolcanic sills intruded into the shallow seafloor are important heat sources that drive hydrothermal circulation cells that form VMS deposits (Campbell et al., 1981). Sills such as the one exposed at this outcrop may have contributed significantly to the formation of Algoma-type banded iron-formation within the overlying Soudan Iron Formation, and perhaps, VMS mineralization within the uppermost parts of the Lower Ely or the Soudan Iron Formation.

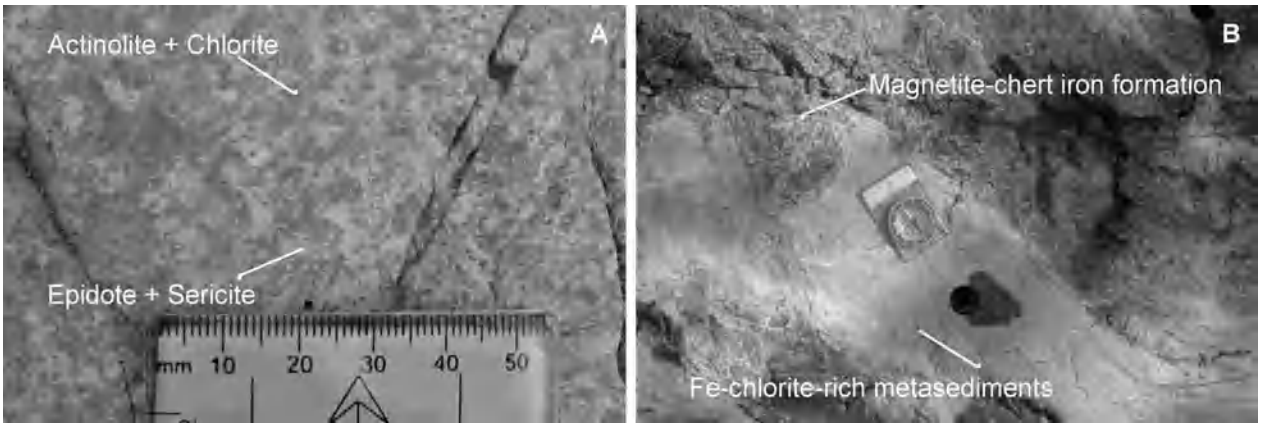


Figure 1-23. Synvolcanic sills and banded iron-formations in the uppermost parts of the Lower Ely. Figure 1-23a shows the uppermost margin of a gabbro/diorite sill that has apparently intruded wet, unconsolidated sediments to form an extremely altered hyaloclastite or sill-marginal peperite. Figure 1-23b shows folded banded iron-formation horizons that occur approximately 10 meters up-section from the sill.

STOP 1-17: <u>BANDED IRON-FORMATION</u>, NORTH OF SIXMILE LAKE ROAD

Location: (SE, NE, S.21, T.62N., R.14W., NAD 83 UTM 565953E, 5298823N)

Description: Two outcrops located within 30 meters north of the Sixmile Lake Road contain relatively thin (<1-2 meters) magnetite-chert horizons interbedded with mafic to intermediate composition volcaniclastic sedimentary strata, and illustrate features common to the uppermost parts of the Lower Ely. Based on detailed mapping by Peterson and Patelke (2003) and Hudak et al. (in prep. a), the uppermost 50-200 meters of the Lower Ely is transitional into the base of the Soudan Iron Formation. Basalt – andesite lava flows and volcaniclastic sedimentary strata decrease in abundance, whereas Algoma-type banded iron-formations become more common as one progresses up-section through this transition zone. This sequence records the change from a period of dominantly effusive submarine volcanism to a period marked by chemical sedimentation.

The outcrop closest to the road contains strongly deformed magnetite-chert iron-formation interbedded with iron-chlorite-rich metasediments. Microfossils are locally present in the Vermilion district iron-formations, and hint at a biogenic contribution to their formation (LaBerge, 1973). Deformation has produced more or less north-northwest striking, steeply dipping beds. The deformation that occurs in these strata is not regionally observed, and is believed to be a local phenomenon resulting from syndepositional intrusion of the mafic sill observed in the previous stop. A later quartz-feldspar porphyry dike occurs at the northeastern end of this small outcrop. Note the localized sulfide staining along the contact between the intrusion and adjacent banded iron-formation.

The second outcrop, located approximately 30 meters north of the road, contains poorly exposed mafic – intermediate volcaniclastic strata and associated banded iron-formation horizons. The best exposures of the banded iron-formations occur at the southeastern edge of the outcrop. At this location, the magnetite-chert iron-formations locally contain veins and stringers of malachite, along with up to 20% pale brown garnet (x-ray analysis performed by Hudak et al. (in prep. a) indicate that the garnet is andradite). The presence of malachite is indicative of the local high concentrations of copper within the uppermost parts of the Lower Ely. Similar copper mineralization within the Soudan Iron Formation has been identified by Peterson and Patelke (2003) in the 12th level of the Soudan Mine.

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FIELD TRIP 2

Geologic Highlights of New Mapping in the Southwestern Sequence of the North Shore Volcanic Group and in the Beaver Bay Complex

Leaders:

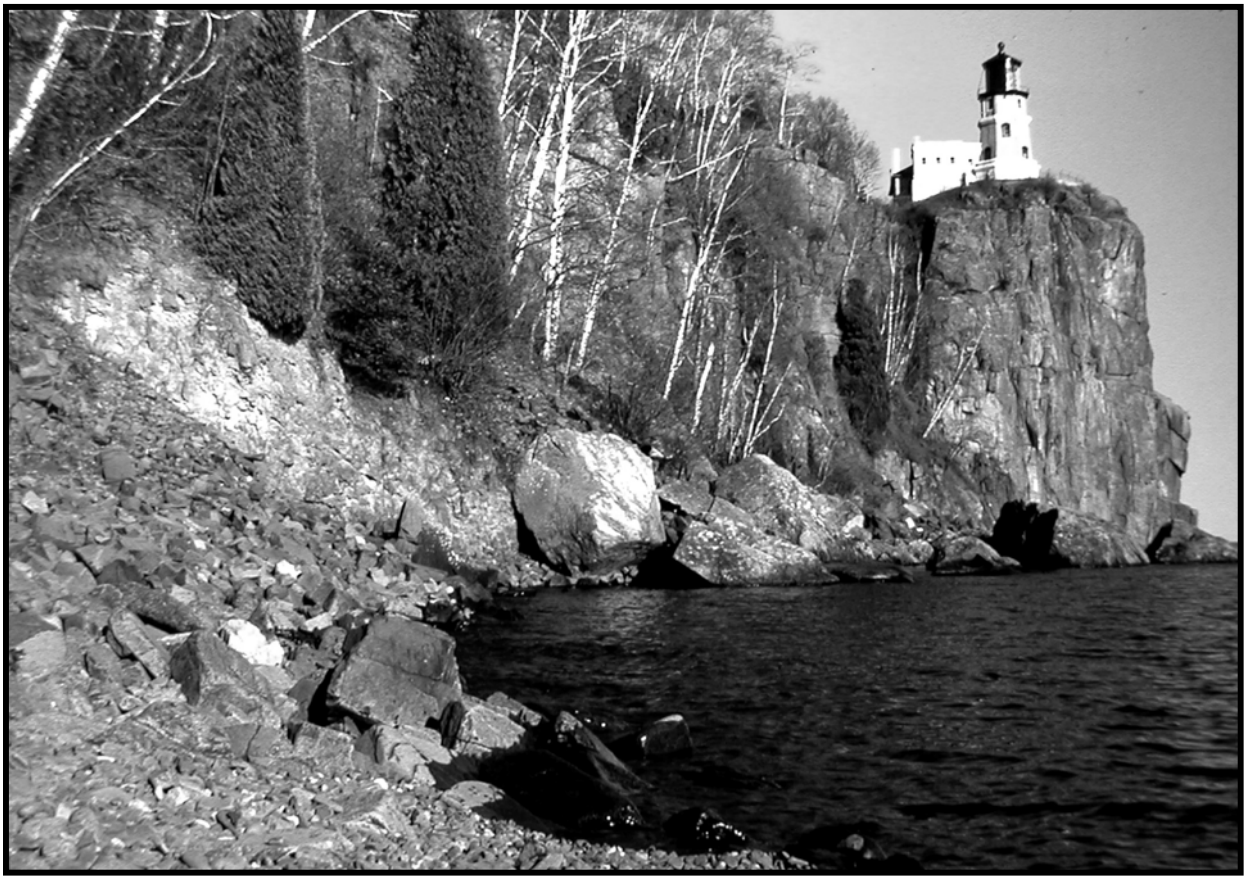
Terry Boerboom, Jim Miller

Minnesota Geological Survey, University of Minnesota

and

John Green

Department of Geological Sciences, University of Minnesota Duluth



Split Rock Lighthouse Point (Stop 2-8)

FIELD TRIP 2

Geologic Highlights of New Mapping in the Southwestern Sequence of the North Shore Volcanic Group and in the Beaver Bay Complex

by

Terry Boerboom, Jim Miller
Minnesota Geological Survey, University of Minnesota
and

John Green
Department of Geological Sciences, University of Minnesota Duluth

INTRODUCTION

For the past 20 years, the Minnesota Geological Survey has been actively conducting detailed (1:24,000-scale) bedrock geologic mapping in northeastern Minnesota. This effort began in 1985 with a mapping project focussed on delineating the geology of the intrusive Beaver Bay Complex in the central part of the region. This eight-year project was partially supported by USGS's COGEOMAP program and resulted in publication of five geologic maps (Miller, 1988; Miller and others, 1989, 1993, 1994; Boerboom and Miller, 1994) covering ten 7.5' quadrangles (Fig. 1). More recently, geologic mapping in northeastern Minnesota has focused on the shoreline quadrangles between Duluth and Split Rock Point, where it joins up with the Beaver Bay mapping. This ongoing mapping, which began in 2001, has produced four geologic maps (Boerboom and others, 2002a, 2002b, 2003a, 2003b) with another currently in production (Fig. 2-1). Mapping in the Two Harbors NE quadrangle, scheduled for this coming field season, and publication of quadrangles of the Duluth metropolitan area later this year, will result in 1:24,000-scale coverage of mappable bedrock along the North Shore from Duluth to Tofte. Long-range plans are to continue to systematically map the shore to the Canadian border.

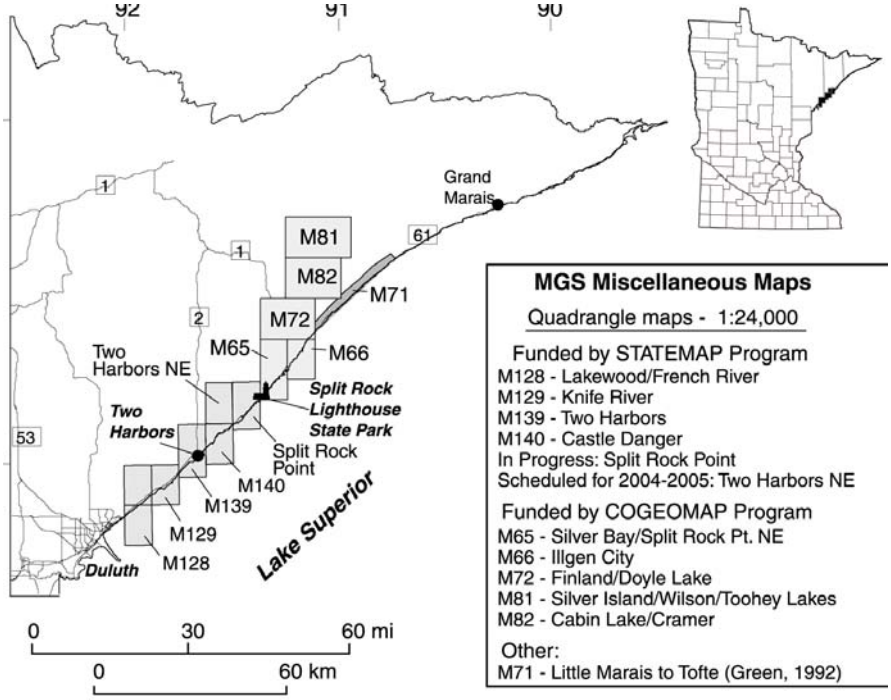


Figure 2-1. North Shore 7.5' quadrangles mapped by the MGS through the USGS-sponsored COGEOMAP and STATEMAP programs.

This field trip is intended to highlight some of the characteristic and more intriguing aspects of North Shore geology that the recent detailed mapping has revealed. Of course, there are many more interesting geologic features than we can show in two days. Hopefully, you will take advantage of the geologic maps that have been produced of this area to explore more of its fascinating geology.

VOLCANIC AND SEDIMENTARY ROCKS OF THE SOUTHWESTERN SEQUENCE OF THE NORTH SHORE VOLCANIC GROUP

by John Green

This overview is a slightly modified expert from the MGS Report of Investigations 58, Chapter 5: "Volcanic and sedimentary rocks of the Keweenawan Supergroup in Northeastern Minnesota" by John Green (Miller and others, 2002).

Magmatic activity related to the 1.1 Ga Midcontinent Rift produced a more than 10 kilometer thick edifice of lava flows and subvolcanic intrusions that are exposed along Minnesota's north shore of Lake Superior. The lava flows and minor sedimentary rocks are referred to as the North Shore Volcanic Group (Goldich and others, 1961) and the intrusive rocks are variably assigned to the Duluth Complex, the Beaver Bay Complex, and miscellaneous intrusions of the Midcontinent Rift Intrusive Supersuite (Fig. 2-3). Although previous publications have subdivided the North Shore Volcanic Group into informal volcanic suites and distinctive flows (Green, 1972, 1982), the 1:200,000-scale map (M-119, Miller and others, 2001) accompanying the report of Miller and others (2002) represents the first time that the North Shore Volcanic Group has been subdivided into formational entities on a geologic map. A brief description of the North Shore Volcanic Group and associated sedimentary formations of the Keweenawan Supergroup is given here to supplement the information provided on the geologic map about these formational units.

Rock classification, recognition, and textures

As a coherent tholeiitic compositional suite, the volcanic rocks of the North Shore Volcanic Group can be described using only a few rock names (Fig. 2-2). The most primitive rocks are olivine tholeiites, which form an iron-enrichment trend with further evolution. They display ophitic textures and pahoehoe surfaces nearly everywhere. Most olivine tholeiites are aphyric, but those that are porphyritic contain dominantly plagioclase phenocrysts, less commonly olivine. Transitional basalts contain somewhat higher alkalies and other incompatible elements than the olivine tholeiites, but generally not enough to classify them as alkalic. Their texture is typically intergranular and fine- to medium-grained. Porphyritic varieties generally contain small phenocrysts of plagioclase, olivine, clinopyroxene, and magnetite. The reversed-polarity Hovland lavas, however, are characterized by transitional basalts (grading to basaltic andesites) that contain abundant, large, tabular, plagioclase phenocrysts. Transitional basalt flow surfaces are generally smooth (pahoehoe), although a few show breccia tops.

The basaltic andesites and andesites (greater than 52 percent SiO_2) are tholeiitic, rather than calc-alkaline; they show iron enrichment and contain only anhydrous ferromagnesian minerals. These rocks are typically fine-grained and intergranular to felty or pilotaxitic, and many contain small phenocrysts of plagioclase, olivine, clinopyroxene, and magnetite. The andesites generally weather to a red-brown color, and have flow-brecciated (aa) tops but not bases. Many flows that contain 50 to 55 percent silica show millimeter-scale oxidation lamination (Green, 1989) parallel to the base. A few highly iron-enriched flows, separable only by chemical analysis, can be called ferroandesites.

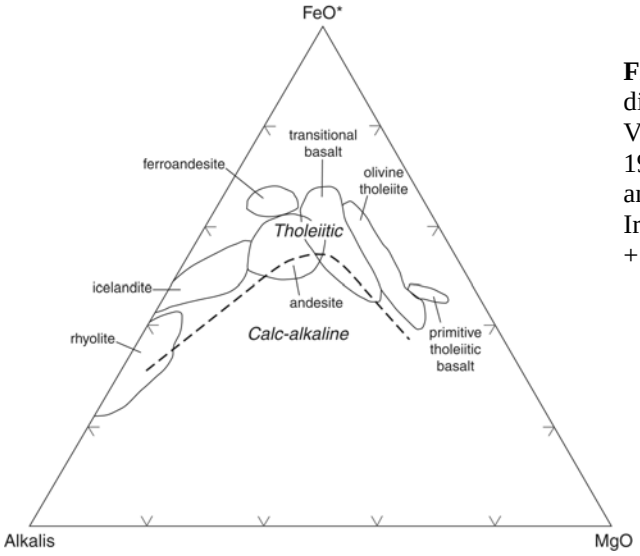


Figure 2-2. AFM compositional diagram for the lavas of the North Shore Volcanic Group (modified from Green, 1982). The boundary between tholeiitic and calc-alkalic rocks is modified from Irvine and Barager (1971). FeO* = FeO + $0.9Fe_2O_3$.

Carmichael (1964) first used the name "icelandite" for rocks intermediate in character between andesites and rhyolites in the Tertiary lavas of eastern Iceland. They might be considered the tholeiitic equivalent of calc-alkaline dacite in orogenic suites. Other examples of these rocks have been described from the Galapagos (McBirney and Williams, 1969) and the Miocene of Nevada–Oregon (Wallace and others, 1980). Very large flows of similar composition in the Etendeka volcanics of Namibia have been referred to as quartz latites by Milner and others (1992). Icelandites in the North Shore Volcanic Group (Green and Fitz, 1993) are characterized chemically by SiO2 contents ranging from 60 to 68 percent, high FeO* (averaging 7 percent), K2O + Na2O values between 6.5 and 9 percent, a potassium/sodium atomic ratio of about 0.9, and an Mg number [Mg/(Mg + Fe) atomic] averaging 0.14. Petrographically, North Shore Volcanic Group icelandites grade continuously from the andesites to somewhat paler colors (brown or tan), but have a similar phenocryst assemblage. Quartz and alkali feldspar are common in the groundmass but never occur as phenocrysts. Flowtop features (crusty to coarsely brecciated) indicate that the icelandites were erupted as lavas.

The rhyolites have higher silica and total alkali contents, and lower FeO than the icelandites. They are generally light gray, pink, or red. Several are very thick, extensive, and voluminous (up to several hundred cubic kilometers; Green and Fitz, 1993). Although most rhyolites are porphyritic (phenocrysts of quartz, alkali feldspar ± plagioclase, and altered fayalite ± ferroaugite), some lack quartz and alkali feldspar phenocrysts, and rare flows are aphyric. Groundmass textures are fine-grained holocrystalline, typically with a meshwork of platy quartz paramorphs after primary tridymite, which may show a flow structure. A "snowflake" texture is common, in which poikilitic quartz patches (coalesced ex-tridymite grains) enclose small, dusty alkali-feldspar grains. In quartz-phyric flows, these poikilitic quartz patches are in optical continuity with adjacent quartz phenocrysts (Green, 1990). Outcrop-scale flow structure, including folding, is common near flow tops and bases. In some flows, distinct fiamme, deformed to varying degrees, are recognizable. These imply an explosive eruption that produced a pyroclastic flow, which welded and remobilized to produce a rheoignimbrite.

Structure and Lithostratigraphy

As stated above, the Midcontinent Rift volcanic rocks and interbedded redbeds in northeastern Minnesota comprise the North Shore Volcanic Group (Goldich and others, 1961; Green, 1972, 1977, 1982; Basaltic Volcanism Study Project, 1981). In general, these rocks form an arcuate stack that is slightly tilted toward the southeast and forms the roof rocks into and under which the Duluth Complex and associated hypabyssal intrusions were emplaced. At the southwest end of the North Shore Volcanic Group near Duluth, the volcanic rocks strike north with a 10º to 20º easterly dip; at the northeast end at Grand Portage, the flows strike east–west with a 10º southerly dip (Fig. 2-3). Thus, traveling northeast along the Lake Superior shore northeast from Duluth, and southwest from Grand Portage, one encounters successively higher flows in the volcanic stratigraphy until the Tofte–Lutsen area in southern Cook County, where the highest exposed flows crop out. Exposure is generally excellent along the eroding lakeshore, and along the lower, high-gradient stretches of tributary streams, providing good control on the stratigraphy. A stack of volcanic rocks approximately 9.7-kilometers-thick has been measured in the "southwest limb" (Table 2.1), and another stack of volcanic rocks about 7.2-kilometers-thick has been measured in the "northeast limb" (See Table 5.2, Miller and others, 2002). This implies nearly continuous subsidence during the rifting process. The difference in stratigraphic thickness between the two limbs reflects major complications in the central area, which appears not to have subsided at the same rate as in the limbs, and into which many of the subvolcanic intrusions were emplaced. Except for the capping Schroeder–Lutsen basalt sequence, no stratigraphic unit can be traced from one limb to the other; each limb has its own stratigraphic column.

To aid in the correlation of intrusive and extrusive rock units throughout the Midcontinent Rift system, their paleomagnetic polarity has been used. Nearly all of the igneous and sedimentary rocks associated with the Midcontinent Rift were formed either during an earlier, reversed-polarity interval or a succeeding normal-polarity interval. Thus, in each limb of the North Shore Volcanic Group, the lower stratigraphic units show reversed polarity, and the upper sequences show normal polarity. This polarity reversal forms the basis for distinguishing upper and lower sequences in the northeast and southwest limbs. U-Pb zircon dates demonstrate that the reversed-polarity magmatism occurred mainly in the time interval from 1108 to 1107 Ma; whereas, around the Lake Superior basin, normal-polarity magmatism occurred mainly in the interval from 1099 to 1094 Ma (for example Davis and Paces, 1990; Paces and Miller, 1993; Davis and Green, 1997). These two pulses were separated by a magmatically inactive time (at least in the upper crust), which appears to be expressed as a slight unconformity in the volcanic sequence on the north shore. However, because intrusions subsequently penetrated along this horizon in the North Shore Volcanic Group, this unconformity has not been recognized in outcrop. The Duluth Complex separates the upper and lower sequences of the southeastern limb (Fig. 2-3, Table 2.1).

One other significant gap in the stratigraphic continuity of the North Shore Volcanic Group occurs near the stratigraphic top, where the Schroeder–Lutsen sequence overlies the upper units of the northeast and southwest sequences (Fig. 2-3, Table 2.1). In the northeast limb, southwest of Grand Marais (Fig. 2-3), the basal flow of the Schroeder–Lutsen sequence (Terrace Point basalt member) overlies a thick sandstone and siltstone unit (Cut Face Creek sandstone), which in turn conformably overlies the Good Harbor Bay andesites of the upper northeast sequence (See Table 5.2, Miller and others, 2002 for details). However, in the southwest limb, the basal Schroeder–Lutsen sequence flow overlies a thinner sandstone and conglomerate unit (the Little Marais conglomerate) that in turn rests in sharp angular unconformity atop structurally disturbed flows of the Belmore Bay lavas near Little Marais (Table 2.1). Furthermore, the gently dipping Schroeder basalts have not been penetrated by the abundant hypabyssal intrusions of the Beaver Bay Complex that complicate the underlying volcanic sequence in this mid-shore area (Green, 1992; Miller and others, 1993). Attempts to date the Schroeder–Lutsen sequence have been unsuccessful to this point.

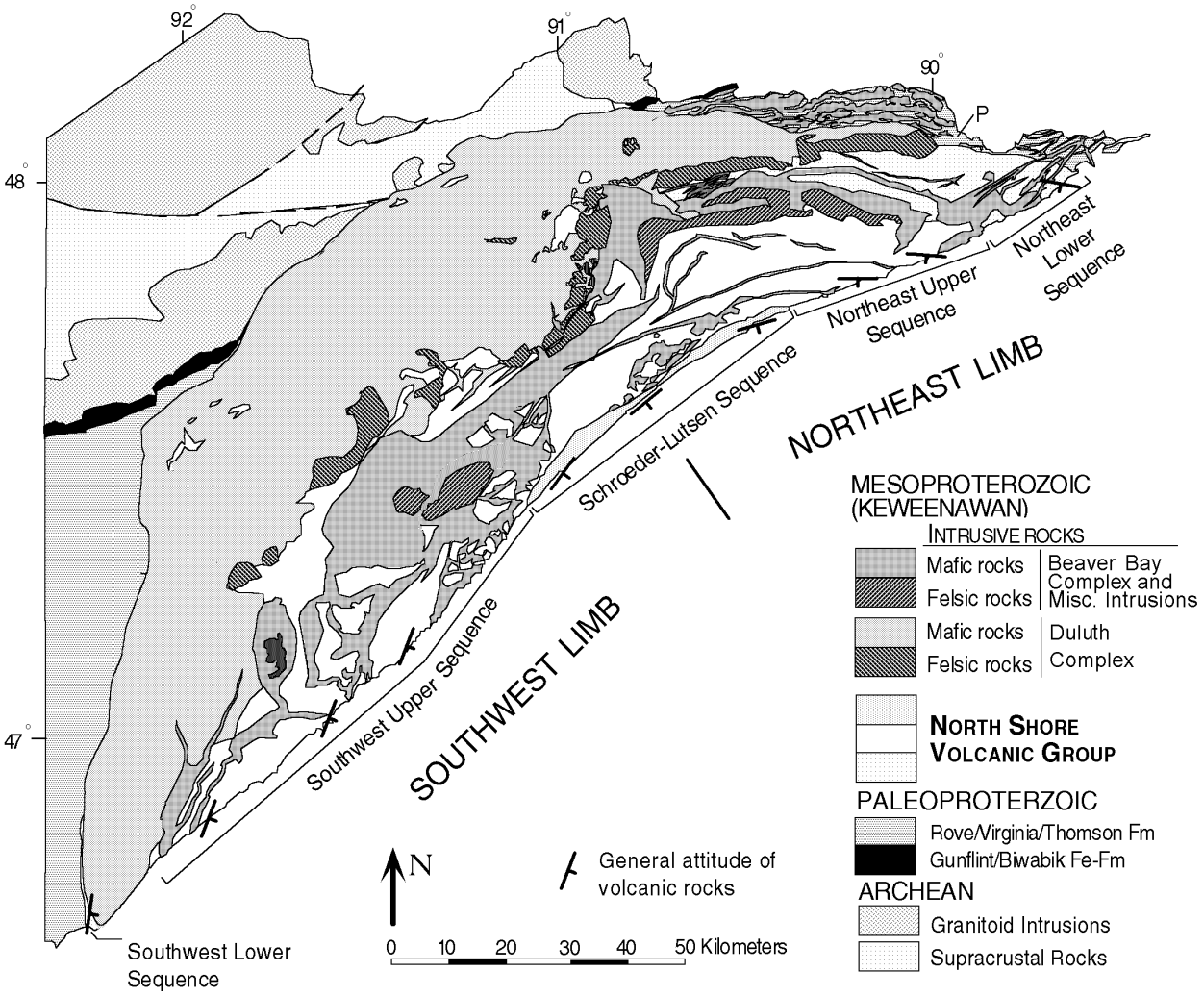


Figure 2-3. Simplified geology of northeastern Minnesota showing the major sequences of the North Shore Volcanic Group (taken from Fig. 5.2, Miller and others, 2002).

The five lithostratigraphic sequences comprising the two limbs of the North Shore Volcanic Group are further subdivided into informal formational units (Table 2-1). Some formational units are individual flows of distinctive lithology and/or substantial thickness and lateral extent. Most are suites of lava flows that have distinct lithologic characteristics or that are separated by intrusions. Some lava formations contain distinct flows or sedimentary rock units within an otherwise homogeneous package of lavas. Such units are given informal member rank (such as the Silver Beaver rhyolite member of the Baptism River lavas, Manitou transitional basalt member of the Schroeder basalts, and Indian Camp sandstone member of the Lutsen basalts). See Chapter 1 of RI58 (Miller and others, 2002) for more details on stratigraphic nomenclature of Keweenwan rock in northeastern Minnesota.

Table 2-1. Generalized stratigraphy of the southwest limb of the North Shore Volcanic Group showing U/Pb ages (Davis and Green, 1997; Green and others, 2001). Positions of intrusions denote approximate stratigraphic level affected and not age of emplacement.

Thickness(m)	Lithostratigraphic units		Lithologic character	U/Pb ages
9735	**Total section**			
	150	Carlton Quarry lavas *(fault bounded)*	basalt, andesite, and rhyolite flows	1094.3±2.0
945	**Schroeder–Lutsen sequence** (normal polarity)			
	900	Schroeder basalts	ophitic olivine tholeiite basalt flows; includes Manitou transitional basalt and Pork Bay breccia	
	<45	Little Marais conglomerate	polymict volcanic conglomerate and sandstone	
	angular unconformity			
8275	**Upper southwest sequence** (normal polarity)			
	565	Bell Harbor lavas	mostly quartz tholeiite basalt and basaltic andesite flows	
	100	Palisade Head rhyolite	gray-pink, porphyritic rhyolite flow	1096.6±1.7
	Beaver Bay Complex			~1096
	700	Baptism River lavas	mixed lavas, mostly basalt; includes 165-meter-thick Silver Beaver rhyolite	
	20	Silver Bay porphyritic basalt	ophitic basalt flow with abundant large plagioclase phenocrysts	
	730	Gooseberry River basalts	mixed basalt flows, mostly ophitic	
	Lafayette Bluff, Silver Creek diabase intrusions			
	315	Two Harbors basalts	mixed aphyric basalt flows; quartz tholeiite flows at base	
	550	Larsmont basalts	ophitic olivine tholeiite flows	
	Stony Point–Knife Island diabase sheet			
	1500	Sucker River basalts	mixed basalt flows, mostly ophitic	
	1350	Lakewood lavas	mostly basalt flows; rhyolite, icelandite, and ferroandesite at base	
	Lester River diabase sill			
	1285	Lakeside lavas	mixed basalt, andesite, icelandite, and rhyolite flows	1098.4±1.9
	Endion diabase sill			
	1160	Leif Erickson Park lavas	mixed basalts, andesites	
	Duluth Complex			~1099
370	**Lower southwest sequence** (reversed polarity)			
	370	Ely's Peak basalts	porphyritic, diabasic, and ophitic basalts; pillowed and pyx-phyric basal flow	
	>8	Nopeming Sandstone	white to tan quartzite and conglomerate	
	angular unconformity			
		Thomson Formation (Paleoproterozoic)		

Physical volcanology

The volcanic rocks of the Midcontinent Rift, including the North Shore Volcanic Group, represent one of the world's oldest and best-preserved examples of plateau lavas. However, they are not typical in that they contain a greater thickness of flows and, in the North Shore Volcanic Group, a higher proportion of evolved compositions. They are similar physically and chemically to the Tertiary lavas that make up eastern Iceland (Sigvaldason, 1974; Green, 1977; Wood, 1978), and they formed similarly over a plume at another major rift. The rocks also resemble the late Tertiary and Quaternary volcanic rocks of the southern Snake River Plain, Idaho and southeastern Oregon, because of their interbedded basalts and large rhyolites (e.g., Bonnichsen and Kauffman, 1987; Manley, 1996).

The basalts range in character from typical flood flows as voluminous as tens of cubic kilometers to more modest, "plains-type" flows (Greeley, 1982) and thin flow units less than a meter thick. At Duluth in the southwest limb, and Grand Portage in the northeast limb, the lowest flows in the volcanic sequences are pillowed, and thus inferred to have erupted subaqueously; however, nearly all of the other flows were erupted subaerially. The flows show different physical characteristics, closely tied to their chemical compositions and viscosities (Green, 1989; Green and Fitz, 1993). Olivine tholeiites, which dominate the North Shore Volcanic Group, all have pahoehoe surfaces, with or without ropy structures. Other physical characteristics of the various rock types were previously discussed in the "Rock classification, recognition, and textures" section of this guide.

All of the flows ranging in composition from basalts to icelandites were erupted as lavas. The rhyolites are notable in their abundance relative to other plateau-lava sequences, their size (up to several hundred cubic kilometers), and extent (Green and Fitz, 1993). Several rhyolites show textural evidence of rheomorphic flow after eruption as ash-flow tuffs, though some were lavas. One of the largest, the Devil Track rhyolite in Cook County, which is as thick as 250 meters and can be traced for 40 kilometers along strike, has ambiguous features that make its mode of eruption difficult to discern; it may be a lava flow. Nearly all the icelandites and rhyolites show evidence of an unusually high temperature of eruption, such as magmatically crystallized groundmass tridymite. The evidently low viscosity of these large rhyolites is attributed to their high temperature, high iron and fluorine contents, and low oxidation state (Green and Fitz, 1993).

Geochemistry and chemostratigraphy

The North Shore Volcanic Group constitutes a subalkalic, tholeiitic suite that ranges continuously from rather primitive olivine tholeiite to rhyolite, and shows a strong iron-enrichment trend (Fig. 2-2; also Basaltic Volcanism Study Project, 1981; Brannon, 1984). However, relative abundances are strongly bimodal; basalts are greatly predominant, but rhyolites make up 10 to 25 percent of the section. The basalts show trace element and isotopic evidence of derivation mostly from a mantle plume (Nicholson and others, 1997), whereas most of the rhyolites include major contributions from partial melting of the Archean basement (Vervoort and Green, 1997). The most common basalt type, ophitic olivine tholeiite, is generally aluminum-rich (16 to 18 percent Al2O3); the most primitive flows have Mg numbers of about 0.65 to 0.68.

The basal few flows in both limbs of the North Shore Volcanic Group have a unique geochemical and petrographic character. Typically they contain augite phenocrysts, are aluminum-poor, and are rich in both compatible (chromium and nickel) and incompatible elements (titanium, phosphorus, and lanthanum) with steep chondrite-normalized lanthanum/ytterbium ratios. This suggests derivation by relatively small-fraction melting of the initial plume head (Nicholson and others, 1997; Green, 1995).

In general, there is little stratigraphic regularity of compositional change within the North Shore Volcanic Group, with the following exceptions. In the middle of the upper southwest sequence, there is a marked upward progression toward more primitive compositions through a 3.4-kilometer section from rhyolite east of the Lester River into a thick group of primitive olivine tholeiites in the Knife River–Two

Harbors area (Brannon, 1984). This includes the Lakewood lavas, the Sucker River basalts, and the Larsmont basalts. In contrast, in the lower northeast sequence, the approximately 1-kilometer-thick basal Grand Portage lavas progress upsection from basalt to increasingly evolved compositions, ending with Red Rock rhyolite (Green, 1995). As mentioned above, the Schroeder–Lutsen sequence, the youngest in the North Shore Volcanic Group, is composed almost entirely of olivine tholeiites.

All of the North Shore Volcanic Group has been affected to some degree by hydrothermal/burial metamorphism. The more permeable (fractured, vesicular) tops and bases of the flows have undergone considerable mineralogical change (deposition of amygdule minerals, alteration of primary minerals), but in many cases the massive flow interiors are remarkably little-altered. Where alteration has approached equilibrium, mineral assemblages range from lower greenschist facies at the base of the North Shore Volcanic Group to zeolite facies at the top (Schmidt, 1993; Schmidt and Robinson, 1997).

Interflow Sedimentary Rocks

Clastic redbed strata occur at many horizons within the North Shore Volcanic Group (Jirsa, 1984). They are lenticular and range in thickness from a few centimeters to about 100 meters. As these rocks are relatively soft and erodable compared to the adjacent volcanic flows, they are mostly covered and are exposed only along actively eroding sites such as streambeds and the lakeshore. They are predominantly red to brown, well sorted sandstone, with minor conglomerate, siltstone, and shale. Conglomerate beds are most abundant in the midshore area from Little Marais to Lutsen.

Compositionally, these redbeds are mainly immature lithic arkose and feldspathic lithic arenite (see Fig. 5.3, Miller and others, 2002). The angular to subrounded clasts are mainly plagioclase, mafic to felsic volcanic rock fragments, clinopyroxene, and Fe-Ti oxides; devitrified or replaced volcanic ash particles and shards are present in a few beds. Quartz is uncommon to absent. The framework grains have been variably cemented with hematite, calcite, prehnite, and a variety of zeolites, depending on the local hydrothermal/burial/contact metamorphic conditions. In some places hydrothermal minerals have replaced many or most of the clasts.

A few of these redbed units have thicknesses in excess of 25 meters. These include a cross-bedded sandstone in Leif Erickson Park in Duluth (35 meters), which disconformably overlies an eroded basalt flow; the Little Marais conglomerate (and sandstone) exposed in the Manitou River area near Little Marais (as thick as 45 meters); the Indian Camp sandstone (68 meters) northeast of Lutsen; and the Cut Face Creek sandstone southwest of Grand Marais (100 meters), which can be traced for at least 4 kilometers along strike. Of these, the Little Marais and Cut Face Creek units occur at the base of the Schroeder–Lutsen basalt sequence. The sandstone in these units is typically planar- or cross-bedded, and some beds are ripple-marked or mud-cracked. The rocks are inferred to be dominantly fluvial, deposited by moderate-gradient, east- to southwest-flowing streams from sources nearly entirely within the subsiding Midcontinent rift basin (Jirsa, 1984).

Many flow-top breccias of andesite and basalt with aa structure contain laminated red sandstone as a matrix because sand filtered down from the flow surface. Similarly, red, laminated sandstone and siltstone form clastic dikes or crevice-fillings a few centimeters wide in the upper parts of some lava flows.

INTRUSIONS WITHIN THE SOUTHWESTERN SEQUENCE OF THE NSVG

by Terry Boerboom

Several hypabyssal sills and sheet-like intrusions occur within the 7-kilometer-thick section of volcanic rocks exposed near Lake Superior from Duluth to Split Rock Point. From lowest to highest in the section, these consist of the Endion Sill, the Northland sill, the Lester River Sill, the Stony Point diabase, the Silver Creek diabase, the Lafayette Bluff diabase, and the Split Rock intrusive felsite. These sheets are variably concordant to the volcanic strata, range from tens of meters to over 500 meters thick, and generally dip gently eastward. They commonly form prominent topographic highlands. This discussion will be limited to those sills visited by this field trip. For a more complete description, see Miller and others, 2002.

Lester River diabase

The Lester River diabase is named for the outcrops adjacent to the mouth of the Lester River (Schwartz and Sandberg, 1940). The diabase forms a prominent topographic high (Moose Mountain) that extends about 6.5 miles inland from Lake Superior.

The diabase forms a sill some 400-m (1,300-feet) thick that dips an average of 19 degrees southeast, conformable with the dip of the surrounding volcanic rocks. The sill generally overlies basalt and underlies rhyolite, the latter of which probably formed a density barrier that trapped the more dense mafic magma as it was emplaced. The upper margin of the sill contains abundant intermediate to felsic rocks that form a continuous, mappable unit along the entire mapped length of the sill, in contrast to the lower margin which contains only thin and discontinuous pods of the hybrid felsic rocks. The hybrid cap rocks likely formed by partial melting of the overlying rhyolite by heat generated from the underplating sill. Jerde (1991) provides considerable geochemical, petrographic, and petrologic data on the Lester River Sill.

Most of the Lester River diabase is a dark grayish-black, medium- to coarse-grained, massive to locally weakly porphyritic diabase. The diabase contains approximately 60% plagioclase, 13% intergranular to subophitic augite, 12% variably altered olivine, 0-4% orthopyroxene, 1-4% Fe-Ti oxides, 0-10% quartz and granophyre, 1-3% diktytaxitic chlorite, and less than one percent apatite, biotite, and hornblende. Quartz commonly contains wispy rutile needles.

Silver Creek diabase

The Silver Creek diabase is named for proximity to Silver Creek, near Silver Cliff, located about 4 miles northeast of Two Harbors. This diabase is composed almost entirely of medium-grained ophitic olivine diabase, with augite oikocrysts from 1 to 5 centimeters in diameter. It forms a subcordant to discordant subhorizontal sill-like intrusion, at least 60-m (200-feet thick), that forms a prominent highland projecting inland about 7 kilometers north-northeast of Lake Superior, and is exposed intermittently for several kilometers beyond. In the area of the Encampment River, the diabase is in subvertical contact with surrounding volcanic and earlier intrusive rocks; this may have been the feeder to the sill. The diabase is free of anorthosite inclusions and contains only rare xenoliths of volcanic rocks. Pope (1976) studied the Silver Creek diabase as part of Master's thesis under the guidance of Dr. John Green. The Silver Cliff tunnel on Highway 61 cuts through this diabase.

Lafayette Bluff diabase

The Lafayette Bluff diabase is named after prominent exposures at Lafayette Bluff, forms prominent knobs near Lake Superior but is typically weathered to a brown crumbly grus. It is an irregular, discordant, sheet- to dike-like body composed of soft, dark greenish-black, amygdaloidal and porphyritic

diabase, with local differentiated masses of coarse-grained to pegmatitic ferromonzodiorite inland from Lake Superior. The diabase contains locally abundant 1-3 meter diameter inclusions of anorthosite and scattered xenoliths of metamorphosed basalt. The Lafayette Tunnel on Highway 61 has been excavated through the Lafayette Bluff diabase.

Split Rock intrusive felsite

The Split Rock intrusive felsite (rhyolite) is a pink, fine-grained, weakly porphyritic, flow-banded rock that has a prominent shingle parting to it (hence: Split Rock River). At a given location, the rhyolite has all the physical attributes of a rhyolite flow, but detailed mapping shows that it forms a north-striking, 3.4 kilometer wide body that cuts north from Lake Superior across the stratigraphy of the surrounding volcanic rocks. The felsite forms cliffs on Lake Superior south of the Split Rock River, and in the Split Rock River and nearby streams forms small, sharp canyons. Field measurements show that the flow banding dips gently towards the central axis of the body, implying that the intrusion is keel-shaped. The lower margins of the rhyolite are intruded by, or commingled with, a dark gray, fine-grained ferrodiabase or ferrodiorite. This felsite is exposed in the Split Rock Point 7.5' quadrangle, which is currently a work in progress as one of the USGS STATEMAP series of geologic maps to be published by the Minnesota Geological Survey. Due to time constraints, this field trip will not visit this unit.

GEOLOGY OF THE BEAVER BAY COMPLEX

by Jim Miller

The Beaver Bay Complex (BBC) is a hypabyssal, multiple-intrusive igneous complex that was emplaced into the upper part of the NSVG over a 600-km^2 area in northeastern Minnesota (Fig. 2-4). Much of this area was the focus of detailed bedrock mapping by the Minnesota Geological Survey between 1985 and 1992 (Miller, 1988; Miller and others, 1989, 1993a, 1994; Boerboom and Miller, 1994). Three general areas of the BBC, southern, northern, and eastern, are distinguished on the basis of distinctive rock types and intrusion form (Miller and Chandler, 1997). This field trip will investigate some of the units composing the southern and northern BBC. The relationship of BBC intrusions to other subvolcanic intrusions within the NSVG (Fig. 2-4) is unclear, because of poor exposure to the southwest and insufficient mapping to the northeast. Within the mapped area of the BBC, thirteen intrusive units have been identified that represent at least six major intrusive events (Miller and Chandler, 1997). Most intrusive activity forming the BBC occurred around 1096 Ma based on U-Pb dates of 1095.8±1.2 Ma for a Silver Bay intrusion, the youngest unit of the BBC, and 1096.1±0.8 Ma for the Sonju Lake intrusion (Paces and Miller, 1993). Whether activity overlapped the main stage of Duluth Complex magmatism at 1099 Ma is unknown, because attempts to date the oldest component of the BBC were not successful (Paces and Miller, 1993). The boundary between the BBC and Duluth Complex is generally marked by a northeast-trending keel-shaped intrusion in the northern BBC (Houghtaling Creek troctolite) that separates largely dike and sill intrusions of the BBC to the southeast from massive granophyric granite and extensive areas of structurally complex gabbroic anorthosite to the northwest that are typical of the roof zone of the Duluth Complex.

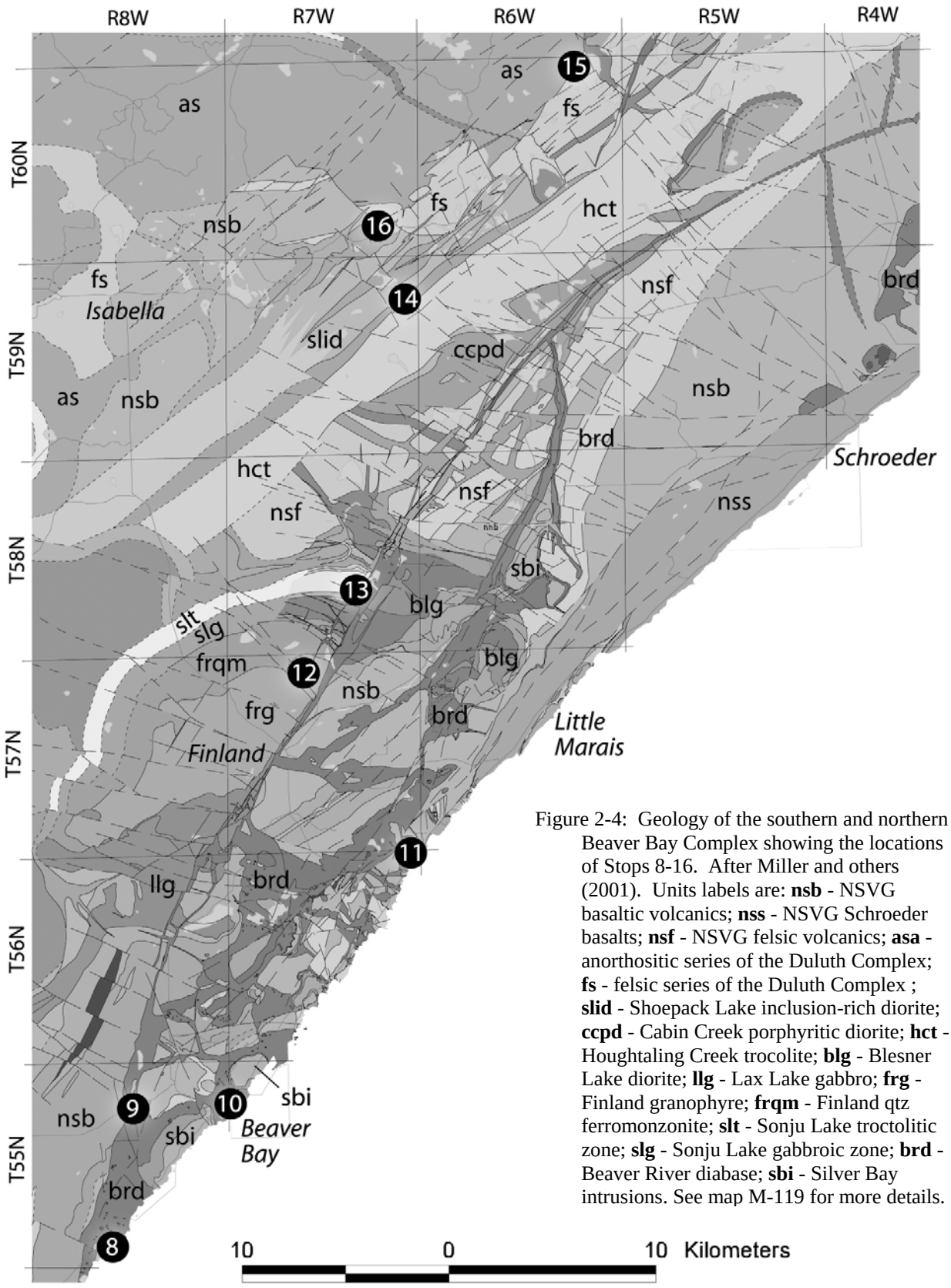


Figure 2-4: Geology of the southern and northern Beaver Bay Complex showing the locations of Stops 8-16. After Miller and others (2001). Units labels are: **nsb** - NSVG basaltic volcanics; **nss** - NSVG Schroeder basalts; **nsf** - NSVG felsic volcanics; **asa** - anorthositic series of the Duluth Complex; **fs** - felsic series of the Duluth Complex ; **slid** - Shoepack Lake inclusion-rich diorite; **ccpd** - Cabin Creek porphyritic diorite; **hct** - Houghtaling Creek trocolite; **blg** - Blesner Lake diorite; **llg** - Lax Lake gabbro; **frg** - Finland granophyre; **frqm** - Finland qtz ferromonzonite; **slt** - Sonju Lake troctolitic zone; **slg** - Sonju Lake gabbroic zone; **brd** - Beaver River diabase; **sbi** - Silver Bay intrusions. See map M-119 for more details.

The range of BBC parent magma compositions is similar to the olivine tholeiite and transitional basalt compositions that dominate the NSVG (Fig. 2-5A). Moreover, like the NSVG, the sequence of intrusion of BBC magmas generally involved progressively more primitive compositions. Compositional variations within the various intrusive units developed as a result of *in situ* magmatic differentiation (Fig. 2-5B), assimilation of footwall rocks, and/or composite intrusions of evolved magma from deeper staging chambers (Fig. 2-5C). The tightly clustered trend of BBC parent magma compositions evident on an AFM diagram (Fig. 2-5A) and the systematic variation of other elemental abundances suggest that all mafic BBC magmas evolved from a common olivine tholeiitic primary magma type. Such a primary composition, which is approximated by the most primitive, high-Al olivine tholeiites of the NSVG, is thought to have given rise to most Midcontinent Rift magmas, especially in later stages of magmatism (Green, 1982; Miller and Weiblen, 1990; Klewin and Shirey, 1992). That even the most primitive of the BBC intrusions, the Beaver River diabase, is significantly evolved from a primitive olivine tholeiite composition (Fig. 2-5A) indicates that all BBC parent magmas were generated in turn by magmatic differentiation of such a primary composition in deeper staging chambers. Petrologic modeling of some BBC intrusions and other hypabyssal bodies that intruded the NSVG (Jerde, 1991) suggests that most magmas experienced multistage, polybaric fractionation between their extraction from the mantle and their subvolcanic emplacement. Although the available radiometric ages do not indicate an overlap of magmatic activity between the BBC and the Duluth Complex (Paces and Miller, 1993), additional dates and more detailed petrologic studies may show that some Duluth Complex intrusions acted as the final levels of staging and differentiation of some BBC-bound magmas.

The focus of emplacement of BBC intrusions appears to have migrated toward the rift axis and toward higher stratigraphic levels with time, perhaps reflecting plate drift and thickening of the volcanic pile. Over the exposed extent of the BBC, intrusion shapes appear to have been controlled by a shallow crustal ridge (Schroeder Forest Center Crustal Ridge) which trends northwest across the BBC. The presence of this buried crustal ridge is indicated by a pronounced saddle in the gravity high over northeastern Minnesota (Chandler, 1990) and the presence of Archean-like inclusions of granitic gneiss, biotite schist, metagraywacke and granodiorite in early intrusions over the gravity low (Boerboom, 1994). The broad network of dikes and sheets, characterizing the southern BBC, becomes tightly focused into a narrow zone of subparallel dikes in the northern portion of the BBC that is situated over the gravity minimum. The eastern BBC opens up again into thick sheet intrusions. Based on geologic, geochronologic, geophysical, and geochemical evidence, Miller and Chandler (1997) suggested that the BBC, particularly the youngest Beaver River diabase dike and sheet network, acted as a magma conduit and structural boundary to the formation and infilling of the western end of the Portage Lake Volcanic basin during the main to late stages of rift volcanism and graben formation. On the North Shore, these volcanic rocks are represented by the Schroeder-Lutsen basalts (Fig. 2-2).

The second half of this trip will investigate seven intrusive units of the BBC and rocks comprising the roof zone of the Duluth Complex (Fig. 2-4). Stops 2-8 through 2-11 will investigate the composite intrusions of the Beaver River diabase and Silver Bay intrusions, the youngest intrusive components of the BBC. Stop 2-12 will visit typical exposures of the Finland granophyre. Stop 2-13 will traverse the middle cumulate units of the Sonju Lake Intrusion - the most completely differentiated intrusion exposed in the Midcontinent Rift system. Stop 2-14 will examine the units defining the northeastern limit of the Beaver Bay Complex - the Houghtaling Creek troctolite, Dam Five gabbronorite, and the Shoepack Lake inclusion-rich diorite. Finally, Stops 2-15 and 2-16 will view anorthositic and granophyric rocks that comprise the roof zone of the Duluth Complex. For more detailed information on these and other units of the Beaver Bay Complex see reports by Miller and Chandler (1997) and Miller and others (2002, Chapter 7) and the 1:24,000 bedrock geologic maps of the area (Miller, 1988; Miller and others, 1989, 1993, 1994; Boerboom and Miller, 1994).

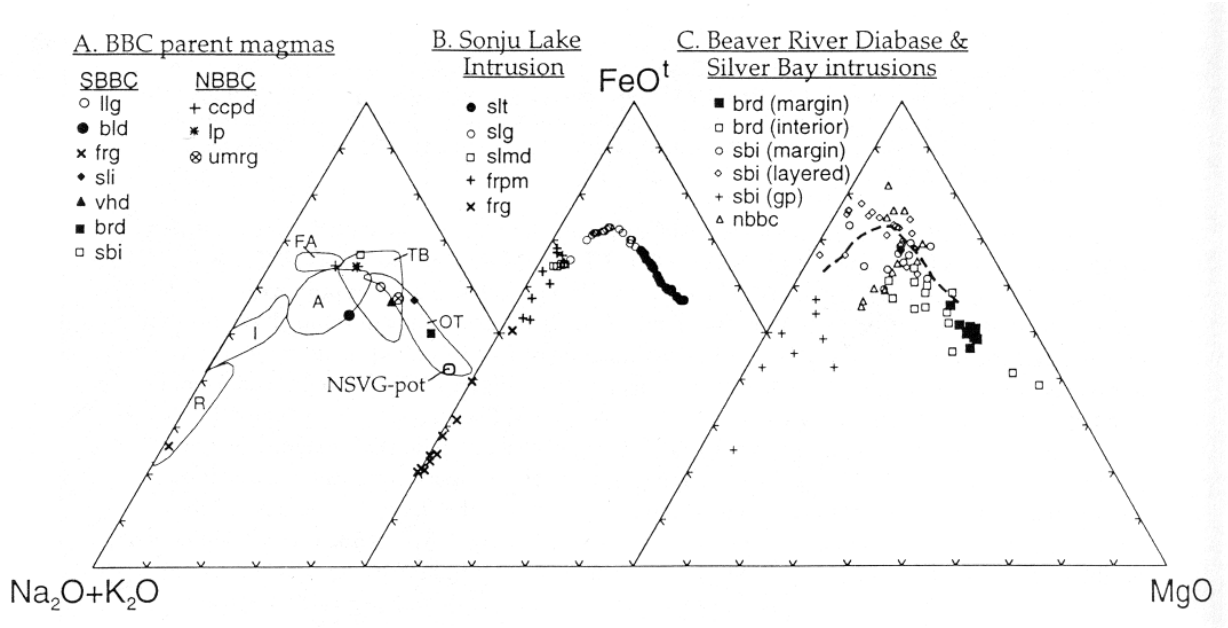


Figure 2-5: AFM diagrams of Beaver Bay Complex (BBC) intrusions. A) Plot of estimated parental magma compositions to various BBC intrusions (see Miller and Chandler, 1997 for unit descriptions and details) compared to major NSVG lava compositions: OT - olivine tholeiite, TB - transitional basalt, A - andesite, FA - ferroandesite, I - icelandite, R - rhyolite (after Green, 1983). NSVG-pot is a primitive NSVG olivine tholeiite composition. B) Calculated liquid line of descent of the Sonju Lake intrusion through troctolitic (slmt-sld-slt), gabbroic (slg-slfg-slad) and monzodiorite (slmd) intervals of the layered sequence. Also plotted are whole rock compositions of Finland granite (frg) and quartz ferromonzodiorite (frpm). C) Whole rock composition plots of Beaver River diabase (brd; ophitic margins and coarse subophitic interiors distinguished), Silver Bay intrusions (sbi; coarse marginal facies, layered ferrogabbroic cumulate interiors, and granophyric compositions distinguished), and composite intrusions from the northern BBC (nbbc). Sonju Lake intrusion differentiation trend (dashed line) is also shown.

FIELD TRIP STOPS

The general locations of all stops are shown on Figure 2-6. A more detailed location map is included with each stop description, with a base made at an appropriate scale from a portion of a 7.5' quadrangle map. This should allow the return visitor to accurately locate the features for each stop. The UTM coordinates listed are all given in NAD83, with easting listed first and northing listed second. The regional setting of the various rock units are given in the introduction, and only facts pertinent to each specific stop are given below.

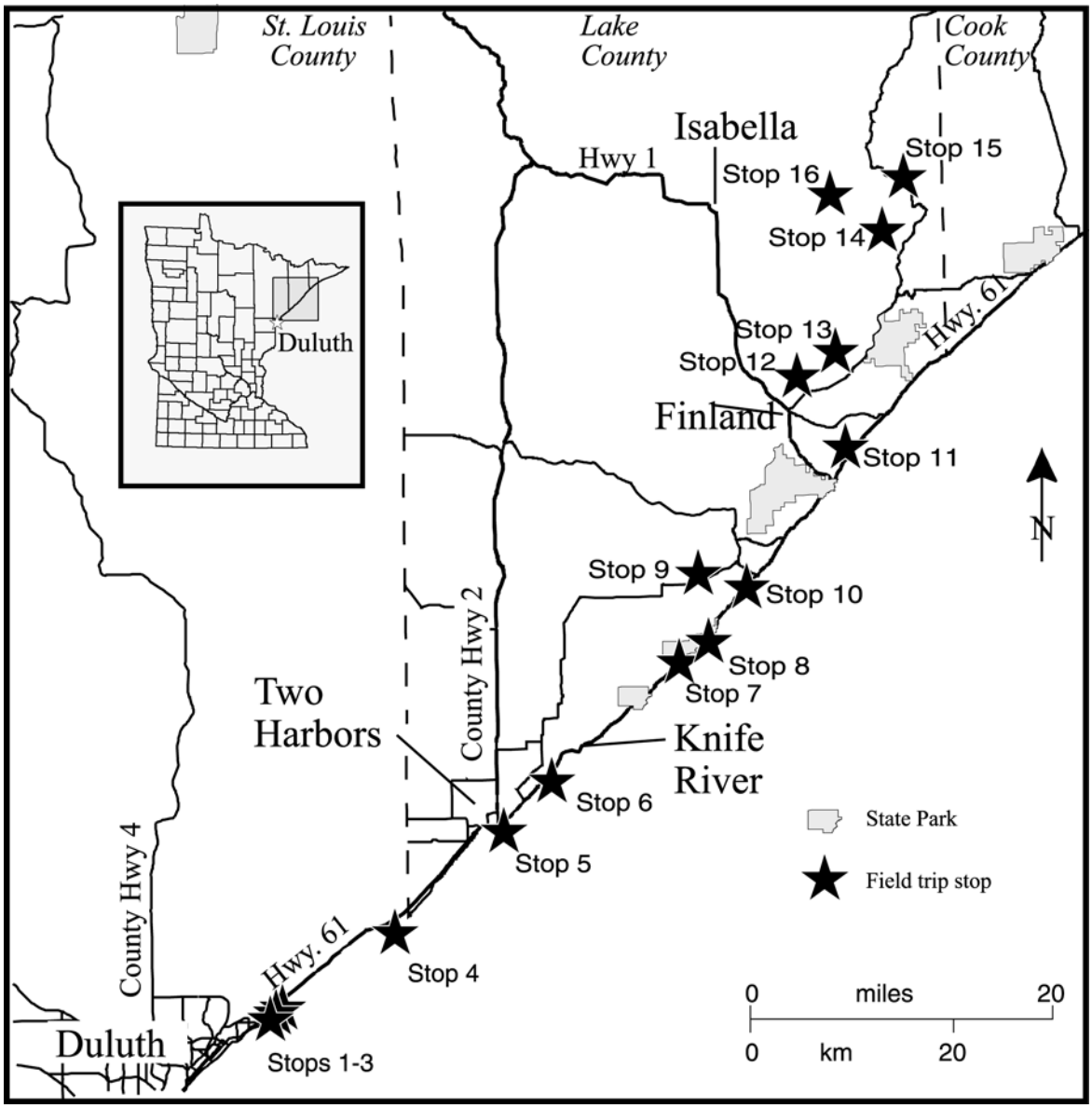


Figure 2-6. Location map for stops for Field Trip 2.

DAY 1

GEOLOGY OF THE SOUTHWESTERN SEQUENCE OF THE NORTH SHORE VOLCANIC GROUP FROM DULUTH TO SPLIT ROCK POINT

Drive Highway 61 from Duluth towards Two Harbors. At the outskirts of Duluth, just after the Lester River, veer right onto Scenic Highway 61 for approximately one-half mile, then turn right on the road that leads to Kitchee Gammee park. Park near the northeast end of this road and walk down to lakeshore.

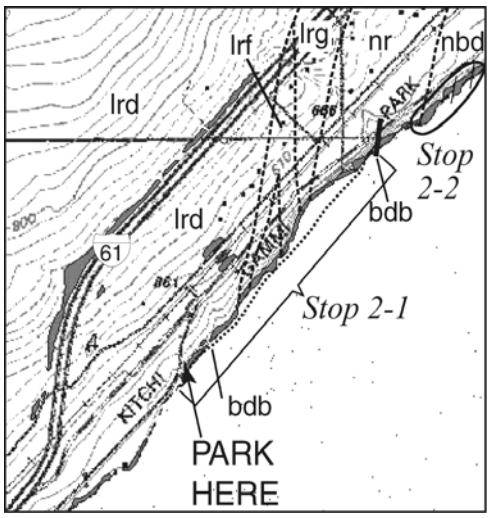
STOP 2-1: Lester River Diabase in contact with rhyolite

T.50N., R.13W., Sec. 4

Lakewood 7.5' quadrangle.

UTM: Start at SW 576929E, 5188417N;
End at NE 577390E, 5188981N

Highlights: Lester River diabase (lrd), hybrid granophyric rocks (lrf, lrg), rhyolite (nr); late diabase dikes (bdb), glacial striations.



The Lester River Diabase is named for the outcrops adjacent to the mouth of the Lester River (Schwartz and Sandberg, 1940). The diabase sill dips southwest about 19 degrees, and forms a prominent topographic high (Moose Mountain) that extends about 6.5 miles inland from Lake Superior. This diabase is mostly intergranular in texture, but does locally exhibit ophitic textures, as seen at this stop. The dominantly intergranular to relatively fine ophitic (less than 1.5 centimeter) texture of the Lester River diabase contrasts with most other Keweenawan diabases such as the Silver Cliff and Beaver River diabases, which typically exhibit a coarse ophitic texture.

This stop starts in the massive interior of the sill, and traverses northeast through increasingly felsic rocks that range from ferrodiorite to granite, and ends at a cliff of fine-grained chilled diabase that cuts the preceding rocks. Felsic rocks first show up as irregular felsic dikes that cut the diabase, but moving northeast towards the rhyolite not only do the felsic dikes increase in abundance, but the diabase itself becomes increasingly granophyric until it grades into dioritic rocks. The granophyric felsic hybrids blend into pink, fine-grained, porphyritic rhyolite, informally termed the Lakewood rhyolite (Green and Fitz, 1993). The exact contact between rhyolite and granophyric hybrid rocks is hard to pinpoint due to the similarity in textures between recrystallized rhyolite and the hybrid melts. Note the thin, shoreline-parallel, fine-grained diabase dikes that cut the Lester River diabase.

On a map scale, the granophyric hybrid phases of the diabase are thick and continuous along the upper sill margin, in contrast with the sill base where hybrid rocks are thin and discontinuous. The diabase sill is sandwiched between rhyolite in the hanging wall, and basalt in the footwall, indicating that the rhyolite acted as a density barrier that trapped and ponded the mafic magma beneath it. The irregularly mixed and distributed hybrid rocks that cap the Lester River diabase are likely due to mixing of remobilized (melted) rhyolite with the underlying mafic magma.

--- Climb up bank , walk NE approximately 0.1 mile to small clearing on east side of highway, climb down to lakeshore, stop 2.2. Alternatively walk to next wayside, climb down, and walk back southwest along lakeshore to basalt/rhyolite contact.---

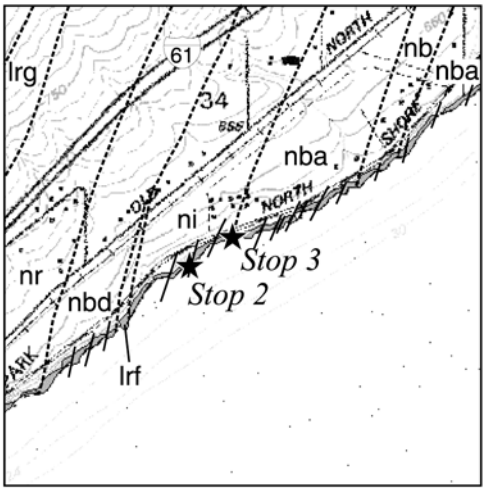
STOP 2-2: <u>Flow contact between rhyolite and basalt of olivine tholeiite composition</u>

T.51N., R.13W., Sec. 34

Lakewood 7.51 quadrangle

UTM: 577648E, 5189145N

Highlights: Contorted flow-banded rhyolite (nr), contact of basalt (nbd) with rhyolite, ropy flow tops in olivine tholeiite ('diabasic'-textured) basalt flows.



Walk southeast to the pink-tinted rhyolite flow that exhibits highly contorted and folded flow banding. Rhyolite contains amygdules of calcite and minor fluorite (please don't beat up the outcrop with hammers), and is capped by a thin, discontinuous layer of fine-grained sandstone. Move northeast over four basalt flows and observe flow features such as (bottom to top of flow) pipe vesicles, vesicle cylinders, amygdaloidal zones in upper portions of flows, and ropy pahoehoe upper surfaces. The pahoehoe crusts indicate a general southeast flow direction.

These basalt flows are of weakly alkaline, transitional intermediate olivine tholeiite composition, with typical olivine tholeiite flow structures. On recently published 1:24,000 scale geologic maps (e.g. Boerboom and others, 2002a), flows such as this are classified as "diabasic-textured' flows. Our field mapping has shown that the textures of the basalt flows does not always correspond to a unique chemical composition, but rather may have more to do with the amount of volatiles in the lava. Some of the most coarse-grained diabasic-textured basalt flows were mapped in the Sucker River upstream from Old North Shore Road. Despite being relatively thin flows (1-3 meters thick), they exhibit a medium- to coarse-grained felty diabasic texture. Some have a coarse pegmatitic zone located about one-third of the way down from the top of the flow, attributed to the concentration of volatiles in this part of the flow as cooling proceeded from the top and bottom. These coarse-grained flows also have unusually prominent pipe amygdules at the base overlain by a zone of bead-like pipe amygdules/vesicle cylinders up to one meter in length that end below the most coarse-grained part of the flow.

The flows observed in stops 2-1 through 2-3 are all part of the Lakewood lavas (see table in introduction), a lithostratigraphic unit characterized by multiple thin 'packages' that range in composition from rhyolite to olivine tholeiite. Each flow package consists of multiple flows of like composition. Brannon (1984) conducted a detailed sampling regime of virtually every flow exposed along the shoreline from Duluth to Two Harbors, starting with the rhyolite just above the Lester River diabase. Her geochemistry verifies the mapped stratigraphy and shows that the most silicic rocks (rhyolite, icelandite, and andesite) occur at the base of the Lakewood lava sequence, whereas primitive tholeiitic basalts dominate the top of the sequence. Vervoort and Green (1997) have shown, through Nd isotopic studies, that the NSVG rhyolites are the product of melting of Archean crustal rocks by underplated primitive

mafic magma, whereas icelandite and basalt are most likely derived directly from a mantle plume source. In their model, early volcanism (lower reversed sequence) tapped a mantle source with little crustal contamination. This was followed by a period of volcanic quiescence (the 1107 -1100 Ma 'volcanic hiatus') during which a mantle plume underplated and melted the crust, generating superheated felsic melts that were later erupted as rhyolite flows and rheoignimbrites, of unusually wide areal extent given their thickness (Green and Fitz, 1993)

--- Climb up bank of small stream to wayside rest (this is the third wayside rest on Scenic Highway 61 past Brighton Beach Road), drive or walk approximately 0.3 miles northeast to wayside rest directly across from Old North Shore Cottages Resort, stop 2-3.---

STOP 2-3: <u>Interflow sandstone between icelandite and basaltic andesite flows</u>

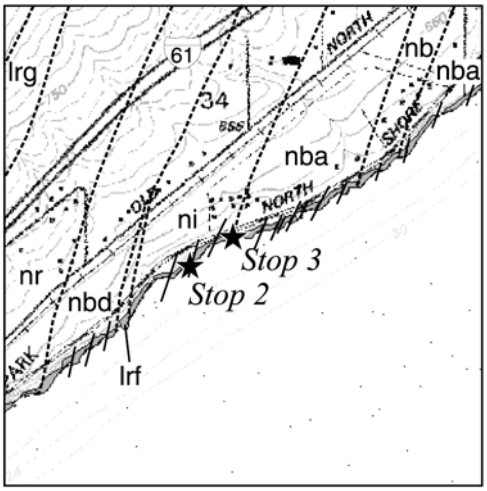
T.51N., R.13W., Sec. 34

Lakewood 7.51 quadrangle

UTM: 578287E; 5189652N

Highlights: Interflow sandstone above icelandite (ni), below basaltic andesite (nba)

Park at wayside across from North Shore Cottages



Area A: Slide down the embankment below wayside (over red clays of Glacial Lake Duluth) to outcrop of sandstone on shoreline. Sandstone is approximately 12-15 feet thick, planar and trough cross-bedded. Jirsa (1984) reports the sand grains are predominantly plagioclase (44-53%) and mafic volcanic rock fragments (18-25%), with lesser quartz (1-4), opaques (2-8%), K-spar, felsic volcanic rock fragments, and mudstone fragments (less than 2% each). The sandstone is cemented by a matrix of calcite, zeolite, K-feldspar, and quartz. Crossbedding indicates an easterly paleocurrent direction.

The sandstone overlies an icelandite flow which has a rubbly flow top (and unfortunately is quite inaccessible due to cliffs), and is overlain by a flow of basaltic andesite composition (will examine these at site B). The upper surface of this interflow sandstone contains curly molds and bits of baked-on chilled basalt from the overlying flow, which is now eroded away.

This sandstone is typical of the thin interflow sedimentary rocks, which are immature, fine- to coarse-grained sediments that were deposited in fluvial and lacustrine environments, although some may interpret some of the bedding in this outcrop to be aeolian in nature. Overall, paleocurrent indicators for the interflow sandstones indicate a transport direction generally toward Lake Superior from a highland of both Keweenawan and pre-Keweenawan sources (Jirsa, 1984).

---Climb back up bank and move about 100 feet northeast and return to lakeshore outcrops.---

Area B: Basaltic andesite. The base of this flow sequence was observed at stop 2-3a. These flows are dark purplish-gray, fine-grained basaltic andesite, which locally have pipe vesicles at the base, and amygdule-rich upper parts that exhibit billowy lobate forms defined by concentrated amygdules. Flow tops are commonly brecciated. Some flows contain sparse plagioclase phenocrysts. Due to time constraints we will not progress up the shore, but the return visitor is encouraged to walk up the shore over at least eight more flows of this composition that are in turn overlain by ophitic basalt flows. Brannon (1984) classifies these flows as ferroandesite, but they are morphologically more similar to flows of basaltic composition.

---- *Climb back up back to highway, back to cars. Continue northeast on the Scenic Highway approximately 8 miles to Alseth Road. Turn right on Alseth Road and follow around to a point where shoreline is close to road, park and walk to shore (Stony Point).*---

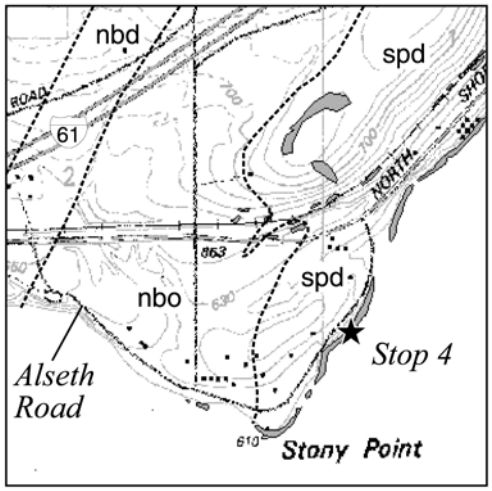
STOP 2-4: <u>Stony Point diabase</u>

T.51N., R.12W., Sec. 1

Knife River 7.5' quadrangle

UTM: 590326E; 5197791N

Highlights: Stony Point diabase (spd)



The diabase forms a sheet-like intrusion that strikes parallel to the shoreline and is discordant to the surrounding volcanic rocks. It is exposed in numerous outcrops between here and Knife River, a distance of nearly 3 miles. It is composed of coarsely ophitic olivine diabase with scattered thin pegmatitic segregations. The island offshore from Knife River forms the northeast-most outcrop of this diabase. The exposed thickness is estimated to be about 100 meters, but a positive aeromagnetic anomaly that corresponds to this diabase, and continues offshore beneath Lake Superior, implies that this diabase is part of a larger intrusive mass whose thickness is unknown. The diabase is chilled against adjacent volcanic rocks, and contains rare small anorthosite xenoliths (neither visible at this stop).

---*Continue northeast on the Scenic Highway, to Highway 61 into Two Harbors. Continue through Two Harbors to the northern outskirts of town, turn right on 1st Street and continue south to the crest of the hill adjacent to the picnic grounds. Walk east to lakeshore.*---

STOP 2-5: <u>Two Harbors Town Park</u>

T.52N., R.10W., Sec. 6

Two Harbors 7.5' quadrangle

UTM: 601851E, 5208673N to
601847E, 5208132N

Highlights: Basaltic flows of quartz tholeiite composition (ntb), rubbly flow tops, diabase dike (db)



This sequence of quartz tholeiite basalt flows is part of the Two Harbors lithostratigraphic unit (see table in introduction). These flows are comprised of brownish-gray to gray, fine-grained, fresh intergranular basalt that typically exhibits a pronounced oxidation-lamination type of flow banding. The upper parts of these flows typically contain a thick and irregular section of rubbly aa rich in laumontite, and red siltstone is common as thin lenses and crack fillings in the flow tops.

Walk south along the shoreline and observe the transition from massive, oxidation-laminated flow interiors to amygdaloidal, rubbly flow tops, and contacts with overlying flows. Each rocky point along this traverse corresponds to the massive, more resistant base of a flow that is underlain by a rubbly, easily eroded, flow top.

A N28W striking diabase dike that cuts the basalt can be seen along the south side of the north-most point, but due to time constraints this field trip will not go there. This dike follows a set of closely spaced brittle fractures in the basalt.

:---Continue northeast of Two Harbors on Highway 61 approximately 4.5 miles to the Silver Cliff Tunnel. Just past tunnel turn right into an informal parking area adjacent to highway.---

****** TRAFFIC HAZARD ****** USE EXTREME CAUTION ******

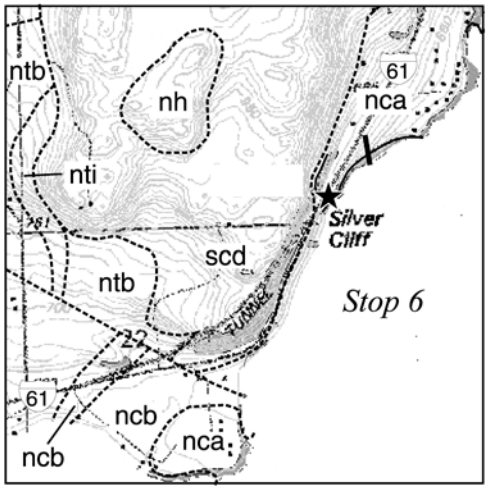
STOP 2-6: <u>Silver Cliff Tunnel</u>

T.53N., R.10W., Sec. 15 and 22

Castle Danger 7.5' quadrangle

UTM: 606993E, 5213952N (park area) to 606755E, 5213486N

Highlights: Silver Creek diabase (scd), contact with subjacent andesitic flows nca), fault



The Silver Creek diabase forms an irregular subcordant, subhorizontal intrusion that is at least 60-m (200-ft) thick. The diabase forms a prominent highland that projects inland several miles from Silver Cliff at Lake Superior. The Highway 61 tunnel has created excellent exposures of the contact between both the top and bottom of the diabase with adjacent volcanic rocks, and has exposed a north striking, 55-degree east dipping brittle fault that cuts the base of the diabase.

The andesitic flows beneath the sill show a very irregular, rather chaotic flow contact, a rubbly flow top, vuggy quartz-lined stretched vesicles and amygdules of gray agates that have been recrystallized by contact metamorphism from the adjacent diabase. The margins of the diabase are marked by a mix of commingled fine-grained, strongly magnetic, dark gray diabasic rock and pink granophyre.

The lower diabase contact strike is approximately N-S, 70 degrees west, and the upper contact (exposed at the south end of the tunnel) strikes N40E, 70 degrees northwest. Out on the old roadbed the diabase displays prominent columnar joints that plunge approximately 60-65 degrees east. From the north edge of the old roadbed, one can look back across the highway down the length of the brittle fault that cuts the base of the diabase. The fault is about 3 meters thick and filled with a mixture of pink zeolites and calcite-filled voids around altered diabase gouge.
The old roadbed affords an excellent view of the Lake Superior coast.

----*Continue northeast up Highway 61. The next tunnel is cut through the Lafayette Bluff diabase, a dark green, porphyritic, amygdaloidal diabase which bears distinct textural differences from the Silver Creek diabase, and that locally contains anorthosite inclusions up to 2 meters in diameter. As we drive past, note the thick cap of reddish-brown grus developed on top of the Lafayette Bluff diabase, which is typical of this unit. Just inland, the grus was formerly excavated for use as road surfacing material. We will see blocks of the Lafayette Bluff diabase at the parking area for the next stop.*---

---*Continue past Lafayette tunnel, pass Gooseberry State Park. From Gooseberry River, go 3.8 miles to a pulloff on lakeward side of road approximately 300 feet before crossing the Split Rock River. Park at wayside and walk northeast to small point to gain access to shore.*---

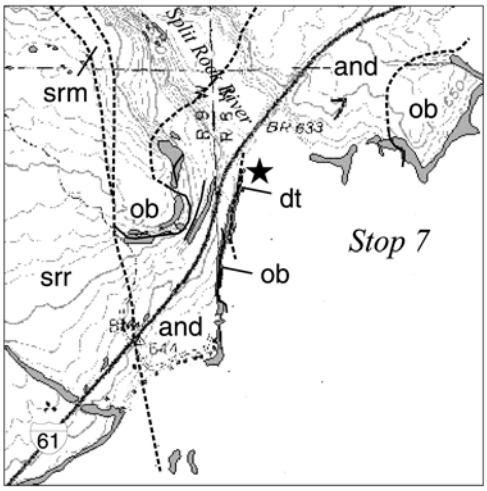
STOP 2-7: <u>Split Rock River diatreme</u>

T.54N., R.8W., Sec. 7

Split Rock Point 7.5' quadrangle

UTM: 620570E, 5226484N to
620497 E, 5226229N

Highlights: Split Rock Creek diatreme breccia (dt), brecciated basalt, basalt/andesite flow contact



Outcrops to north along shore are of weakly to moderately brecciated ophitic basalt that has apparently been slightly disrupted by the adjacent diatreme dike. A small "island" just offshore of the biggest point here consists of dark gray, medium-grained, ferromonzodiorite that petrographically consists of felty plagioclase, prismatic augite, skeletal to blocky oxides, and red-dusted anhedral-interstitial K-feldspar. The edge of this outcrop nearest the shore shows a dramatic brecciated texture due to diatreme emplacement, and the cliff adjacent to that exposes the diatreme breccia proper.

The best diatreme breccia is exposed further south along the beach. Here, the breccia contains visible clasts of massive ophitic basalt, fine-grained amygdaloidal basalt, and intergranular basalt that vary from sub-millimeter to 25 cm in diameter and are all strongly magnetic. Small fragments of interflow sedimentary rocks in the breccia have been noted in thin section. The zeolitic matrix varies from pink to pale green in color. At the south end of this breccia, one can observe the sharp, vertical contact between the diatreme and layered amygdaloidal basalt. The contact here strikes approximately N20W, but overall it is more or less north-south.

A short distance south of the diatreme contact are vertical cliffs formed by the blocky jointing pattern of a dark gray, very hard and fresh basaltic andesite flow which overlies amygdule-layered flows that exhibit minor low-amplitude billowy surfaces. The flow contact can be observed southward down the beach to the next point out into the lake. The top of this andesitic flow may be visible across the bay to the north, where it is overlain by massive ophitic basalt.

The diatreme exposed here is identical to another to the south, near Crow Creek, that is much more well exposed but unfortunately only accessible by water. There, the diatreme forms a vertical, 10 meter thick, cliff-forming dike that cuts ophitic basalt flows along the shoreline, and in some small outcrops inland. Although poorly constrained, the Crow Creek diatreme may be in the form of a ¼ mile diameter ring-like circular dike, or it may simply have an irregular, 'wandering' strike direction that varies from north-south to east-west. The Crow Creek diatreme is a heterolithic breccia that contains clasts of basalt up to 3 meters in length, and has irregular pyrite-rich pods within it. Like the diatreme here, the Crow Creek diatreme has prismatic ferromonzodiorite associated with it, which is itself brecciated, and contains varied types of basalt clasts as well as clasts of interflow sedimentary rocks.

The varied types of basalts coupled with interflow sedimentary rocks that make up the clasts in both places demonstrate that the diatreme has cut through a substantial thickness of volcanic strata. Although prismatic ferromonzodiorite is apparently associated with the diatreme in some fashion, the abundance of zeolitic matrix mixed with rock flour implies that the diatremes were formed by a gaseous, volatile-rich explosive mechanism. The volatile source is not known. However, the Lafayette Bluff diabase, which occurs in close proximity to the Crow Creek diatreme, contains abundant zeolitic (mainly laumontite) amygdules, and overall exhibits moderately strong deuteric alteration, which implies that the magma may have been gas-charged. Thus in a highly speculative scenario, degassing during cooling of the Lafayette Bluff diabase may have led to a buildup of volatiles that subsequently exploded through the overlying basalts.

Both the Crow Creek and Split Rock river diatreme form vertical dike-like bodies which brecciated clasts of ophitic and intergranular basalt, interflow sandstone and siltstone, fine-medium-grained ferrodioritic rocks, isolated unit quartz grains, and isolated blocky plagioclase grains. At this locality the clasts are generally less than 0.5 meter in diameter, but the diatreme near Crow Creek contains clasts of basalt as large as 4 meters in length. Medium-grained prismatic ferrodioritic intrusive rocks are exposed next to the diatreme dikes at both localities, and at both places the ferrodiorites are cut by and hence predate the diatreme dikes.

---Continue up Highway 61 to Split Rock Lighthouse State Park and History Center. Pull into park lot for the History center and proceed toward the lighthouse. To visit the lighthouse area (Area A) during summer tourist season, the history center charges a fee or requires Historical Society membership. A fee is not required to access the shoreline exposures at Area B.

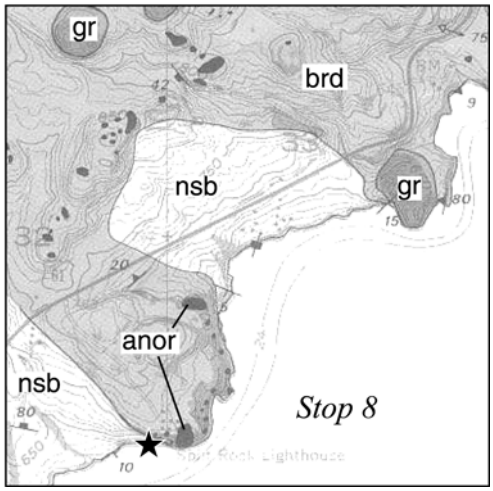
STOP 2-8: Split Rock Lighthouse Anorthosite

T.55N., R.8W., Secs. 32 and 33

Split Rock Point NE 7.5' quadrangle

UTM: Area A- 623690E, 5228678N
Area B - 623625E, 5228665N

Highlights: large anorthosite inclusions (anor), ophitic diabase (brd), lower chill, basaltic lava flow with silt-filled aa flow top (nsb)



The Beaver River diabase is the most areally extensive intrusive phase of the entire BBC and is found in contact with most other BBC units. In the southern BBC (Fig. 2-4), it occurs as a series of dikes, sills, and sheets of ophitic olivine gabbro that grades into coarser and more subophitic to intergranular gabbro in the medial portions of thicker sheets. One of the most distinctive characteristics of the diabase is that it commonly hosts large (as much as several hundred meters in diameter), rounded to angular inclusions of nearly pure anorthosite. These inclusions, which locally are brecciated and recrystallized (Morrison and others, 1983), are particularly common in the upper and lower margins of the larger diabase sheets. As reported for this stop in the Field Trip 7 guidebook (stop 24), these inclusions have been the focus of geologic debate for over 150 years.

Area A: Outcrops of fine-grained, ophitic olivine diabase with centimeter-wide augite oikocrysts, typical of the margins of Beaver River diabase, are exposed just northeast of the lighthouse atop a sheer 30-m-high sea cliff. The diabase here forms a sill that dips gently (<15°) into the lake and whose basal contact with a basalt flow top is exposed around the base of the point. The diabase is a moderately evolved (mg#=57), high-Al, olivine tholeiite (Fig. 2-5, BRD margin). The prominent point just to the northeast (Rusty Point) is held up by a very large (~200 m) inclusion of medium-grained leucogranite lying at the base of the sill. The diabase is slightly chilled around this granite inclusion, which closely resembles the Finland granophyre (see Stop 2-12).

Most of Split Rock Lighthouse Point is held up by a single, large (>35m), layered anorthosite inclusion. Around the base of the lighthouse, the inclusion displays meter-scale modal layering of coarse-grained granular (cataclastic?) anorthosite (>99% Pl) and noritic anorthosite (20% Opx [$En_{70}Fs_{28}Wo_2$], 80% Pl [An_{60-80}]; Morrison and others, 1983). The steeply dipping layers are cut by thin dikes of medium-grained, granular augite leuconorite (An_{56}, $En_{73}Fs_{24}Wo_3$, $En_{46}Fs_{12}Wo_{41}$, Morrison and others, 1983).

.....Head down the tramway path to the lakeshore and carefully make way along base of slope toward lighthouse. PROCEED CAREFULLY OVER LARGE TALUS BOULDERS.

Area B: To the southwest of the lighthouse along the shore, the base of the sill can be observed to conformably overlie an amygdaloidal flow top breccia with a matrix of horizontally bedded siltstone. The diabase here contains a variety of types and sizes of anorthosite inclusions, which stand out as unjointed masses within the highly jointed fine-grained diabase. The diabase is strongly chilled at the basal contact against the flow top, but shows no sign of chilling against the anorthosite inclusions. The variety of anorthosite types can be viewed in the talus blocks at the base of the slope. Looking to the northeast, one can see that the near-vertically, layered inclusion beneath the lighthouse extends to lake-level.

This and other anorthosite inclusions hosted in the Beaver River diabase are different from most anorthositic series rocks of the Duluth Complex; the latter are rarely layered, more compositionally evolved, and rarely contain cumulus hypersthene. These inclusions are similar, however, to some anorthosite inclusions within the anorthositic series. The highly disordered structural state of plagioclase (Miller, unpublished data) and the absence of any discernable chill of the diabase against the anorthosite indicate that these inclusions were derived from a middle to lower crustal source. Isotopic and trace-element compositions of these crustal xenoliths suggest that they may be pre-Keweenawan in age (Morrison and others, 1983), but the data are overall ambiguous. Alternatively, if a plagioclase crystal mush origin for Duluth Complex anorthositic rocks is correct (Miller and Weiblen, 1990), a corollary of such a model is that significant amounts of Keweenawan anorthosite, generated by plagioclase flotation under high pressure, should have formed in the deep crust prior to BBC magmatism at 1096 Ma. Under deep crustal conditions, such plagioclase cumulates would probably be distinctive in texture and composition from their shallow crustal counterparts. Moreover, the ambiguous isotopic compositions of the inclusions may indicate that anorthosite-forming Keweenawan magmas were contaminated by older crust, rather than older anorthosite being contaminated by interaction with Keweewawan magmas, as concluded by Morrison and others (1983).

Return to Duluth for the evening.

DAY 2

Drive Highway 61 from Duluth past Two Harbors to Lake Co Rd 3. Take this road approximately 25 miles. About 2.3 miles past the Silver Bay municipal airport, look for a prominent roadcut on the adjacent railroad grade on the north side of the road. Pull into gravel pullout.

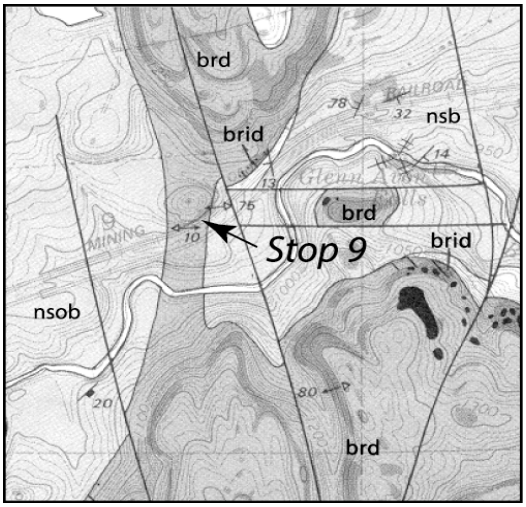
STOP 2-9: Fanning columnar joints in Beaver River Diabase dike/sill

T.55N., R.8W., Sec. 9

Silver Bay 7.5' quadrangle

UTM: 624590E, 5235865N

Highlights: columnar jointed ophitic diabase (brd), anorthosite inclusion (dark outcrops), contaminated and inclusion-rich diabase (brid)



The western and northern extent of the Beaver River diabase is marked by a 110-km-long dike (or dike set) across which vertical downward displacement of 1.5 to as much as 6 km on the southeastern (rift axis) side is indicated by offset of older geologic units (Miller and Chandler, 1997). The feature, which has a very pronounced aeromagnetic expression (Fig. 2-7), is termed the Finland tectono-magmatic discontinuity (FTMD). Within the concavity of the FTMD are several other large, often bifurcating dikes and large sheets dipping gently southeast. The complex of sheets holding up tabletop highlands in the southern BBC (Figs. 2-4 and 2-7) may be part of an originally continuous, nearly horizontal sheet that has locally been eroded through to expose volcanic rocks forming the footwall.

The fanning pattern of columnar jointing exposed in this railroad cut through Bear Lake Ridge reflects the transition from the FTMD dike into a thick sub-horizontal sheet developed to the east of the ridge. The rock type here is a fine-grained ophitic olivine diabase. The larger railroad cut 150 m to the east reveals the irregular, well-jointed base of the diabase sheet overlying the amygdaloidal flow-top breccia of a deeply weathered ophitic basalt (CAUTION, OUTCROP FACE IS UNSTABLE). Abundant weathered anorthosite inclusions and some quartz- and feldspar-phyric rhyolite inclusions (similar to the Palisade rhyolite) are evident in the diabase. This diabase sheet holds up the extensive 120-m cliff visible across (south of) the Beaver River.

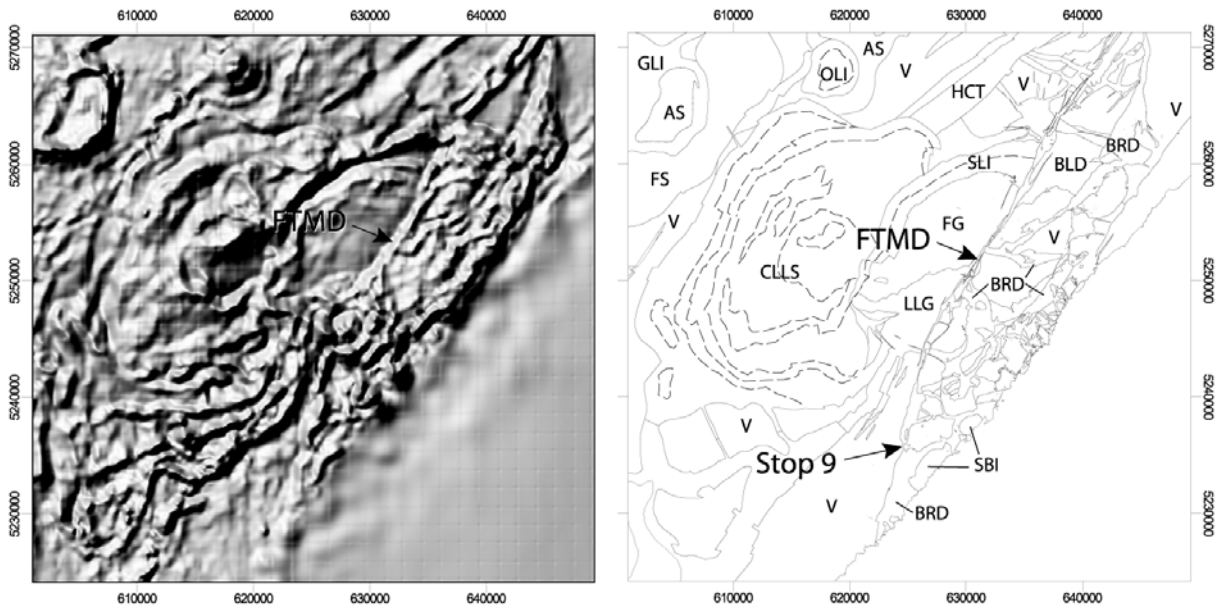


Figure 2-7. Shaded-relief image of total magnetic field data over the southern Beaver Bay Complex and generalized geologic map (from M-119) at the same scale. FTMD - Finland Tectono-magmatic Discontinuity. Labelled intrusions of the BBC include the Cloquet Lake layered series (CLLS), the Houghtaling Creek troctolite (HCT), the Sonju Lake intrusion (SLI), the Finland granophyre (FG), the Lax Lake gabbro (LLG), the Blesner Lake diorite (BLD), the Beaver River diabase (BRD), and the Silver Bay intrusions (SBI). Duluth complex units include the anorthositic series (AS), the felsic series (FS), the Osier Lake intrusion (OLI), and the Greenwood Lake intrusion (GLI). North Shore Volcanic Group rocks are denoted as V. Solid lines denote inter-intrusion contacts; dashed lines denote intra-intrusion contacts. See M-119 for subunits of intrusions. Number labels are NAD 83-based UTM grid coordinates. Modified from Figure 7-2 (Miller and others, 2002).

Two types of diabase are exposed in the upper ledge of the railroad cut, which is accessible from the east. The more common type is the same fine-grained ophitic olivine diabase observed in the dike, but here it contains scattered anorthosite inclusions and locally is plagioclase porphyritic. Exposed across the eastern half of the ledge, the ophitic diabase is found in sharp contact with a dense, black, aphanitic, intersertal diabase rich in quartz and feldspar xenocrysts, numerous small, blebby inclusions of felsite, and larger inclusions of anorthosite. The presence of anorthosite inclusions argues against it being a volcanic xenolith. Near the western end of the ledge, the ophitic diabase is noticeably chilled and columnar jointed adjacent to a sharp steep contact with the inclusion-rich diabase. In the center of the ledge, the inclusion-rich diabase is itself pseudo-columnar jointed near its sharp contact with the ophitic diabase, but the joints are not orthogonal to the contact. This inclusion-rich diabase is locally observed near the margins of Beaver River diabase intrusions throughout the BBC. It probably represents strong contamination of early anorthosite-bearing Beaver River magmas as they were emplaced into brittle conduits in the volcanic edifice.

.... Continue east on Lake Co Rd 3 to junction with Co Rd 4. Turn right and proceed to junction with Highway 61 in Beaver Bay. Turn left, after crossing the Beaver River pull on to shoulder. Walk paths down to beach.

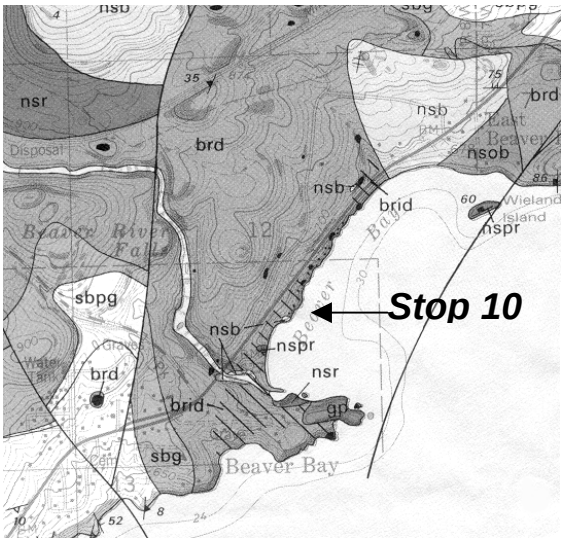
STOP 2-10: <u>Inclusion-rich Contact Zone of the Beaver River Diabase at Beaver Bay</u>

T.55N., R.8W., Sec. 12

Silver Bay 7.5' quadrangle

UTM: 629115E, 5235420N to 629240E, 5235770N

Highlights: porphyritic rhyolite inclusion (nspr), hornfels basalt inclusions (nsb) with granophyre ring dikes, anorthosite inclusions (black masses), inclusion-rich Beaver River diabase (brid)



Exposed along shoreline outcrops on the north side of the Beaver River is a complex mix of rock types that is commonly characteristic of the basal? contact of the Beaver River diabase. Exposed in the first outcrop north of the Beaver River sandbar is hybrid mix of fine-grained diabase and irregular masses of felsic rock in various stages of mixing and assimilation. While some of the felsic material looks to be rounded irregular melted? inclusions in the diabase, some felsic material clearly occurs as dikes in the diabase. At the steep north end of this exposure is a quartz-feldspar porphyritic rhyolite inclusion that is very similar to the Palisade rhyolite, which holds up Palisade Head and Shovel Point several miles to the north of here. At the base of the steep north side of the outcrop, the rhyolite is in sharp contact with a fine-grained diabase. The diabase is riddled with irregular blebs of felsic material that likely represent melted fragments of the rhyolite inclusion. Wieland Island in the northern part of Beaver Bay is composed of a similar porphyritic rhyolite. The cliffs across the mouth of the river to the southeast are held up by large inclusions of aphyric flow-banded rhyolite and medium-grained massive granophyre.

Progressing to the northeast, we cross a small outcrop of an epidotized fragmental mafic rock. This rock may be contaminated diabase or, as implied by similar appearing rock we shall see to the north, it is more likely a metamorphosed flow top breccia of a basalt inclusion. The next low outcrop is of deeply altered (green-tinted) diabase charged with similarly altered anorthosite inclusions. Coming on to a large expanse of semi-continuous outcrop along the rest of the shore, several rock types are observed. Most of the exposure is altered fine-grained diabase. Another prominent rock type is granophyre, which occurs both as irregular masses of varied size and as a curving network of granophyre dikes ranging from several centimeters to over a meter wide. In some areas, granophyre composes almost half of the exposure. In several locations, granophyre dikes are found encircling blocks of dense basaltic hornfels. The basaltic hornfels are sometimes only distinguishable from the fine-grained diabase by a very prominent orthogonal joint pattern and the ring of granophyre. In one block occupying a low spot in the outcrop, a flow contact between massive and flow breccia basalt is recognizable. These blocks of basalt flows were likely thermally metamorphosed by their incorporation into the diabase. The origin of the granophyre is less clear, however. Melting of rhyolite inclusions has evidently given rise to at least some of the granophyre, especially that occurring as irregular masses. However, given the mantling relationship of some granophyre ring dikes to the basaltic hornfels inclusions, is seems possible that some of the granophyre dikes were "sweated out" from the originally hydrated basalt inclusion. Alternatively, the

ring dikes may represent rhyolite-derived felsic melts that simply intruded along contacts between the diabase and the basalt inclusions.

Following the exposure further to the north, this mix of diabase, granophyre, and basalt abruptly gives way to diabase heavily charged with large anorthosite inclusions. At this point, we will head upslope to the highway and back toward the river.

.... Continue along Highway 61 to the northeast past the North Shore taconite plant at Silver Bay and Tettegouche State Park. Proceed 3.0 miles past the junction with MN Highway 1 to a road cut across from (NW of) a Jehovah's Witness Hall.

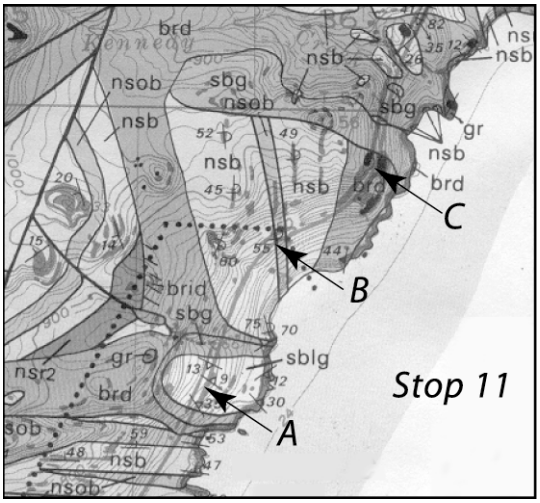
STOP 2-11: Silver Bay intrusions, Beaver River diabase and overturned lava flows

T.56N., R.7W., Sec. 1

Illgen City 7.5' quadrangle

UTM: A: 637890E, 5247155N
B: 638150E, 5247660N
C: 638421E, 5247980N

Highlights: ferrogabbro layered intrusion (sblg), overturned lava flows (nsb), monzodioritic composite intrusions (sbg) into diabase (brd) containing anorthosite and basalt inclusions



The Illgen City quadrangle contains some of the most complex geology exposed in the Keweenawan system (Miller and others, 1988). This complexity resulted from block faulting, displacement, and rotation of volcanic rocks caused by the dike and sill emplacement of the Beaver River diabase and later composite intrusions. The present level of exposure along the shoreline has cut into the footwall of a major sheet-like intrusion of diabase that holds up the highlands about 2 km in from the shore. This is perhaps the same sheet that the Beaver River cuts through near Stop 2-9. The footwall is highly dissected with a complex network of large (50-800m wide) diabase dikes and the intervening blocks of volcanic rocks have been significantly displaced and rotated into a variety of orientations (Fig. 2-8). Multiple composite intrusions of ferrogabbroic to intermediate magmas have produced a variety of intrusions hosted by the Beaver River diabase, which are collectively termed the Silver Bay intrusions. Three roadcuts along a one-and-a-half kilometer stretch of the northwest side of Highway 61 display some of the best examples of the magmatic and structural complexities attending the emplacement of the Beaver Bay Complex.

Roadcut A: This roadcut exposes a nearly complete cross section through a small, zoned Silver Bay intrusion. This intrusion, termed the Jehovah body, has the form of a broad, shallow asymmetric synform that plunges about 15° to the east and measures about 450 m north to south. It is intrusive into the axial portion of a Beaver River diabase dike about 500-600 m wide, which dips steeply to the north.

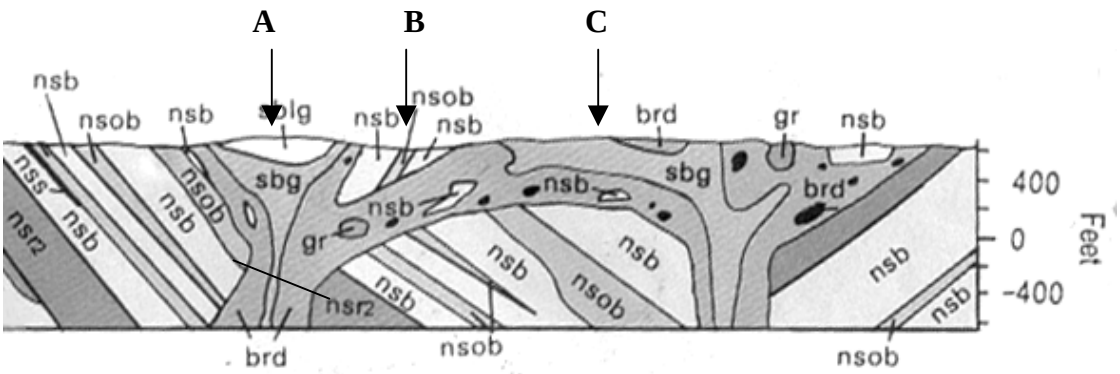


Figure 2-8. Geologic cross-section of the shoreline in the vicinity of STOP 2-11 (From Miller and others, 1989). Unit abbreviations are: nsb- basalt; nsob-ophitic basalt; nsr2-rhyolite; nss-interflow sandstone; brd-Beaver River diabase; sbg-Silver Bay gabbro; sblg-Silver Bay laminated gabbro;, gr-granite inclusion. Approximate locations of roadcuts A, B, and C projected to the shore are shown. No vertical exaggeration.

On either end of the roadcut west of Highway 61 (Fig. 2-9A) are exposures of a deeply weathered, varitextured but generally coarse-grained, nonfoliated apatitic ferromonzodiorite forming the margins of the intrusion (unit **sbg**). Toward the center of the roadcut, the ferromonzodiorite abruptly grades into a medium-grained, well-foliated, subprismatic ferrodiorite with poikilitic olivine (unit **sbpg**). The poikilitic olivine ferrodiorite is composed of about 50-55% prismatic to lath-shaped plagioclase (An_{45-15}); 15-20% subprismatic and partially uralitized augite (En'_{62-42}) and minor intergrowths of pigeonite (En'_{42-36}); 5-10% subequant to bladed iron-titanium oxide; 5-15% subpoikilitic to poikilitic, altered olivine ($<Fo_{48}$); and 5-15% granophryic mesostasis and trace apatite. Olivine is almost completely altered to a glassy black sheet silicate mineral, called hisingerite by Gehman (1957). In thin section, the hisingerite appears to be mostly dark-greenish-brown biotite, partially altered to chlorite, with some serpentine and iron oxide. The oxidation of iron oxide, which is especially common around the outer margin of the olivine oikocrysts, produces purplish coronas around the black clots.

Variations in the abundance and size of olivine oikocrysts (1.5 to 5 cm across) impart a layering to the gabbro, which is parallel to plagioclase foliation. This internal structure defines an asymmetric synform whose axial plane projects through the northern half of the intrusion. Cryptic variations in the Mg/Mg+Fe contents of mafic minerals are noted upward through this sequence (Fig. 2-9A) and are suggestive of magmatic differentiation.

We interpret the coarse-grained, decussate rocks to represent equilibrium crystallization of a stagnant margin around a small convecting core of fractionally crystallizing magma, which formed the foliated gabbro. The dikes of coarse-grained nonfoliated ferrodiorite internal to the zoned intrusion may represent late intrusions of parental magma or remobilization of interstitial magma from within the intrusion.

Roadcut B: The next major roadcut to the northeast displays a series of basalt lava flow that have been overturned 125° from horizontal. These flows are part of a large (~1 x 2 km) block of basalts that is completely enclosed by intrusive rocks of the Beaver River diabase and Silver Bay intrusions (Fig. 2-8 and location map).

The basalts range in composition from olivine tholeiites, with smooth upper flow surfaces and pipe amygdules at their base, to quartz tholeiites with brecciated (aa) flow tops (Fig. 2-9B). They range in thickness from 3 to 20 meters. Two interflow sandstone units occur beneath the thickest lava flows. These flows are part of the Bell Harbor lavas, the uppermost unit of the upper southwestern sequence of the NSVG (Table 2.1).

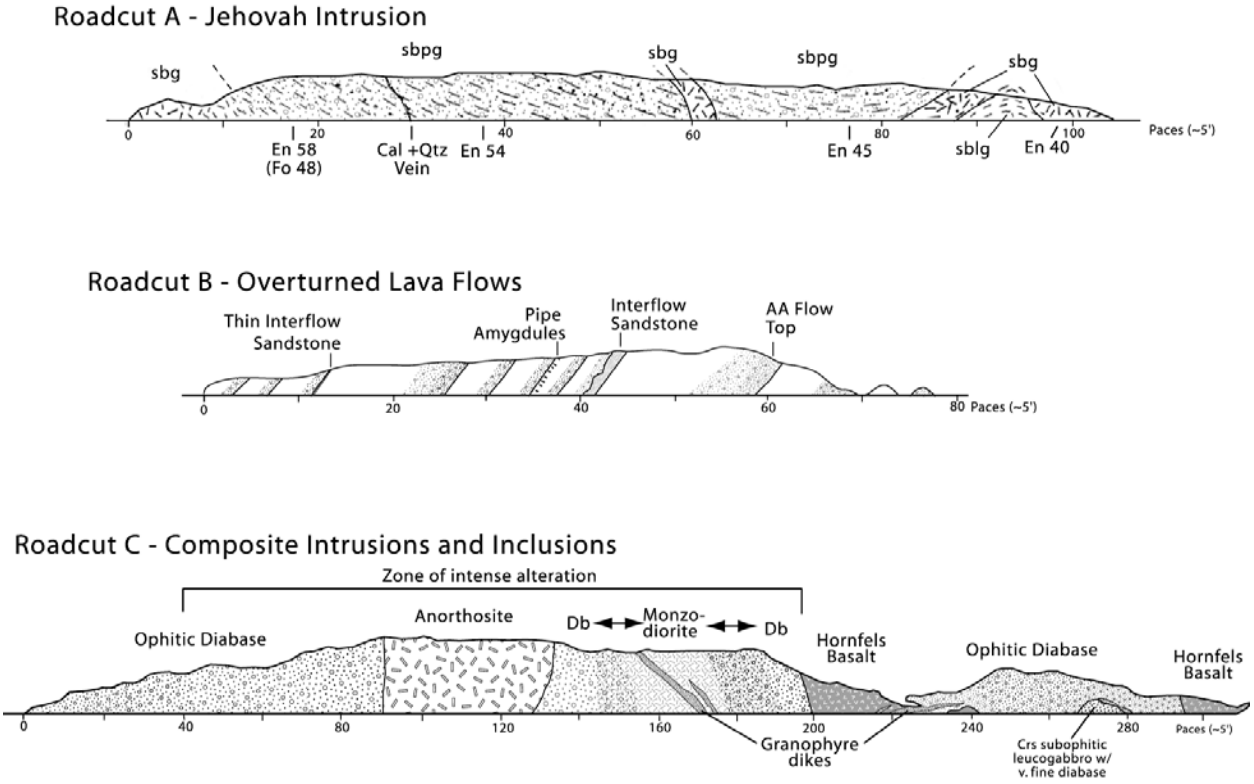


Figure 2-9. Schematic diagrams of geologic relationships exposed along three roadcuts at STOP 2-11. Foliation development portrayed in Roadcut A by alignment of tick marks. Units: sbg- nonfoliated ferromonzodiorite; sbpg - foliated poikilitic olivine ferrodiorite; sblg - foliated intergranular (non-poikilitic) olivine ferrodiorite. Compositions of augite shown by En content (= MgO/(MgO+FeO), mole %).

Roadcut C: A complex sequence of intrusive and inclusive relationships are exposed in the 70' high roadcut to the northeast (Fig. 2-9C). This is one of the tallest roadcuts on the North Shore. Starting at the southernmost end of the roadcut (0 pace) is a medium-grained ophitic olivine diabase with 2cm augite oikocrysts. The diabase gradually coarsens to where by 80 paces the oikocrysts are about 5 cm across. Beginning at about 40 paces, the diabase displays an intense oxidation/alteration that persists through about 200 paces, where it is in contact with hornfels basalt. Between 90 and 130 paces is a coarse - grained anorthosite inclusion, which has near vertical contacts with the enclosing diabase and displays the same alteration as the diabase. This anorthosite holds up the crest of the hill that the roadcut traverses. On the north side of the inclusion, the diabase is more subophitic, with coarse (~2 cm) augite clots, and it contains granophyric patches. Progressing north, augite gradually becomes more prismatic and granophyre content increases to where by 160 paces the rock is a ferromonzodiorite typical of the marginal phases of Silver Bay intrusions. Between 170 and 175 paces, two steeply inclined granophyre dikes cut the ferromonzodiorite. At about 180 paces, a subophitic texture again is evident. This rock gradually becomes more ophitic and more medium-grained as a contact with hornfels basalt is approached at 200 paces. The northern contact of this basalt inclusion occurs at a low spot in the roadcut and is obscured by two shallow dipping granophyre dikes. The larger dike (~1.5m wide) cuts medium fine-grained ophitic diabase, which contains another hornfels basalt inclusion near the base of the exposure at about 240 paces. The exposure from 240 to 295 paces is relatively unaltered, medium fine to medium-grained diabase. Between 270 to 280 paces, near the base of the roadcut, is an odd inclusion of coarse-

grained, subophitic leucogabbro with apophyses? of very fine, amygdaloidal mafic rock. The inclusion is mantled on its upper right margin by a thin rind of granophyre. The northernmost 10 paces of the roadcut leading to a gravel side road is hornfels basalt.

The sequence of events interpreted from the rocks displayed in Roadcut C are:

1) intrusion of diabase dike carrying anorthosite inclusions from depth and incorporating locally derived basaltic wall rock.
2) composite intrusion of ferromonzodioritic magma into the semi-molten core of the diabase dike;
3) injection of granophyre dikes (possibly auto-intrusions of late-stage felsic melt segregated from the ferromonzodiorite); and
4) fluxing of oxidizing fluids through the central zone of the dike.

--- Continue on Hwy. 61 to Lake Co. Hwy 6. Take this to MN Hwy 1 in Finland. Turn right on Hwy 1 about 1/4 mile and turn right on to Lake Co Hwy 7. Just past the Finland Community Center and just before the pavement ends, turn left on to the road to the site of a former USAF Radar Station. Take the paved road about 1.4 miles to gravel road crossing and pull out to the left.---

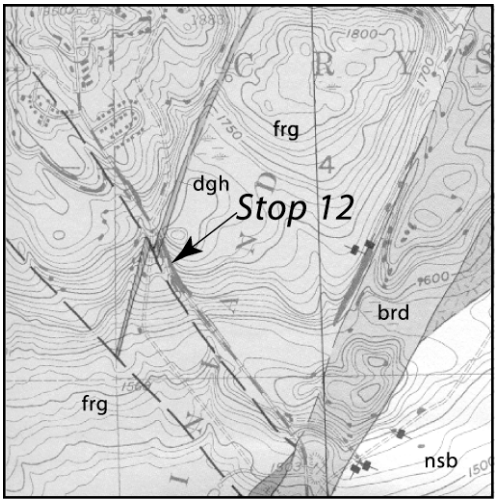
STOP 2-12: <u>Finland granophyre and hybrid dikes</u>

T.57N., R.7W., Sec. 4

Finland 7.5' quadrangle

UTM: 633180E, 5256520N

Highlights: granophyric leucogranite (frg), locally with blebby mafic enclaves (dgh)



Exposed over a 40-km^2 area in the northwestern part of the southern BBC (Fig. 2-4) is a distinctive, oval-shaped mass of leucogranite and quartz ferromonzodiorite, termed the Finland Radar Station granophyre by Miller and others (1993), or simply the Finland granophyre. The main phase of the granophyre is a homogeneous, salmon-colored, micrographic leucogranite with abundant miarolitic cavities (unit **frg**, Fig. 2-4). To the north, this granitic phase, which contains less than 5% Fe-silicates and oxides, abruptly grades to a quartz ferromonzodiorite characterized by 5-20% prismatic, iron-rich pyroxene, amphibole, and locally olivine and 20-40% micrographic felsic matrix (unit **frqm**, Fig. 2-10). At this stop, the typical character of the granitic phase of the Finland granophyre is displayed in the roadcuts on either side of the road to the former USAF radar station (decommissioned in 1974).

In the rubble at the north end of this roadcut, the medium-grained granophyre contains small rounded enclaves of fine-grained mafic rock. Some of the rubble blocks are composed completely of this fine-grained rock. This occurrence is in line with what to the north is clearly a mafic to intermediate dike cutting the granophyre. This dike is part of an orthogonal network of dikes composed of heterogenous mixtures of mafic and felsic rock types that cut the Finland granite and the underlying Sonju Lake

intrusion and that parallel the adjacent Beaver River diabase (BRD-SLI hybrid dikes in Fig. 2-10A). Miller and Chandler (1997) have interpreted these dikes to represent hybrid mixtures of Beaver River diabase magma and residual felsic melts in the upper Sonju Lake intrusion and overlying Finland granophyre. The textures displayed here are consistent with those resulting from two magma mixing, which would imply that the Finland granite was significantly molten when the mafic dikes were emplaced.

The intrusive relationships of the Finland granophyre, the older Lax Lake gabbro along its southern margin (**llg**, Fig. 2-4), and the younger Beaver River diabase at its eastern extent (unit **brd**, Figs. 2-4 & 2-10A) are well-established by several exposures of sharp contacts. However the genetic relationship of the granophyre to the underlying Sonju Lake layered intrusion (Stop 13) along its gradational northern boundary (Fig. 2-10A) is more problematic. Stevenson (1974), noting aplitic granophyric dikes cutting the mafic cumulates of the Sonju Lake intrusion, considered the Finland granophyre to be intrusive into the Sonju Lake intrusion. However, mapping by Miller and others (1993) showed these aplite dikes to be phases of the late hybrid dikes (Fig. 2-10A) that cut the Sonju Lake intrusion and both phases of the Finland granophyre. Again, the termination of such a dike is thought to give rise to the mafic enclaves at this stop. Given the parallel zonation of the two phases of the granophyre with the strike of Sonju Lake cumulate units (Figs. 2-4 & 2-10; also see Miller and others, 1993a) and the generally smooth compositional variations across these units (Fig. 2-10B), a comagmatic relationship resulting from crystallization differentiation could be envisioned. Such an interpretation is inconsistent, however, with the large amount of granophyre relative to the layered mafic rocks apparent from the geologic map (Figs. 2-4 & 2-10). Modeling of gravity and aeromagnetic data across these units (Miller and others, 1990) confirms that the volume of granophyre is at least as great as the volume of underlying mafic rocks and thus is too great to be a differentiate of the layered intrusion. Also, preliminary Sm-Nd and Rb-Sr data indicate that the granophyre has an isotopic composition distinct from the mafic rocks and therefore could not have evolved from them (J. Vervoort, unpublished data). A more plausible interpretation is that the mafic magma that produced the Sonju Lake intrusion was trapped by the less dense Finland granophyre. This underplating resulted in partial melting of the granophyre and the development of a diffuse contact. Petrologic and isotopic studies are needed to determine the extent of assimilation across the mafic-felsic contact.

--- Return downhill to Lake Co Rd 7. Turn left and proceed 5.6 miles to Sonju Lake Forest Rd. Take the Sonju Rd 3.1 miles to an area of low outcrops---

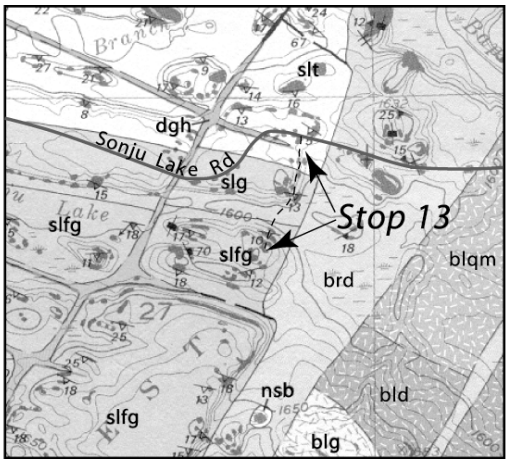
STOP 2-13: <u>Sonju Lake Intrusion</u>

T.58N., R.7W., Sec. 27

Finland 7.5' quadrangle

UTM: 633180E, 5256520N

Highlights: cumulate stratigraphy of Pl+Ol (slt) -> Pl+Cpx+Ol (slg) -> Pl+Cpx+Ox+/-Ol (slfg)



The Sonju Lake intrusion (SLI) is the most completely differentiated mafic layered intrusion recognized in the Keweenawan magmatic system (Stevenson, 1974; Weiblen, 1982). The intrusion occurs as a shallow (15-30°) south- to southeast-dipping, 1200-m-thick sheet of mafic cumulates underlying a granophyre body of similar thickness. The SLI is exposed in outcrop over an area extending about 3 km west from its abrupt truncation by the Beaver River diabase dike of the FTMD structure (Fig. 2-10A). However, it can be traced beneath glacial cover by its aeromagnetic signature for a strike length of about 15 km (Fig. 2-4). Above a basal contact zone of fine- to medium-grained melatroctolite (unit **slmt**) intrusive into older gabbroic rock, five map units within the SLI are distinguished by their cumulus mineralogy: Ol (unit **sld**)→ Pl+Ol (unit **slt**)→Pl+Cpx+Ol (unit **slg**)→Pl+Cpx+Ox±Ol (unit **slfg**)→Pl+Cpx+Ox+Ol+Ap(unit **slad**). At the upper contact with the overlying Finland granophyre is a complex unit (**slmd**) characterized by a cyclical loss of igneous lamination and enrichment in granophyre at the expense of mafic phases. This igneous stratigraphy is complimented by a smooth, logarithmic cryptic layering exemplified by decreasing mg# values of pyroxene (Fig. 2-10B). Taken together, these characteristics are consistent with formation of the SLI by closed-system fractional crystallization of a moderately evolved tholeiitic basaltic magma (Miller and Ripley, 1996).

At this stop, we will traverse the middle cumulate units of the SLI. Exposed in low outcrops on either side of the Sonju Lake Forest Road is coarse-grained, moderately foliated, ophitic olivine gabbro typical of the upper part of the **slt** unit. It is a Pl+Ol cumulate with abundant (15-20%) augite oikocrysts up to 8 cm in diameter. Downsection from here, the **slt** unit is more typically an ophitic augite troctolite with olivine about twice as abundant as augite and the latter as 1- to 3-cm oikocrysts (Fig. 2-11). The increased mode and size of augite oikocrysts here presages the approaching cumulus arrival of augite.

The arrival of cumulus augite is indicated by its abrupt change from ophitic to granular habit (Fig. 2-11), which can be observed in intermittent outcrop about 70 paces (~100 m) south of the road. This singular cumulus phase transition from Pl+Ol to Pl+Cpx+Ol is traceable across the entire exposure area of the SLI and serves as a useful datum horizon (Fig. 2-10). The **slg** unit is about 70 m thick and is typically composed of 60-65% plagioclase, 28-32% subprismatic augite, 5-10% olivine and 1% interstitial Fe-Ti oxide. The rock is consistently coarse-grained and well foliated.

Continuing south across a low swamp, the outcrop exposed on the north face and crest of the next ridgeline is medium-grained, well foliated oxide gabbro of the **slfg** unit. The cumulus arrival of Fe-Ti oxide is defined by its granular habit and its increase in modal abundance to 7-10% (Fig. 2-11). Olivine is commonly in low abundance or absent. This unit is over 200m thick, and in its medial section hosts an 80m-thick interval enriched in PGE (Miller, 1999; Joslin, 2004).

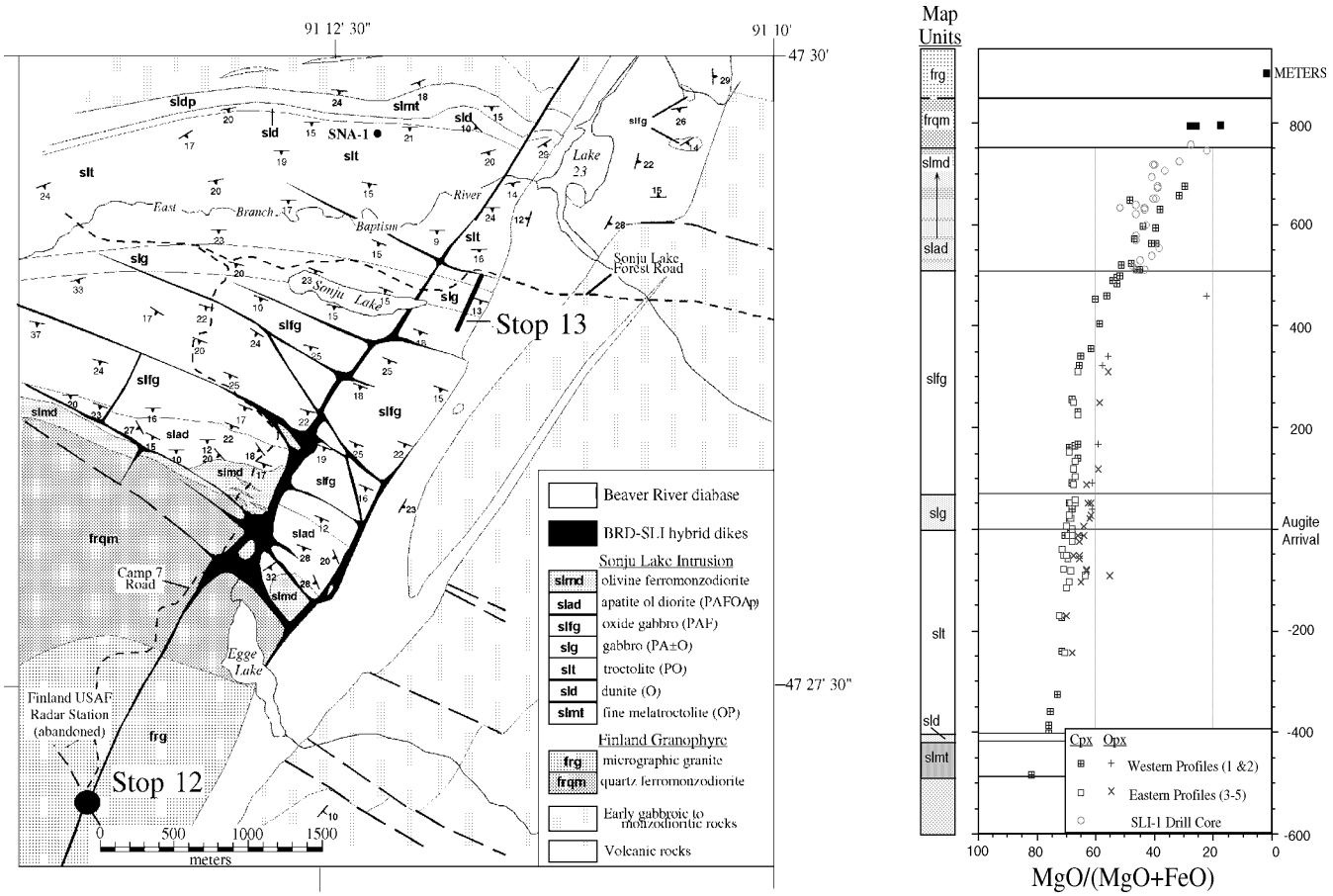


Figure 2-10. A) Geology of the Sonju Lake intrusion and overlying Finland granophyre showing the locations of stops 12 and 13 (modified from Miller and others, 1993). B) Cryptic variation of MgO/(MgO+FeO) (mole %) in augite and orthopyroxene through the Sonju Lake intrusion and Finland granophyre (from Miller and Ripley, 1997).

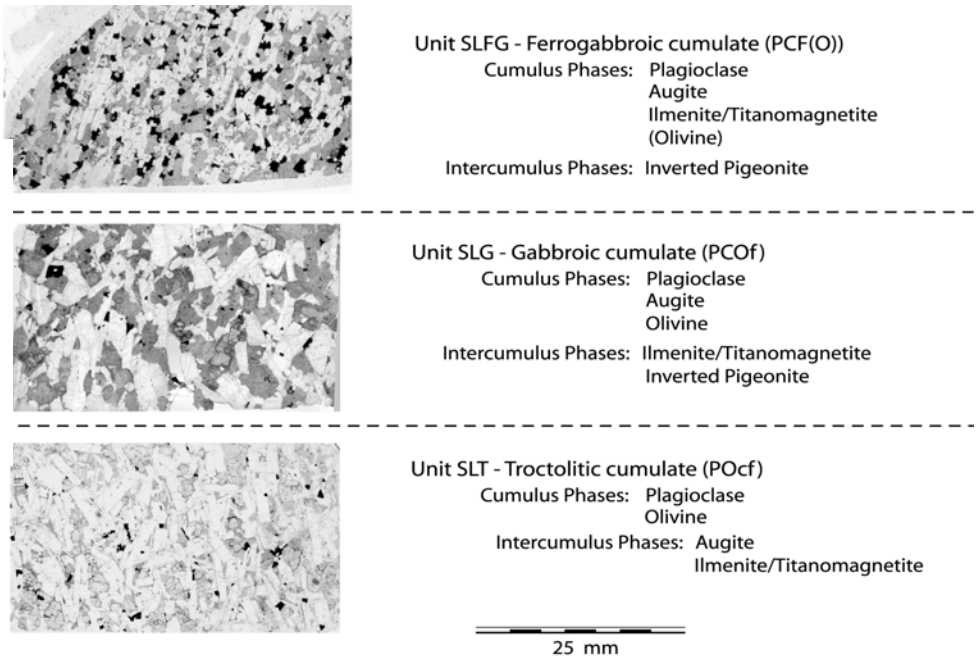


Figure 2-11. Textures of cumulate rocks from the middle units of the Sonju Lake intrusion. Light phase is plagioclase, outlined light gray phase is olivine, dark gray phase is augite, and black phase is Fe-Ti oxide.

--- Return to Co Rd 7. Proceed north (to left) about 10 miles to Hoist Lake Rd (just past the Trestle Inn on Crooked Lake). Turn left and go 4.5 miles to junction with snowmobile trail heading to right---

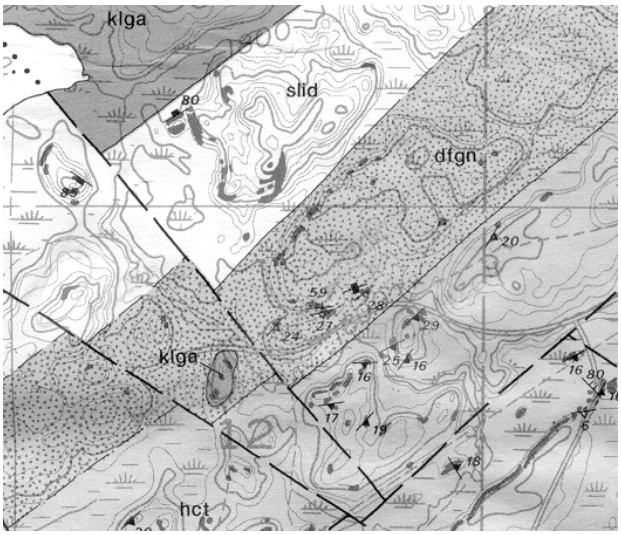
STOP 2-14: <u>Northwestern margin of the Beaver Bay Complex</u>

T.59N., R.7W., Sec. 12

Cabin Lake 7.5' quadrangle

UTM: A - 638780E, 5274975N
B - 638460E, 5274980N
C - 638125E, 5274755N
D - 638565E, 5275005N

Highlights: Houghtaling Creek trocolite (hct); Dam Five gabbronorite (dfgn) with inclusion of gabbroic anorthosite (klga); Shoepack Lake inclusion-rich diorite (slid)



This sequence of four exposure areas shows the major units comprising the northwestern margin of the Beaver Bay Complex (Miller and others, 1994). The first exposure (Area A) is of a medium coarse-grained, moderately foliated, layered troctolite cumulate with minor (<5%) ophitic augite forming the northwestern margin of the Houghtaling Creek troctolite (Figs. 2-4 and 2-12). The layering and foliation dip about 30º SE. This intrusion occurs as a 2-5 km wide, northeast-trending macrodike, which can be traced along strike for over 50 kilometer in outcrop and by its distinctive aeromagnetic signature (Fig. 2-7). Intermittent olivine layering and foliation define an asymmetric trough structure with its shorter limb on the southeast side (Fig. 2-12B).

---Following the snowmobile trail north of the road for about 100 m to a junction of trails. Take the trail to the left (west)---

Pavement outcrops on the snowmobile trail and in a larger outcrop on the north side of the trail (Area B) are medium-grained, well-foliated gabbronorite cumulates of the Dam Five gabbronorite (dfgn, Fig. 2-12). The rock is composed of 50% plagioclase, 20% subprismatic augite, 27% subprismatic inverted pigeonite, and 3% subpoikilitic Fe-Ti oxide and would classify as a Pl+Cpx+IPig cumulate. Well-developed foliation dips about 27º SSE. This unit occurs exclusively along the northwestern margin of the Houghtaling Creek troctolite. Although a contact between the troctolite and gabbronorite is not observed and their internal structures are nearly conformable, the troctolite does show some fining toward the inferred contact and is considered to be intrusive into the gabbronorite. The very different cumulus mineralogies of the two units, which cannot be related by any conventional differentiation mechanism, is further evidence that these are distinct intrusive units (Fig. 2-12B).

--- Continue along snowmobile trail to junction with main road. After crossing Hoist Creek bridge and meeting another road junction, take the trail to the south and then follow flags to Area C outcrops ---

The exposures at the crest of this hill (Area C) are of coarse-grained gabbroic anorthosite, which occurs as a large inclusion in the Dam Five gabbronorite. The inclusion is composed of 90-95% moderately foliated plagioclase, 3-7% ophitic augite and 2-3% poikilitic oxide. It is mapped as an inclusion of the Katydid Lake gabbroic anorthosite (klga; see location map), but its ophitic texture is more like lithologies found in the deeper anorthositic series of the Duluth Complex (**as**, Fig. 2-4; **olga**, Fig. 2-12A).

--- Return to road junction and continue west and north to clear cut area. Follow flags over crest of hill to outcrop ledge---

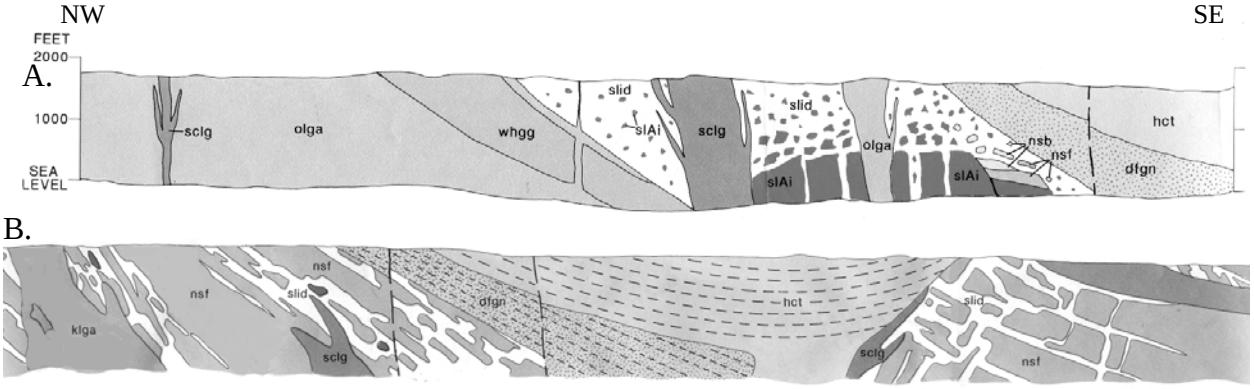


Figure 2-12. Interpretive cross-sections through the northern Beaver Bay Complex from geologic maps by A) Boerboom and Miller (1994) and B) Miller and others (1994). Map unit abbreviations: **nsf**- felsic volcanics, **slid** - Shoepack Lake inclusion-rich diorite; **slAi** - Archean inclusions in **slid; whgg** - Whitefish Lake granophyre, felsic series of the Duluth Complex; **olga, sclg, klga** - phases of the anorthositic series of the Duluth Complex; **dfgn** - Dam Five gabbronorite; **hct** - Houghtaling Creek troctolite.

Exposed in a 5' ledge on the western crest of the clear cut hill (Area D) is a typical exposure of the earliest intrusive unit of the Beaver Bay Complex - the Shoepack Lake inclusion-rich diorite (**slid**). This unit is composed of a heterogenous mix of aphanitic to very fine-grained diorite and inclusions of predominantly felsite (icelandite and K-feldspar-phyric rhyolite) and lesser amounts of basalt, gabbroic anorthosite, gabbro and locally Archean? rock types (granite, gneiss, amphibolite, schist). Inclusions range in size from single quartz xenocryts to blocks 50m across. Most inclusions exposed here are of felsic volcanic rocks irregularly mixed and partially assimilated into the diorite. However, at the northern end of the exposure, is an inclusion of a cross-bedded interflow sediment hornfels. The rock is composed of a granular mix of plagioclase, oxide and minor pyroxene, with oxide and plagioclase defining graded bedding. The pyroxene (~5-10%) is probably metamorphic.

To the north, Archean xenoliths are common. These occurrence reinforce the interpretation that the pronounced gravity low in this area reflects a prominent southeast-trending basement ridge that divides the Keweenawan igneous rocks into two basins of intrusion and eruption (Boerboom, 1994; Miller and Chandler, 1997). The Shoepack Lake diorite is interpreted to be an intrusive breccia that was the first northeast-trending intrusion to breach this crustal ridge and intrude into a rhyolite-dominated part of the NSVG overlying the ridge (Fig. 2-12). This breach of the crustal ridge paved the way for subsequent intrusions, including the Dam Five gabbronorite and the Houghtaling Creek troctolite, to intrude along this same crustal weakness and thus create the composite grouping of northeast-trending intrusions that form the northern BBC (Figs. 2-4, 2-12). The presence of anorthositic series inclusions in this rock suggests that it may be one of the older intrusions in the greater Duluth Complex, perhaps correlative with early stage magmatism.

--- Return east to Co Rd 7. Turn left and proceed to Forest Rd 170. Go right about 2.5 miles and then turn left on the secondary forest road to Whitefish Lake. Proceed about 4.0 miles to a gravel pit road on left. Park and walk about 150m south on the forest road to low outcrops on the west side of the road. ---

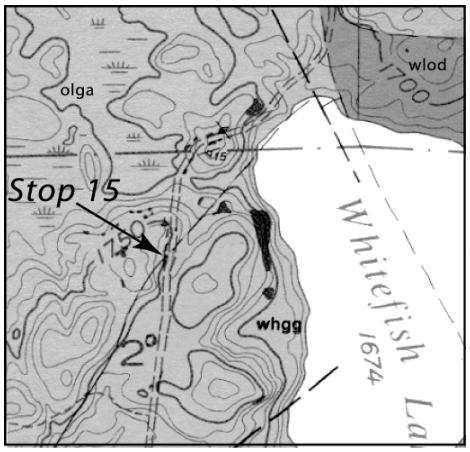
STOP 2-15: <u>Contact between the felsic and anorthositic series of the Duluth Complex</u>

T.60N., R.6W., Sec. 2

Wilson Lake 7.5' quadrangle

UTM: 646050E, 5286185N

Highlights: Contact between Whitefish Lake granophyre (whgg) and flow foliated gabbbroic anorthosite of the Outlaw Lake gabbroic anorthosite (olga).



In the low pavement exposure, a sharp contact can be observed between medium-grained pink granophyre and medium-grained, swirly foliated, ophitic oxide gabbroic anorthosite. A bimodal grain size of well-foliated plagioclase in the gabbroic anorthosite imparts a trachytic texture to the rock. Several lines of evidence indicate that the gabbroic anorthosite is intrusive into the granophyre:

1) the gabbroic anorthosite shows a subtle decrease in grain size toward the contact whereas the granophyre show no change in texture as the contact is approached. Gabbroic anorthosite in gravel pit outcrops, about 50 m to the north, is considerably coarser grained than observed here. Two miles to the northeast, a more obvious chill of the gabbroic anorthosite against granophyre is noted.
2) well-developed, but irregular-oriented foliation in the gabbroic anorthosite appears to be concordant within a meter of the contact; and
3) felsic apophyses emanate from the granophyre and into the gabbroic anorthosite, probably caused by back veining of partially melted granophyre at the intrusive contact.

This exposure provides rare field evidence for the older age of the felsic series relative to the anorthositic series of the Duluth Complex. This inferred age relationship was reaffirmed by U-Pb zircon dating, which yielded an 1109.6±4 Ma age for the Whitefish granophyre (Sandland and others, 2001) compared to 1099.1 and 1099.6 Ma ages for two anorthositic series samples dated by Paces and Miller (1993). The early stage emplacement of felsic magmas, which gave rise to the granophyre bodies that now dot the upper contact of the Duluth Complex (Fig. 2-3), helps explain the subsequent emplacement of the multiple, massive intrusions of the Duluth Complex. These felsic bodies probably acted as density traps that triggered underplating of first, ferrogabbroic magmas, creating the early gabbro series, then, plagioclase-phyric magmas (Fig.2-12A), creating the anorthositic series, and finally, aphyric mafic magmas, creating the many discreet intrusions of the layered series (Miller and Ripley, 1996; Miller and others, 2002).

In the gravel pit are exposures of several varieties of anorthositic rocks that show complex lithologic and structural relationships typical of the anorthositic series (Miller and Weiblen, 1990).

*--- Return south to Forest Rd 170, then west to Co Rd 7. Proceed north (right) to Forest Rd 172 (Wanless Rd). Go left on 172 for about 2.9 miles to quarry entrance on right side of road (**Permission required for entry)** ---*

STOP 2-16: "Lake Superior Green"quarry

T.60N., R.7W., Sec. 35

Silver Island Lake 7.5' quadrangle

UTM: 636532E, 5278342N

Highlights: Dimension stone quarry; porphyritic gabbroic anorthosite (klga); diabase dikes.



Cold Spring Granite Company has been operating a dimension stone quarry at this site for the past 10 years. The rock is marketed as Lake Superior Green, but its geologic name is the Katydid Lake gabbroic anorthosite (Boerboom and Miller, 1994; Miller and others, 1994). The rock is a coarse-grained plagioclase-porphyritic gabbroic anorthosite and is composed of 60-70% large (2-3 cm), blocky plagioclase phenocrysts in a medium-grained matrix composed 65% plagioclase, 20% prismatic augite, 5-10% granophyric mesostasis, and 5-10% accessory phases of oxide, biotite, amphibole, and apatite. The subequant habit of the plagioclase precludes the development of a foliation, though locally some hint of alignment is evident. Locally, small (10-30 cm) rounded inclusions of coarse-grained anorthosite or granulated (tectonized) anorthosite are found. Along the southern wall of the quarry, the gabbroic anorthosite is cut by 10- to 50-m-wide dikes of fine-grained diabase.

This rock is thought to be an offshoot from the main mass of the anorthositic series, which occurs about 2 kilometers to the north and is locally called the Outlaw Lake gabbroic anorthosite (Boerboom and Miller, 1994).

END OF TRIP

---Continue west along Forest Road 172 10 miles to MN Hwy. 1. Turn left and proceed southeast to Hwy 61 at the shore. Turn right and proceed to Duluth---

References

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FIELD TRIP 3

LATE WISCONSINAN SUPERIOR-LOBE DEPOSITS IN THE SUPERIOR BASIN NORTHEAST OF DULUTH

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INTRODUCTION

The recession of the Superior lobe was marked by many readvances. We will examine glacial striations and tills of the Superior lobe along the scenic North Shore of Lake Superior from Duluth to Two Harbors. The stratigraphy in this area will be tied to ice margins and events elsewhere, leading to a consistent glacial history. Each time the ice lobe retreated into the Superior basin, it created a glacial lake, and each subsequent readvance incorporated reddish-brown lake sediment. Each of these lakes is here considered to be a phase of glacial Lake Superior in a broad sense. Each readvance covered a smaller area and reached a lower elevation than the last, and deposited a finer-grained till than the previous one.

STRATIGRAPHY

Cromwell Formation

The oldest glacial sediment exposed in these stops is a rocky, gravelly, reddish-brown (5YR 4/4) till. Its matrix texture is typically a loam, but some samples are extremely low in clay (Fig. 3-1). It is non-to-slightly calcareous, and is deeply leached where exposed at the surface, even where calcareous at depth. This till and associated meltwater sediment has been named the Cromwell Formation (Wright and others, 1970). The characteristic reddish color and abundant red and black clasts, which define the Superior provenance, are derived from the passage of the ice through the Superior basin, and are presumably absent from deposits up-ice from the basin. The till matrix derived its reddish color and some of its red clasts from pulverized red sedimentary rock from the floor of the basin: sandstone, siltstone, and shale. The other red and black rock fragments are primarily extrusive and intrusive igneous rocks from the northwestern lake shore: basalt, rhyolite, diabase, and gabbro. Black slate fragments from the Proterozoic Rove Formation are locally present.

In most of its extent (down-ice from the basin), the Cromwell Formation contains many clasts other than those from the Superior basin: gray to black Proterozoic metamorphic clasts, Archean granite and greenstone, and Paleozoic carbonate. These clasts are mainly derived from older Quaternary sediment and local bedrock. Within the Superior basin, the clast assemblage in the Cromwell Formation in many places is dominated by Superior basin clasts. I call this assemblage a “hyper-Superior” assemblage, as opposed to the mixed Superior assemblage down-ice. Inasmuch as the ice itself did not originate inside the Superior basin, the scarcity of clasts from up-ice of the basin implies that the rate of bedrock erosion is much higher within the basin than outside, which explains why the basin formed in the first place. The standard grain size for clast counting and determining provenance is 1-2 millimeters (very coarse-grained sand), and this size is used unless otherwise specified.

Barnum Formation

The remainder of the glacial and glaciolacustrine sediment in the area is referred to as the Barnum Formation (Fig. 3-2). The Barnum Formation was originally defined to include clayey

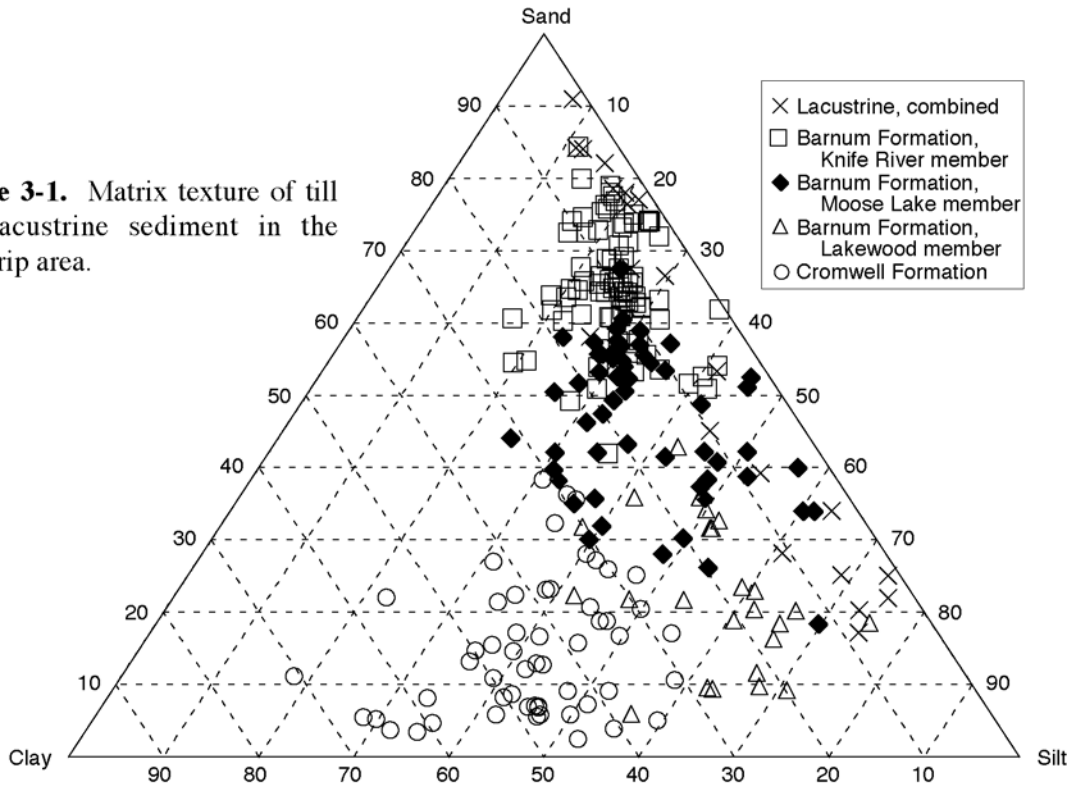


Figure 3-1. Matrix texture of till and lacustrine sediment in the field trip area.

till overlying the Cromwell Formation (Wright and others, 1970). Hobbs (2002, 2003a, b) recognized three texturally distinct tills above the Cromwell Formation. The tills are overlain and separated in places by reddish lake and deltaic sediment, which ranges in texture from silt to clay (Fig. 3-1), and contains very little sand and almost no coarse-grained fragments. This sediment ranges from massive to laminated, and from grayish-brown to reddish-brown. More detail is provided in the stop descriptions.

The individual tills contain inclusions of similar lake sediment in places. These tills and lacustrine sediments are to be included in a modified Barnum Formation (Johnson and others, unpub. data). Until this work is published, the member names will be used informally.

Lakewood member

The lowest till in the Barnum Formation is the Lakewood member. It is typically reddish-brown (5YR 4/3 to 3/4) noncalcareous silt loam (Fig. 3-1). Its average clay content is only a little more than that of the Cromwell Formation. Its proportion of coarse-grained fragments is typically less than that of the Cromwell till, but more than that of the other tills of the Barnum Formation. In contrast to the Cromwell Formation, large pebbles and boulders are not common. Most samples have a hyper-Superior clast assemblage in the 1-2 millimeter size fraction. A few samples have a small number of Paleozoic carbonate clasts. The till is relatively thin and patchy in the field trip area; it probably averages less than 2 meters in thickness.

Moose Lake member

The middle till in the Barnum Formation is the Moose Lake member. It is typically reddish-brown (5 to 2.5YR 5/3 to 4/4) calcareous clay loam to silty clay loam (Fig. 3-1). Coarse-grained fragments are sparser than in the brown silty till of the Lakewood member, and large rocks are rare. The typical clast assemblage is Superior, but most are not hyper-Superior. The Moose Lake

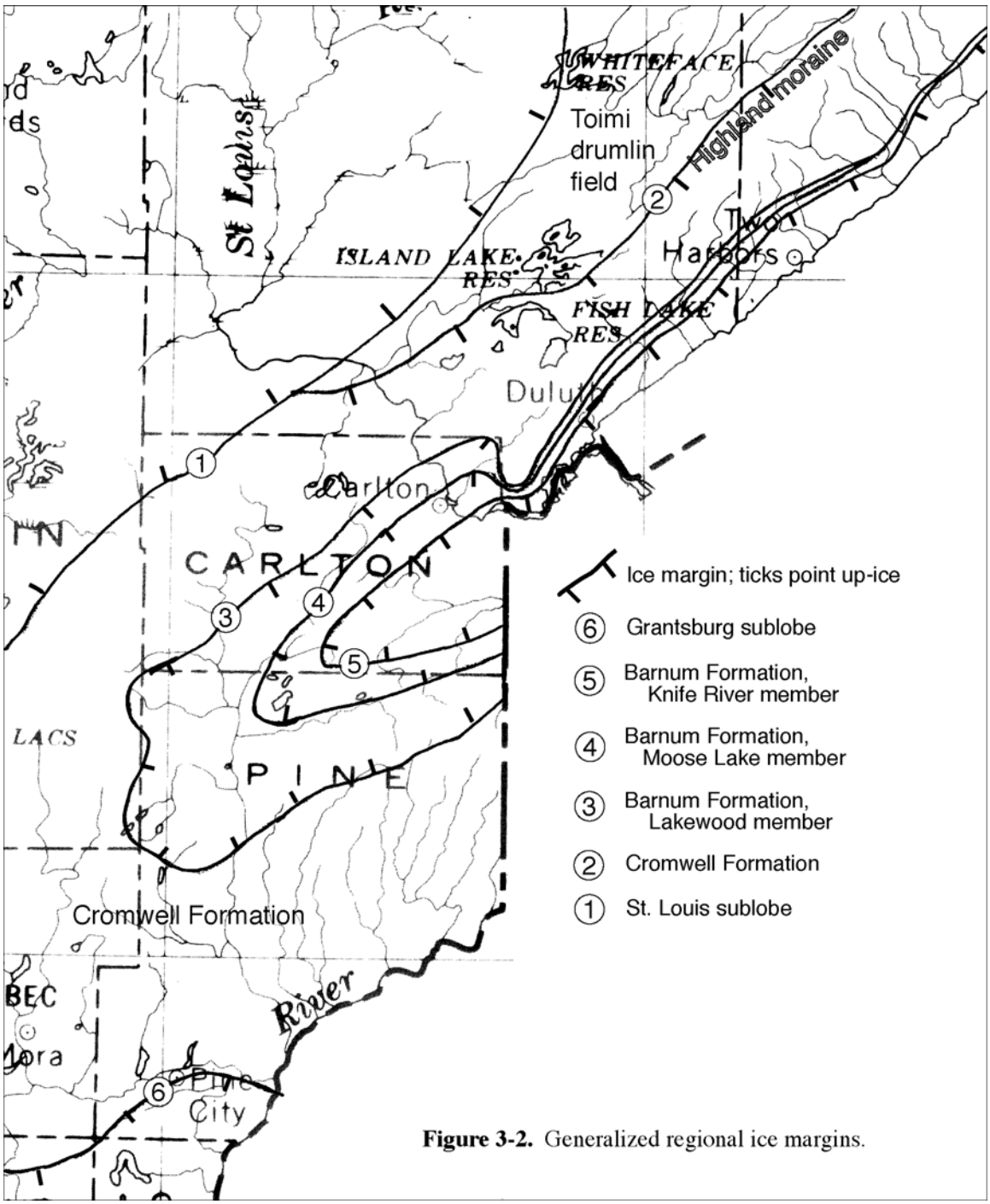


Figure 3-2. Generalized regional ice margins.

member contains more clasts from outside the Superior basin (mostly granitic) than does the Lakewood member. Many samples contain secondary carbonate precipitated by soil processes, but fewer than half of the samples have primary carbonate clasts. The number of primary carbonate clasts is small, even in samples that appear to be quite calcareous, which suggests that the bulk of carbonate is in the finer grain sizes. The average thickness of the till in the field trip area is about 3 meters.

Wrenshall member

Laminated red and gray lacustrine silt and clay in the Nemadji lake plain was named the Wrenshall Formation by Wright and others (1970). It is stratigraphically above the Moose Lake member, and behind the Nickerson moraine at the margin of the Moose Lake member. I am redefining the Wrenshall Formation as a member of the Barnum Formation, because it is a part of the same stratigraphic package of glacial lake sediment and lake-influenced tills. This member has not been observed in the field trip area.

Knife River member

The upper till of the Barnum Formation is informally named the Knife River member. It is typically reddish-brown (2.5 YR 5/3 to 4/4) calcareous clay (Fig. 3-1), containing few coarse-grained fragments or large rocks. A rough field criterion for differentiating the Moose Lake and Knife River tills is the sound they give off to a shovel. The Moose Lake member goes "clink," but the Knife River goes "clunk." That is, almost every shovel thrust into the Moose Lake till encounters a pebble, but most shovel thrusts into the Knife River do not. Where the Knife River till overlies the till of the Moose Lake member, its color is commonly on the same color chip, but a little lighter and redder. The clast assemblage is commonly Superior, rarely hyper-Superior, and occasionally a mixed Superior-Winnipeg assemblage (the Winnipeg assemblage is found in tills deposited by ice that advanced through the Winnipeg lowland. It is characterized by common to abundant Paleozoic carbonate, granitic grains that are more common than dark grains, few red grains, and little to no Cretaceous shale). More than half of the samples contain primary carbonate grains, and many samples contain common to abundant secondary carbonate. Its average thickness is 3 to 4 meters, but appears to be absent in many places close to Lake Superior. Its clay texture is atypical for glacial till, which led Moss (1977) and Gross (1982) to interpret this till as massive glacial lake clay.

GLACIAL HISTORY

During the most recent glaciation, the field trip area was dominated by the activities of the Superior lobe, which advanced through the axis of the Superior basin from northeast to southwest as far south as the St. Croix moraine (Wright, 1972). The advance to the maximum, and the earlier retreatal phases, deposited till and associated sediments of the Cromwell Formation. Many retreatal or readvance ice margins have been recognized between the St. Croix moraine and the Superior basin (Mooers, 1988; Patterson, 2001), but only the most recent ones are discussed here. As long as the ice did not front a substantial glacial lake between advances, the texture of the till remained more or less the same: rocky loam to sandy loam.

The first retreat into the Superior basin did not create a large continuous lake, but rather a set of small ice-marginal lakes trapped between the ice and the sides of the basin. Meltwater escaped by a series of ice-marginal channels, building ice-contact deltas of sand and gravel into ponds on its way out of the basin to the southwest (Stop 3-8). The next advance of the Superior lobe overrode these deltas and deposited a loamy Cromwell till that ends at the Mille Lacs–Highland moraine. The texture of this till does not reflect any large amount of overridden lacustrine sediment. An outwash plain graded to the Highland moraine buries the adjacent part of the Toimi drumlin field. It is graded to a set of channels that seems to end at the St. Louis River. The present day St. Louis River flows east and south, into the Superior basin, so meltwater flow from the Highland moraine maximum must have flowed west (reversed) into glacial Lake Upham I. The flowpath out of Lake Upham I is unknown; deposits of the St. Louis sublobe later covered this whole area (Fig. 3-2).

As the ice receded from the Highland moraine, a set of meltwater channels was formed between the receding ice in the basin, and the higher ground to the northwest. These channels are discontinuous, and must have partly formed on the ice itself. Their general flow direction is to

the southwest, and they probably exited the Superior basin at its southwestern corner, where bedrock elevations are the lowest. Deposits of subsequent readvances have since covered this area, but the bottom of the most recently used lake spillway is about 1,050 feet in elevation. Gravel in these channels is somewhat younger than the gravel in the ice-contact deltas, but is also included in the Cromwell Formation. Further retreat opened up a lake in the southwestern corner of the Superior basin that extended about as far northeast as Two Harbors.

The advance that followed was named the Split Rock phase by Wright (1972). It spilled out of the basin and deposited a thin, reddish-brown (5YR) till, which I have correlated to the Lakewood member of the Barnum Formation. This advance extended well out of the Superior basin deep into Pine County, where it apparently formed the Split Rock, Askov-Lookout Tower, and Kerrick ice margins. This correlation is uncertain, and the till texture is not exactly the same in Pine County as in the field trip area, but till behind this set of margins is finer-grained than that of the earlier phases (Patterson, 2001). The Lakewood member does not exhibit a morphologically distinct ice margin in the field trip area. Its margin is recognized only by the change in texture of the surface till, from that typical of the Cromwell Formation to that typical of the Barnum Formation. This change occurs fairly consistently between 1,150 and 1,200 feet in elevation.

All three till members of the Barnum Formation have fairly consistent margin elevations in the field trip area. The ice was confined in a steep-sided basin, so their margins are similar to bathtub rings. Theoretically, one would expect the margins to rise to the northeast, both to account for the down-ice slope of the glacier, and differential postglacial rebound. There is little evidence of this rise in the margin of the Lakewood member. This is partly because the data points on which it is based are widely separated, and partly because the mapped margin of the member is not necessarily the actual ice margin, but the margin of preserved deposits. A thin patchy till deposited on a sloping surface is subject to extensive erosion, both by waves and by sheet runoff. In addition, the area mapped so far is rather small, and the expected rise in ice margin is not great.

Another retreat opened up a larger portion of glacial Lake Superior; it is unclear how much of the basin was ice-free at this time. Apparently, this lake accumulated a considerable amount of clay, because the advance that followed deposited the clayey Moose Lake member and built the Nickerson moraine at the southwestern margin of the basin. In the field trip area, the Moose Lake member was deposited up to about 1,100 to 1,150 feet in elevation. If the glacier were merely recycling its own clay, it is puzzling why the Moose Lake member is so much more clayey than the Lakewood member. Another puzzle is why the Moose Lake member is so calcareous compared to the Lakewood member.

One possibility is a change in the nature of debris carried into the Superior basin from the central Laurentide Ice Sheet. For example, Mooers and Lehr (1997) proposed that the change from calcareous to noncalcareous debris carried by the Rainy lobe (in a broad sense) was correlated to the migration of the ice divide south of the Hudson Bay lowland, so that the Paleozoic carbonates in the lowland could no longer be transported south. The eastern boundary of these lowlands is up-ice from the Superior basin, and could have supplied some carbonate. In fact, a small number of Paleozoic dolomite grains are sporadically present in the otherwise noncalcareous Cromwell Formation and Lakewood member tills, so a Hudson Bay component is quite possible. However, the mechanics of how a reverse transition from noncalcareous to calcareous would work in this setting is difficult to envision.

Another possibility is that glacial Lake Upham I drained into Lake Superior between the Split Rock and Nickerson advances. This lake may have been fed by meltwater from the calcareous Itasca or Koochiching lobes to the north, or even the incipient St. Louis sublobe. If so, some of the meltwater probably funneled down the St. Louis River to glacial Lake Superior. Fine-grained calcareous lake sediment was thus potentially available for entrainment into the Moose Lake till. Although the sediment carried down the St. Louis River should have been mostly brown and

gray, there was abundant red clay deposited in the lake from the melting Superior lobe, and the red color tends to dominate in mixtures (the gray color is not the result of a pigment, but the lack of pigment).

The later advance of the St. Louis sublobe filled the Aitkin-Upham basin and buried the deposits of Lake Upham I. This advance coincided more or less with the Nickerson moraine, as shown by diversion channels around the moraine that can be traced upstream to the St. Louis River (Wright and others, 1970). There is no way to be sure which of these channels were fed by outwash from the St. Louis sublobe, and which were fed by overflow from glacial Lake Aitkin-Upham. The general pattern is that the earlier, higher diversion channels are narrower, and presumably carried less water than the later, lower ones. One might speculate that the earlier diversion channels carried meltwater from the ice margin, and that the later ones carried water from the glacial lake. At times, Lake Aitkin-Upham was fed by overflow from Lake Koochiching (the eastern arm of Lake Agassiz) at its highest levels (Hobbs, 1983). Thus, a much larger flow of water was available than could have been generated by local melting in the St. Louis River watershed.

The succeeding retreat first opened up a small ice-marginal lake in the southwestern corner of the basin, called glacial Lake Nemadji by Winchell (1899). This lake appears to have been small and fronted by ice at first—the rapid decline of its depositional surface from the proximal edge of the Nickerson moraine to Duluth can be most easily explained if the Superior lobe still occupied most of the basin. The outlet of Lake Nemadji was the Portage River channel, which cuts through the Nickerson moraine and joins a diversion channel of the St. Louis River at the town of Moose Lake. The bottom of the Portage River outlet is about 1,050 feet in elevation, but lake-smoothed topography and minor beach ridges can be recognized up to 1,080 feet. This does not mean that the water in the outlet channel was necessarily 30 feet deep; the higher wave-washed topography could have formed while the channel was being deepened. A great thickness of laminated glacial clay and silt (Wrenshall member) accumulated at this time, between the proximal side of the Nickerson moraine and the ice remaining in the basin.

Further retreat of the ice uncovered the lower Brule outlet in Wisconsin, which lowered the level of the lake to about 1,020 feet and dried up the Portage outlet (Leverett, 1932). This lower level of the lake has been called glacial Lake Duluth, but I prefer to use Duluth and Nemadji as levels or phases of glacial Lake Superior, rather than giving every phase a separate lake name.

A large outflow from glacial Lake Aitkin-Upham II came down the St. Louis River after the Superior lobe melted back from the Nickerson moraine. It washed a large area (about 10 square miles) down to bedrock, upstream from the Nemadji lake plain. Although this scouring event was extremely widespread, it did not expose any substantial area of bedrock below 1,050 feet in elevation, suggesting that the local base level was no lower than 1,050 at the time of the event. Subsequent stream erosion has incised a narrow valley into the bedrock below 950 feet in elevation, but this cutting seems to have been accomplished by long erosion of a much smaller stream graded to a much lower base level. Thus, the last major meltwater flood down the St. Louis River seems to have come during the short time that glacial Lake Superior stood at the Nemadji level.

As the Superior lobe retreated, the lake declined in stages as lower outlets were uncovered, until the entire lake basin was ice-free. The water level was much lower than it is today (at least in the field trip area). As the ice retreated north of Lake Superior, Lake Agassiz spilled east into Lake Nipigon, into Lake Superior, and then out into Lake Huron (Clayton, 1983). A considerable amount of suspended sediment from glacial Lake Agassiz was deposited in the basin. Combined with the earlier sediment from Lake Aitkin-Upham, the Superior basin now had a considerable amount of clay-rich calcareous sediment to be entrained by the next glacial advance. Lake Superior remained ice-free for about 1,000 years, from 1100 to 1000 B.P. (Drexler and others, 1983).

The last, or Marquette advance, filled the basin with ice again, depositing a stone-poor, very clayey 2.5YR till, which I have included in the Knife River member of the Barnum Formation. This till is much more clayey on average than the clayey Moose Lake member (Fig. 3-1); its matrix is more calcareous, and it contains more actual carbonate clasts. These differences are attributed mainly to incorporation of sediment from Lakes Agassiz and Aitkin-Upham. In the field trip area, the Knife River member extends up to a little less than 1,050 feet in elevation; it extends laterally over the Wrenshall member as far as the proximal side of the Nickerson moraine (Howard Mooers, unpub. data).

The field trip takes place in the area mapped in the French River, Lakewood, Knife River, Two Harbors, and Castle Danger quadrangles (Fig. 3-3). I have recently mapped the surficial deposits of the first 4 quadrangles under the USGS STATEMAP program (Hobbs, 2002, 2003a, b). A surficial map of the Castle Danger quadrangle is in progress. I combined the knowledge gained from this mapping with earlier mapping and interpretations in the region to create this new interpretation of Superior lobe recessional history.

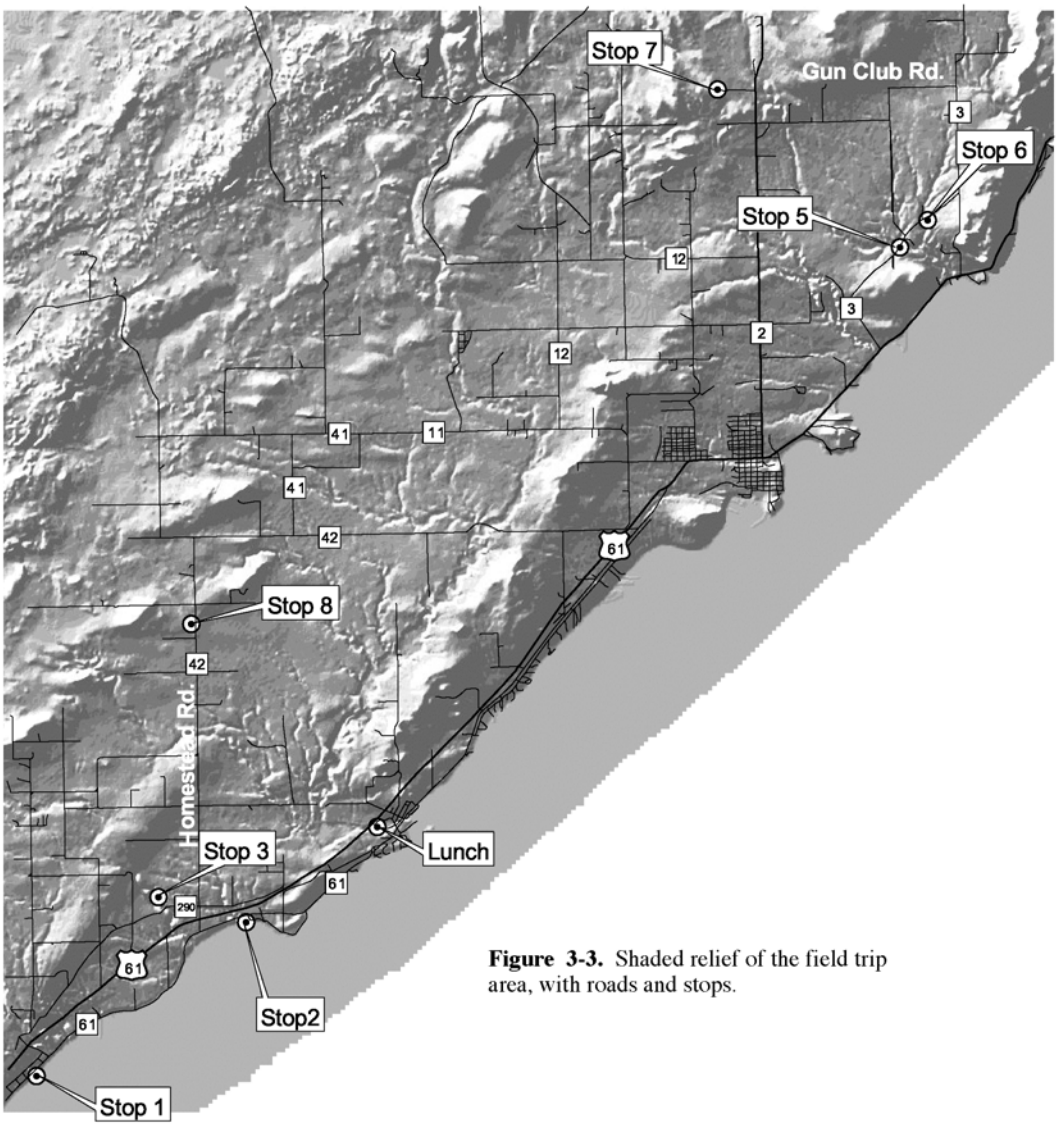


Figure 3-3. Shaded relief of the field trip area, with roads and stops.

FIELD TRIP STOPS

8:00 a.m. Leave Radisson Hotel, Duluth; get onto I-35/ Highway 61, heading northeast through Duluth. Just beyond the where US Highway 61 becomes a 4-lane expressway, turn right onto the Scenic North Shore Drive (County Road 61). Continue about 6 miles to the village of French River.

Stop 3-1. Superior bluff cut at French River. French River Quad., T.51N., R. 12 W., Sec 19 NE corner; 583815E, 5194075N NAD83 UTMs

This wave-cut bluff exposes about 25 feet of glacial sediment above basalt bedrock. The top of the bedrock is 3 to 4 feet above lake level. It has been broken up by weathering and wave action in most places, but there are some surfaces that have been smoothed by glacial abrasion. Striations strike northeast, almost parallel to the shore but a little more westerly. These striations were carved by the main flow of the Superior lobe through the Superior basin. Wherever striations are exposed by wave action along the shoreline, they have a similar orientation.

The lower third of the cut is the Lakewood member of the Barnum Formation [(5) 28 63 9] over till of the Cromwell Formation [(18) 52 42 6]. Numbers in brackets represent individual texture samples: (gravel) sand, silt, and clay. Sand, silt, and clay are considered matrix, and their numbers add to 100 percent, rounded to the nearest 1 percent. The number in parentheses is the total of all fragments larger than 2 millimeters, and is expressed as a percent of the original sample: matrix plus coarse-grained fragments. Large rocks were avoided for the texture samples. Both tills are firm but not hard, with a platy structure parallel to the bluff face. The platy structure is probably a weathering phenomenon. These tills are a little less red than they usually appear—7.5 YR rather than 5YR. The Cromwell sample has a hyper-Superior grain assemblage; the Lakewood sample has a Superior assemblage with a trace of carbonate.

The middle third of the bluff cut is red clayey till of the Moose Lake member: [(1) 13 29 58 and [(2) 19 30 51]. Its lower contact is flat and sharp. Both the middle and upper tills are fairly firm, with a strong angular blocky structure. The structure and consistency are the best way to distinguish in-place till from the soft structureless mudflow sediment that coats the bluff. Two samples of the Moose Lake member have Superior grain assemblage with a trace of carbonate.

The upper third of the bluff is red clay till of the Knife River member: [(0.1) 4 22 74] and [(0.2) 2 26 72]. Both tills are reddish-brown (2.5 YR 5/4); the contact between them was not observed, but inferred from the texture of the samples. This till contains irregular, subhorizontal clay inclusions, which look grayish by contrast to the red till. They range from pinkish gray (7.5YR 6/2) to light brown (7.5YR 6/3). These inclusions are more calcareous than the red till; they are interpreted as incompletely digested clumps of Lake Agassiz-derived clay. The texture samples from this interval contain some of these clay inclusions, which explains why they are so poor in gravel and sand, and so rich in clay, even for till of the Knife River member. Each sample contains only a tiny number of 1-2 millimeter grains—one is a Winnipeg assemblage, one is a Superior-Winnipeg mixture.

NEXT: Continue along the scenic drive about 4.5 miles. Turn right on a road leading to a scenic overlook. Park and scramble down the bluff. Notice how the stone wall has broken and dropped down in places where the red clay till it was built on has slumped. This bluff is actively retreating by wave action and slope failure. The stop is the bluff just across a small gully west of the scenic overlook.

Stop 3-2. <u>Superior bluff by overlook</u>. Knife River Quad., T.51N., R.12W., Sec 2, SE of SW of NW; 5888830E, 5197725N NAD 83 UTMs

The bluff is about 28 feet high, but only the upper half is exposed (unless a storm has washed the lower part since I saw it). The top of the exposed bluff is not at its original elevation. There are a few slump scarps hidden in the trees above the top, and their combined offset makes the top of the bluff 10 to 12 feet lower than the original surface. No bedrock is exposed at this site.

The lowest part of the exposure is 4 feet of till of the Cromwell Formation [(5) 36 59 5] and [(14) 43 48 9]. This till is unusually low in clay, even for Cromwell till. It is calcareous, apparently protected from leaching by the calcareous tills above it. Both samples of Cromwell till contain hyper-Superior grain assemblages.

Above the Cromwell till is a layer of brown silty till [(11) 17 59 24] only 2 feet thick, overlain by a foot of massive silt [(1) 7 73 20], which is similar in color and texture to the silty till, but lacks stones. The brown silty till is included in the Lakewood member of the Barnum Formation; the brown silt could also be included in the Lakewood member. It could either be massive lacustrine, or actual till that happens to contain few coarse-grained fragments. More digging may help us figure this out. Both are calcareous. Both contain hyper-Superior grain assemblages with no carbonate grains, but the 1-2 millimeter sample from the massive silt is very small.

Above the massive silt is a layer of interbedded red (2.5YR 4/4) and gray (10YR 4/1) calcareous lacustrine clay just over a foot thick. A sample of the red clay is slightly more clayey [0.2) 3 15 82] than the gray [(0.7) 3 18 79], but the gray clay feels more slippery, possibly because of a higher content of expandable clays. Both have very small Superior grain assemblages. The main red and gray beds are about an inch thick, but there are red and gray micro-laminations within each bed. There are no silt beds, making it unlikely that these beds represent varves. The bedding is more or less horizontal, but not very regular. Beds pinch and swell; is this original, or were the beds deformed by overriding of the next glacier? There are numerous slickensided joints running through both the brown silt and the red and gray clay. Were these caused by overriding ice, or by modern slumping?

The uppermost layer exposed in this bluff is about 2 feet of calcareous red clayey till [(5) 31 25 44] of the Moose Lake member. It is weak red (2.5 YR 4/2) in the lower part, and reddish brown (2.5YR 4/4) in the upper part. A sample has a Superior grain assemblage. Red clay till of the Knife River member is not exposed here, but it may be present at the site. The top of the exposure is at least 3 feet below the bluff top, and the bluff top itself is probably not the original surface. A soil boring taken along the Scenic Highway from a bluff about 20 feet higher than this one penetrated about 16 feet of red clay till over lacustrine sediment, with an intercalated contact.

NEXT: Leave the parking lot, drive back north to the Scenic Highway, turn left and go west about 0.5 mile to the intersection of Homestead Road (County Road 42). Turn right and go north across the railroad tracks and the Highway 61 expressway, and turn left on Old North Shore Road (County Road 290). Proceed about 0.5 mile west into the Big Sucker Creek valley, and park by the bridge. Walk upstream about 0.3 mile, following the trail on the northeast side of the stream. The trail traverses a slumped stream cut.

Stop 3-3. <u>Big Sucker Creek cut</u>. Knife River Quad., T.51N., R. 12W., Sec 4, SW of NE of SE of NE; 586740E, 5198325N NAD83 UTMs

The creek has cut down to and into bedrock in this whole area. At this site the basal 2 feet of the stream cut is basalt. The base of the drift section is rocky till of the Cromwell Formation, about 4 feet thick. Color is dark reddish brown (5YR 3/2). The lower sample is gravelly and very low in clay [(31) 48 46 6]. The upper sample is more clayey but even more gravelly than the

lower one [(37) 34 47 19]. The till is slightly calcareous, but the grain assemblage is hyper-Superior with no carbonate grains.

The Cromwell till is overlain by about 9 feet of the Lakewood member; this is one of the thickest Lakewood layers exposed in the field trip area. The trail crosses the cut at the level of the Lakewood member. This till is dark reddish brown (5YR 3/2), slightly to moderately calcareous, less rocky than the Cromwell till, but much more so than the overlying till. The results of three texture samples were: [(12) 24 54 22], [(6) 16 50 34], and [(11) 17 51 32]. All samples have a hyper-Superior grain assemblage, with only one carbonate grain among the three.

The Moose Lake member is not exposed here. The contact between the Lakewood member and the overlying Knife River member is above the trail, and is not exposed. What appears to be the contact is actually a slump plane. Till of the Knife River member is not well exposed, and only one sample was obtained [(1) 7 22 71]. It is reddish brown (2.5YR 5/4), slightly to moderately calcareous, and has a Superior grain assemblage, with a few carbonate grains and one grain of spherical frosted quartz derived from Paleozoic sandstone.

NEXT: Return to the road, drive back to Homestead Road, and turn right (south). Turn left (east) on the Highway 61 expressway, and drive about 3.5 miles, cross the Knife River, and turn right at the first street leading to the town of Knife River. Turn right at the first intersection, by the school, and head back south and west to a rest area overlooking the river.

Stop 3-4. Lunch at Knife River rest area. Knife River Quad, T.52N., R.11W., Sec 31, NE of SW of NW; 591965E, 5200050N NAD83 UTMs

While eating lunch, we can observe the erosion along the Knife River, and attempts to control it. This stream, like most streams emptying into Lake Superior, is fairly short, with a steep gradient and a greatly fluctuating discharge. Most of the time the level is low, but snowmelt and big rainfalls can turn it into a roaring torrent.

NEXT: Return to Highway 61. Drive about 10 miles northeast, through Two Harbors. About 2.5 miles northeast of Two Harbors turn left (northwest) onto Lake County Road 3. Drive up the hill, and down into the Silver Creek valley. Park where the road crosses Silver Creek, about 2 miles from the turnoff. Walk 200 feet west (upstream) along the creek to a cut on the outside of a meander bend.

Stop 3-5. Silver Creek cut. Castle Danger Quad., T.53N., R. 10W., Sec 16, SE of NW of SW of SW; 604470E, 5213910N NAD83 UTMs

This is a small cut, but it illustrates some things rarely seen in this area: a good exposure of postglacial stream channel and overbank sediment, and a layer of artificial fill. The total height of the cut is about 7 feet. It exposes a foot of red clay fill over alluvium (overbank and channel sediment) over red clay till of the Knife River member, over brown sandy till of the Cromwell Formation.

The Cromwell till is reddish-brown (5YR 4/3), moderately calcareous, fairly hard and very pebbly [(10) 30 44 26]. Its grain assemblage is Superior, with some calcite rhombs. They all appear to be amygdule fragments. Even though the stream has only cut a few feet into the till, it has developed a continuous pavement of rocks on its bottom.

The Cromwell till is overlain by clay till [(2) 10 27 63] of the Knife River member, with a fairly sharp, wavy contact. This till is reddish-brown, on the same color chip but slightly lighter and redder than the color of the Cromwell Formation. It is calcareous, with a medium to coarse-grained blocky structure. Its grain assemblage is hyper-Superior, with no carbonate grains.

The overlying alluvium was not sampled. It consists of 1.5 to 2 feet of dirty gravel channel sediment, overlain by 1 to 1.5 feet of silt and fine-grained sand overbank sediment. There is a thin soil (possibly truncated) in the top of the overbank sediment. The gravel contains clasts up to boulder size, but large clasts are much less common in this layer than in the currently active channel sediment. The difference is that at the time the gravel was being deposited, the stream was still eroding the clay till; the stream is now eroding till of the Cromwell Formation, which has more and larger coarse-grained clasts.

The uppermost layer here is a layer of red clay that I interpret as recent fill dirt. It is texturally indistinguishable from the red clay till of the Knife River member, and was probably excavated from that material. But it is out of order stratigraphically, being underlain by a soil developed in recent alluvium, and not itself having a soil cover.

NEXT: Return to vehicle. Drive 0.75 mile uphill (northeast) and down into another small valley, a tributary of Silver Creek. Park near the double culvert. Walk downstream (south) 200 feet to a slumped cut on the outside of a meander.

Stop 3-6. <u>Tributary to Silver Creek.</u> Castle Danger Quad., T.53N., R.10W., Sec 16, SW of SW of NE; 605195E, 5214645N NAD83 UTMs

This cut is a series of slumps, so there is a mixture of good exposure and covered areas. The description is sort of a composite section. The base of the exposure is loamy till of the Cromwell Formation [(16) 39 34 27]. This texture is somewhat clayey for Cromwell till, but not unusual for inside the Superior basin. The till is slightly calcareous, unoxidized brown (7.5YR 4/2); it is not especially pebbly for Cromwell till. The grain assemblage is hyper-Superior; a large 1-2 millimeter sample contains one grain of dolomite.

The Cromwell till is overlain by red clay till of the Knife River member. There is an inclusion of Cromwell till near the base of the clay till, and another several feet higher. It is not clear whether the inclusions were created by glacial erosion, or if they are related to slumping on this slope. Elsewhere on the slope is a contact between red clayey till of the Moose Lake member [(13) 31 30 39] and red clay till of the Knife River member, at a higher elevation than the contact between the Cromwell Formation and the Knife Lake member. This suggests that some or all of the contacts may have shifted downslope.

Near the top of the cut is a contact between red clay till of the Knife River member and red lacustrine clay. Here the texture of the till is [(3) 4 24 72] and the lacustrine texture is [(0) 3 18 79]. There is scarcely any difference other than the number of coarse-grained fragments. The till has a small hyper-Superior grain assemblage with no carbonate rock fragments, though there is a grain of secondary carbonate. The lacustrine sample has a very small 1-2 millimeter fraction, half of which is secondary carbonate.

NEXT: Return to vehicle and continue northeast and north about 2 miles to the intersection of Gun Club Road. Turn left (west) and follow the road 1 mile west, 0.5 mile south, and 2 more miles west. Turn right on Lake County 2; go 0.5 mile north and turn left (west) on Reeves Road, an unpaved township road. Go up the hill, under a power line, and stop in a clearing about 0.5 mile west of County Road 2.

Stop 3-7. <u>Striated bedrock</u>. Two Harbors Quad., T.53N., R.11W., Sec 1, SE of SE of NW; 600155E, 5217670N NAD83 UTMs

There are several small hornfels outcrops here, flush with the ground. They are striated northwest–southeast, about 90 degrees from the striations we saw earlier today. These striations reflect ice moving up and out of the Superior basin toward the Highland moraine. This is the

general orientation of striations in the highlands. The landforms in this area tend to be streamlined northwest–southeast parallel to the striations. These landforms, mainly bedrock, but partly mantled by drift, are named the Highland Flutes (Wright, 1970). This hill is one of the southwesternmost of the flutes. They extend northeast inside the Highland moraine as far as Lutsen.

NEXT: Return to Lake County Road 2; turn right and go south 2.5 miles; turn right on Lake County 12, and go 3 miles west and 2.5 miles south where it dead-ends on Lake County 11 (Valley Road). Turn right and travel 7 miles south and west to Homestead Road. Lake County Road 12 becomes St. Louis County Road 41 at the county line. County Road 41 ends at County Road 42, which turns and becomes Homestead Road. Turn left on Homestead Road, travel 1.3 miles south to the entrance of a gravel pit.

Stop 3-8. Peterson Gravel Pit. Knife River Quad., T. 52N., R.12W., Sec 15, NE of SE of NW; 587620E, 5204895N NAD83 UTMs

This pit is developed in one of the many ice-contact deltas that developed between the Superior lobe and the North Shore highlands. They occur in the elevation range of 1,150 to 1,210 feet, and have been mapped in the French River, Knife River, and Two Harbors quadrangles. They have not been observed east of the Two Harbors quadrangle.

The general stratigraphy here is sand and gravel of the Cromwell Formation overlain by till of the Cromwell Formation overlain by till of the Lakewood member of the Barnum Formation. The section currently exposed may be different from the one described, due to continued mining.

The stratified unit ranges from fine-grained sand to medium-grained gravel, moderately well sorted. The color of the finer beds is dark reddish-brown (5YR 3/4); the color of the coarse-grained beds is dominated by the individual rock particles, which are mostly red, black, and dark gray. The unit is non-to-slightly calcareous; two samples show hyper-Superior grain assemblages with no carbonate grains. The gravel is subangular to subrounded; it is more rounded than the gravel in the till, but not much. It does not appear to have been transported far by water.

Depending on the orientation of the cut, the sand and gravel appears to have either trough crossbedding, or foreset bedding. On the south side of the pit, the foreset beds dip west.

The sand and gravel delta is overlain by a noncalcareous loamy till [(4) 39 52 9], ranging from 3 to 8 feet thick. The contact is sharp and wavy. Some of the underlying beds are deformed and truncated, and there is a considerable amount of sand and gravel incorporated into the till in places. The till is hard and dense, with a coarse-grained platy structure; this suggests that it was deposited under the weight of a glacier, rather than as a mudflow deposit. The grain assemblage is hyper-Superior, with no carbonate grains.

On the east side, there is a discontinuous layer of silty till [(3) 20 71 9] that overlies the Cromwell till and the delta sand and gravel. Where sampled, it overlies the delta with no intervening till. This is interpreted as till of the Lakewood member. It is noncalcareous, reddish-brown (5YR 4/4), and contains only a few stones. Its grain assemblage is hyper-Superior, with no carbonate grains.

The uppermost layer at this site is a sandy, silty, unbedded deposit, 1 to 2 feet thick, interpreted as colluvium. It does not bury a soil, but the surface soil is developed in it. It must therefore not be much younger than the glacial sediments.

NEXT: Return to Homestead Road, turn right, and travel south to the Highway 61 expressway; turn right to return south to Duluth.

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FIELD TRIP 4

Geology of the Eastern Mesabi Iron Range, Northeastern Minnesota

by

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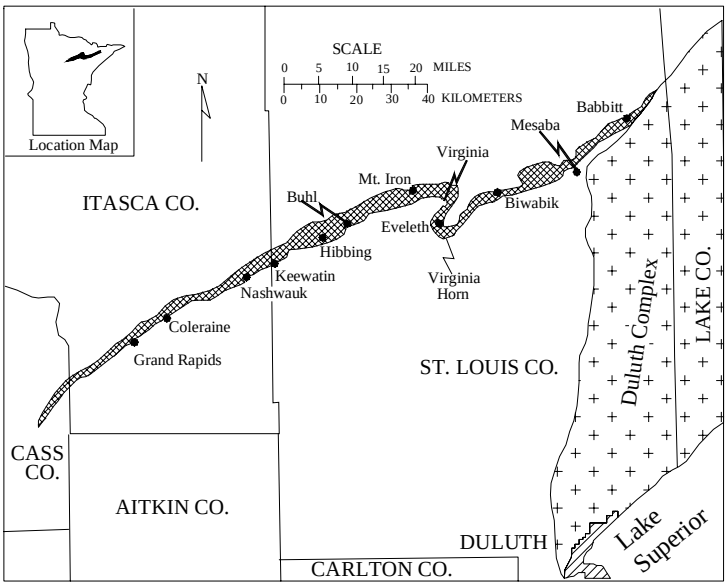
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Generalized map of the Mesabi Range (cross-hatched). Note Duluth Complex at east end.

INTRODUCTION

Iron-formation was described as early as 1866 by Henry Eames for rocks on what was to become the Mesabi Range. Several attempts were made by individuals on their way to the iron mines of the Vermilion Range to find ore on the Mesabi. However, it was not until November 16, 1890, that the first rich iron ore on the Mesabi Range was discovered by the Merritt brothers near what is now Mountain Iron, MN. In 1892, the first shipment from this mine was 4245 tons (White, 1954). Exploration for iron ore ensued and within the next few years, most of the productive parts of the Mesabi Range were discovered.

The Mesabi Range is the largest iron range in the United States and is one of the largest in the world. It is 0.25 - 3.0 miles wide and 120 miles long (cover page). The Biwabik Iron Formation, as thick as 750 ft, in general dips gently to the southeast at an angle of about 7-15 degrees. The iron-formation, called taconite, typically contains 30-40% iron and 40-50% SiO2, plus other components (Morey, 1992). In numerous places along the length of the range, silica was leached out, thereby enriching the iron content to 55%+. These pockets became the high-grade natural ore mines; there were more than 500 individual mines prior to merging into larger mines as the ore between adjacent properties was removed. These were very important in making the United States an industrial giant, and were instrumental in providing raw material for WWI and WWII. As the high-grade ore was becoming depleted, the taconite process was developed. In 1967, taconite production exceeded natural ore production. Currently, six taconite plants are in production (Fig. 4-1).

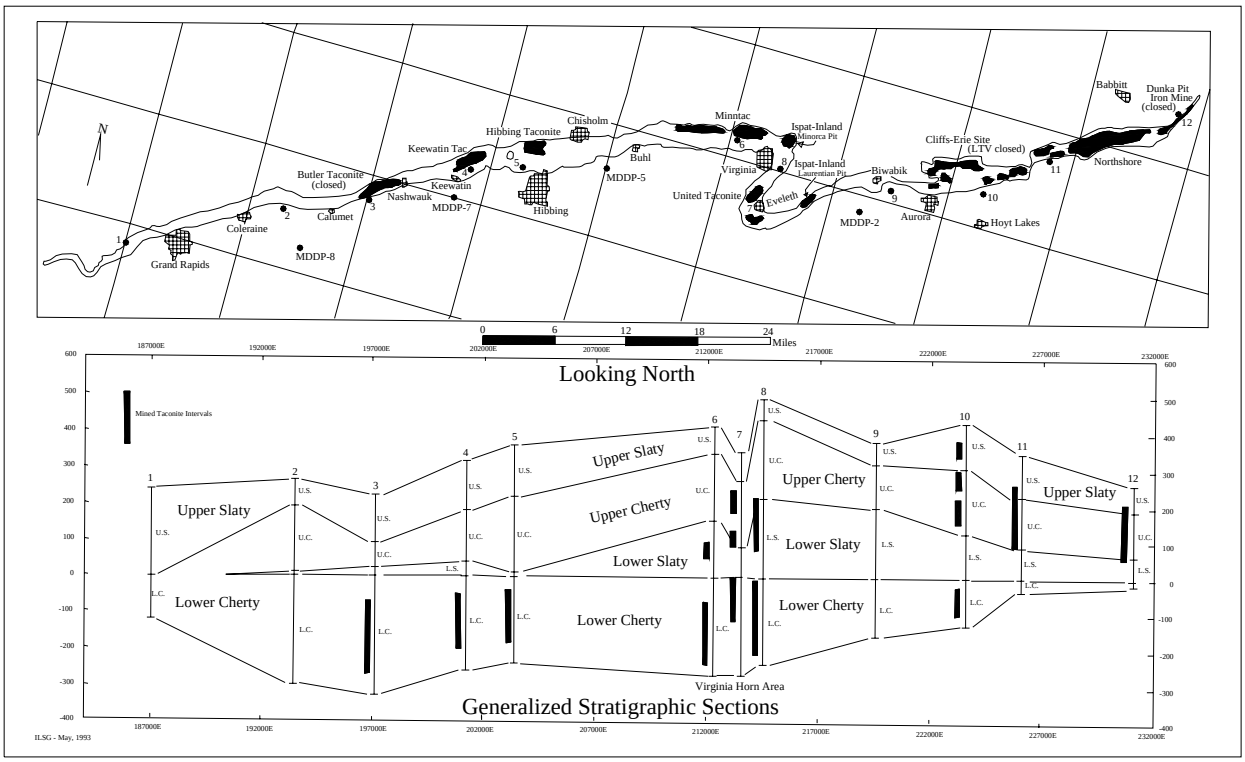


Figure 4-1: Generalized map of the Mesabi Range showing taconite pits (black) and cities, and in the lower half, a longitudinal section of the Biwabik Iron Formation (compiled by Henry Djerlev, 1993). Mined taconite intervals are shown as black columns adjacent to sections. Modified from Morey, 1993.

The name of Biwabik Iron Formation was chosen by Van Hise and Leith (1901, p. 356), "because the word Biwabik is the Chippewa word for a piece or fragment of iron." The word taconite is also used in discussions pertaining to hard, unoxidized portions of the iron-formation. H.V. Winchell (1882, p. 135) originally called portions of the BIF "taconyte" because he thought the rocks correlated with lower Cambrian rocks in the Taconic Mountains in northern New England. Since that time, many geologists have used taconite in their descriptions of the iron-formation and it has thus become firmly established. Perhaps a more proper definition for "taconite" is an economic term for iron-formation from which iron can be profitably extracted after fine-grinding, followed by magnetic separation and pelletizing (Morey, 1993).

REGIONAL GEOLOGY

The peneplaned Archean craton in the Lake Superior region formed a platform upon which a Paleoproterozoic continental margin assemblage was deposited in Minnesota, Michigan, and Wisconsin. Extension resulted in localized rifts that received thicker accumulations of sediments and volcanic rocks than did adjacent parts of the platform. Seas transgressed onto the continent one or more times and an ocean basin opened south of present-day Lake Superior. Island arcs that formed during southward subduction collided with the craton margin to the north as the ocean basin closed. A remnant of this oceanic crust is poorly preserved as a dismembered ophiolite sequence in Wisconsin (Schulz, 1987, 2003). The arc volcanics are preserved as the Wisconsin magmatic terranes. The collision resulted in a fold-and-thrust belt known as the Penokean orogen. To the north of the fold-and-thrust belt, a northward-migrating foreland basin—the Animikie basin—developed as the stacked thrusts weighed down the crust (Fig. 4-2). Thick turbidite successions were deposited along the basin axis, and terrigenous clastics and Lake Superior-type iron-formation were deposited on the shelf along the northern margin (i.e., the foreland or peripheral bulge) of the basin. See Ojakangas et al. (2001) and Severson et al. (2003) for more detailed summaries on Paleoproterozoic basin development in the Lake Superior region.

The development of the Midcontinent Rift System (MRS) at 1.1 Ga severed the basin into northwestern and southeastern segments (Fig. 4-2). If the MRS rocks are removed from the geologic map, the different portions of the Animikie basin become contiguous and the fold-and-thrust belt rocks of Minnesota and Wisconsin-Michigan become continuous (Fig. 4-3).

Figure 4-4 is an interpretive cross-section of the Animikie basin during its formative stages, with sediments derived from the Archean basement to the north and from the fold-and-thrust belt to the south.

The Paleoproterozoic supracrustal rocks in the northwestern segment, including east-central and northeastern Minnesota and the adjoining part of Ontario, are for the most part poorly exposed. However, mining of iron ore on the Mesabi and Cuyuna Ranges and continued mining of taconite on the Mesabi Range have resulted in excellent artificial exposures and an abundance of drill hole information. Geophysical surveys and stratigraphic test drilling by the Minnesota Geological Survey also have been major sources of information (e.g., Southwick et al., 1988).

Animikie Group

The Animikie Group unconformably overlies the Mille Lacs and North Range Groups to the south and the Archean basement to the north (Southwick and Morey, 1991). Magnetic data show North Range structures are present beneath Animikie strata to the east of the exposed North Range Group (Chandler, 1993).

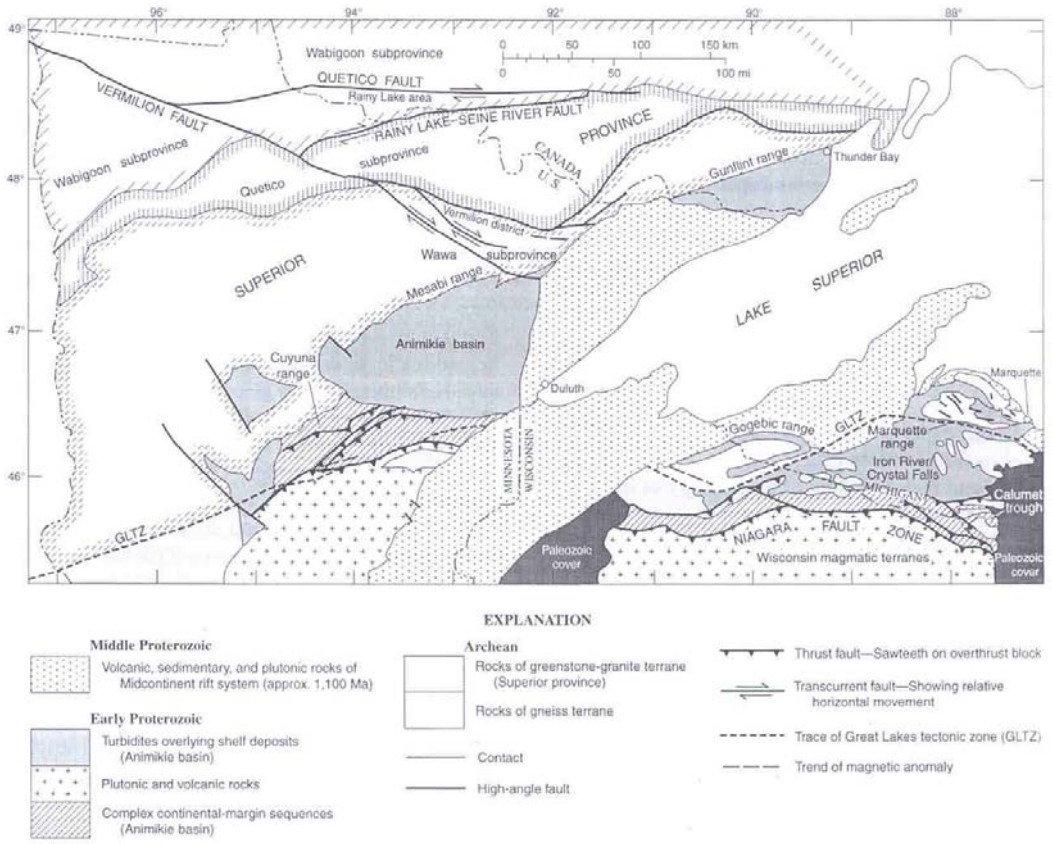


Figure 4-2: Generalized geologic map showing the distribution of Precambrian rocks and structural elements of the Lake Superior region, modified from Ojakangas 1994 and references therein (from Ojakangas, et al., 2001).

The group consists of three conformable major formations on both the Mesabi and Gunflint ranges. The respective units on the two ranges 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 correlative with the Virginia and Rove Formations. The Biwabik and Gunflint are on strike with each other and were probably continuous prior to the intrusion of the Duluth Complex at about 1100 Ma.

The Animikie Group in Minnesota-Ontario on the Mesabi and Gunflint ranges and the Baraga Group of Michigan-Wisconsin on the Gogebic range were both deposited in the Animikie foreland basin. The basal units comprised of siliciclastic sediment derived from the Archean basement, and the overlying iron-formation, were deposited in a shallow sea on the northern edge (i.e., the peripheral bulge or foreland) of the northward-migrating Animikie basin (e.g., Ojakangas, 1994). Additional details are provided below in the section titled, Environments of Deposition.

The siliciclastic and iron-formation units are exposed on the Gogebic Range of northern Michigan and Wisconsin (Palms Quartzite and Ironwood Iron Formation), on the Mesabi Range of northern Minnesota (the Pokegama Formation and the Biwabik Iron Formation), and on the Gunflint range of northeasternmost Minnesota and Ontario (the Kakabeka Quartzite and the

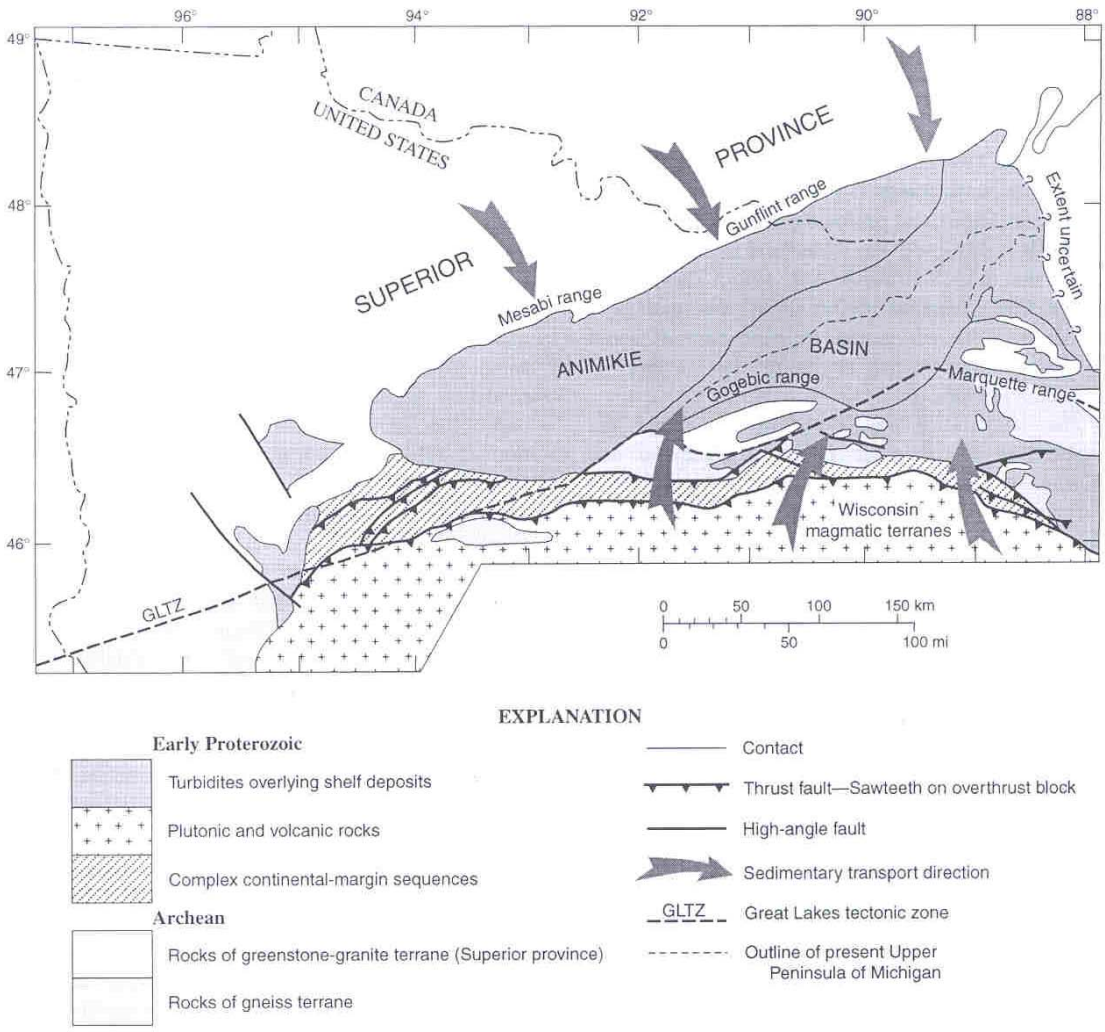


Figure 4-3: Schematic hypothesized paleogeography at the time of sedimentation of the Animikie Group turbidites that overlie shelf deposits in the Animikie basin. The rocks of the 1.1 Ga Midcontinent Rift System have been removed from the map, and Michigan and Wisconsin are thus positioned 60 miles closer to Minnesota-Ontario than they were after the formation of the Midcontinent Rift System. Arrows denote generalized transportation directions of sediment from major source areas. Compare with Fig. 4-2. Modified from Ojakangas (1994) and references included therein (from Ojakangas, et al., 2001).

Gunflint Iron Formation), and are lithostratigraphic equivalents. They probably were continuous from south to north prior to development of the Midcontinent Rift System in Mesoproterozoic time. A consequence of this model is that they are diachronous, with the units in Michigan-Wisconsin (located about 60 miles to the south of the Mesabi range during deposition) thus somewhat older than those in Minnesota-Ontario. The thickest and uppermost units in the basin, essentially lithostratigraphic correlatives but probably differing somewhat in age, are the Michigamme, Tyler, and Copps Formations of the southeastern segment and the Thomson, Virginia and Rove Formations of the northwestern segment. These are typical turbidite-shale (flysch) sequences, with graded beds and intercalated muddy “rain-out” sediment.

Figure 4-5 is a regional correlation chart of the aforementioned units, as well as others that cannot be discussed herein.

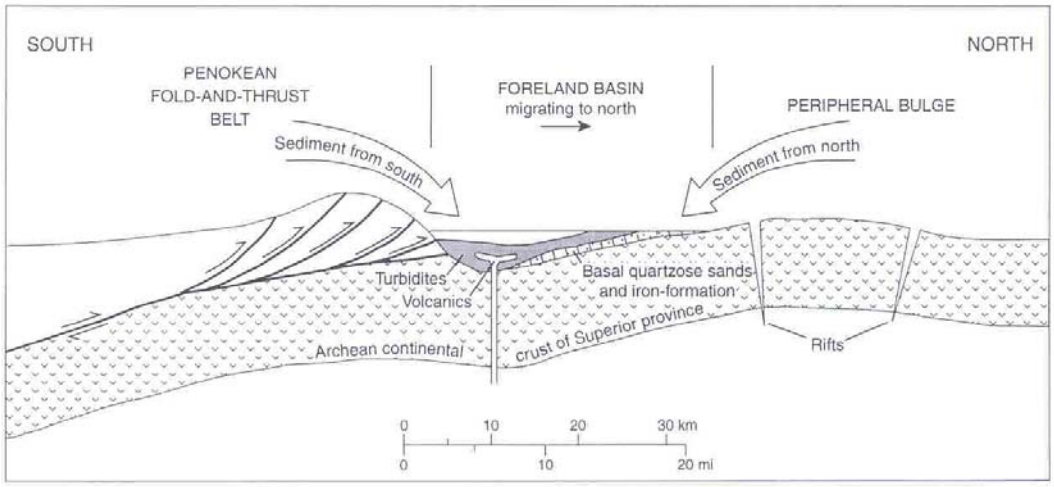


Figure 4-4: Schematic cross-section depicting deposition of the Animikie Group turbidites that overlie shelf deposits in the Animikie basin, with sediment derived from both the north south. The southern area, the fold-and-thrust belt, comprises a complex assemblage including: 1. accreted Paleoproterozoic volcanic and plutonic rocks and volcanic rocks of the Wisconsin magmatic terranes; 2. accreted Archean miniplate terranes; 3. older Paleoproterozoic passive-margin sedimentary rocks and volcanic rocks produced during initial rifting of the continental margin, both scraped off the southward-subducting Archean Superior craton; and 4. recycled initial foredeep deposits, possibly including basal shallow water sandstones deposited in the transgressing sea of the northward-migrating foreland basin. The peripheral bulge comprises a source-rock assemblage of Archan granitic rocks and Archean volcanic-sedimentary (greenstone) belts. Scale is approximate. Compare with Fig. 4-3. Modified from Ojakangas (1994) and references included therein (from Ojakangas, et al., 2001).

Ages

Along the Mesabi Range, the Pokegama Formation rests unconformably on diabase dikes of the Kenora-Kabetogama dike swarm that gives a Rb-Sr isochron age of 2125 +/- 45 Ma (Southwick and Day, 1983; Beck, 1988) and this provides a maximum age for deposition of the Pokegama. A minimum age of 1930 +/- 25 Ma (Pb/Pb) for the Pokegama was obtained by Hemming et al. (1990) from quartz veins that cut the Pokegama. A U/Pb age on euhedral zircons from an ash layer in the lower Gunflint Iron Formation of Ontario is 1878 +/- 2 Ma (Fralich and Kissin, 1998; Fralick et al., 2002). A similar age of 1874 +/- 9 Ma was obtained on zircon from rhyolite in the Hemlock Formation that is adjacent to (and is interlayered with?) the Negaunee Iron Formation in the Marquette Range Supergroup of Michigan (Schneider et al., 2002). A zircon age from an ash layer near the base of the Virginia Formation is 1850 Ma (Hemming et al., 1996), and an age of 1821 +/-16 Ma has been obtained from an ash layer in the Rove Formation about 70 m above the Gunflint Iron Formation (Kissen et al., 2003). Several of these ages are shown on Figure 4-5.

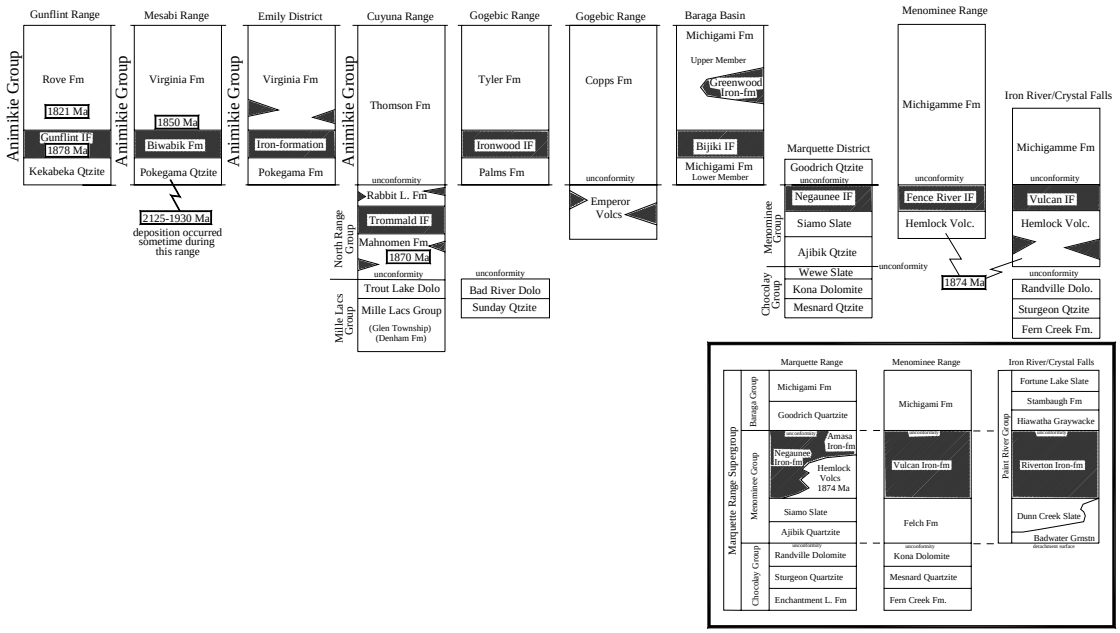


Figure 4-5: Generalized correlation chart of Paleoproterozoic strata in the Lake Superior region (after Morey and Southwick, 1995). Note that recently obtained age dates are shown for the Gunflint Formation, Mahnomen Formation, Hemlock Volcanics, and the Rove and Virginia Formations (see text for references). Also included is a correlation chart (lower right corner) of strata in Menominee, Iron River-Crystal Falls and surrounding terranes (LaBerge et al., 2003 – includes changes to previous usage not yet officially adopted by the USGS). Iron-formations are shaded. Modified from Severson et al., 2003.

Pokegama Formation

The formation has long been called the Pokegama Quartzite, but because it contains appreciable argillite and siltstone, the name Pokegama Formation is more appropriate. It has been studied by several workers since it was named by Winchell (1893) for exposures at the western end of the Mesabi Range. Much of the previous work has been summarized by Morey (1972, 1973, 2003).

Few natural exposures exist, as thick glacial drift generally covers the formation. Outcrops, roadcuts, and mine cuts occur at a few places along the length of the range, but most exposures are in the central portion of the range. A few drill holes have penetrated the entire formation. One is located just south of Eveleth (NE 1/4, NE1/4, Section 5, T. 57 N., R. 17 W.) and another is southwest of Mountain Iron (SE 1/4, SE 1/4, Section 8, T. 58 N., R. 18 W.); the thicknesses are 167 ft and 85 ft, respectively (Fig. 4-6). Three other drill cores, recently rediscovered, have not yet been studied. Numerous drill holes have penetrated only the upper few feet of the formation, as the drilling was generally undertaken in relation to iron ore exploration and development. The Pokegama is thin at the eastern end of the range and thickens to the western end where it may be more than 300 ft thick.

The formation is composed of three main rock types—argillite, siltstone, and quartzite. The latter is generally a silica-cemented quartz sandstone, and is therefore an orthoquartzite rather than a metaquartzite. (Morey, 2003, has determined that mineralogical changes in the Pokegama and the Biwabik are the result of diagenesis rather than metamorphism, except at the eastern end of the range adjacent to the Duluth Complex.) These three rock types make up three gradational

members-- lower, middle, and upper-- respectively, as shown in Figure 4-6. Minor thin conglomerates occur at the base of the formation, and seem to represent a weathered residuum on the surface of Archean rocks, perhaps reworked by fluvial processes.

The Pokegama unconformably overlies Archean metavolcanic, metasedimentary, and plutonic rocks. There may be as much as 100 ft of relief on the Archean surface (Grout and Broderick, 1919), but the surface was, nevertheless, essentially a peneplain. Some Archean "knobs" were islands when the Pokegama was being deposited, and are present in the wooded areas between Eveleth and Virginia. The Pokegama-Biwabik contact is gradational, with some cherty horizons in the upper Pokegama. Various geologists have placed the contact at different stratigraphic levels.

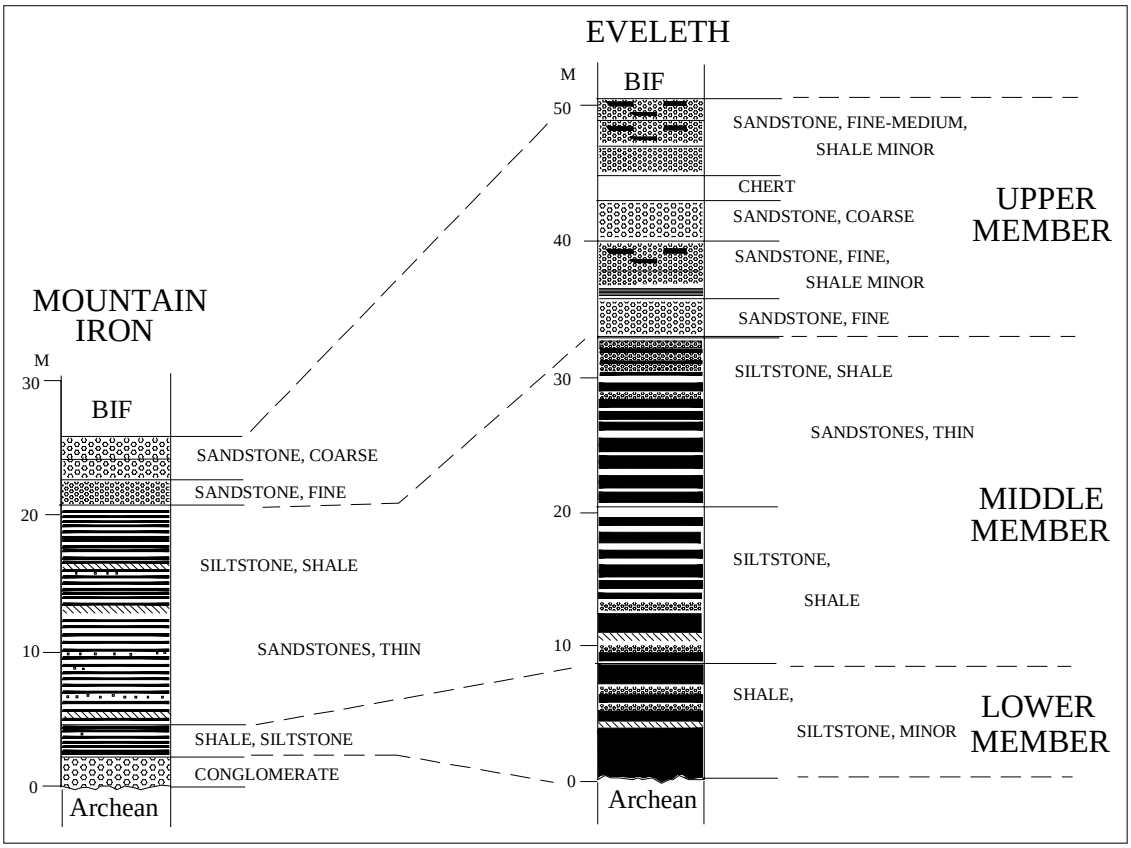


Figure 4-6: Measured sections from two drill holes that penetrate the entire Pokegama Formation. Dark shading represents shale, thin blank units represent siltstone, and dotted pattern represents sandstone. BIF is Biwabik Iron Formation. After Ojakangas (1983).

Biwabik Iron Formation

This is one of the world's major iron-formations, and the largest in the United States. The formation is 200 to 750 ft thick and consists of four divisions as defined by Wolff (1917). These lithostratigraphic units, now informal members, are from the bottom up, the Lower Cherty, the Lower Slaty, the Upper Cherty, and the Upper Slaty. (These are miners' terms, and do not indicate metamorphism.) The cherty members are dominantly granular (i.e., sand-textured), thick-bedded (several inches to several feet), and are largely composed of chert and iron oxides. The slaty members are dominantly fine-grained (i.e., mud-textured), thin-bedded (<1 inch) and

composed mostly of iron silicate and iron carbonate with local chert beds. However, these two rock types are interbedded on all scales and are generally gradational. They contain about the same high quantities of silica, 42-47% (Morey, 1992). The Lower Slaty is not present at the far western end of the range.

There are some diagnostic marker units within the formation. Two stromatolite-bearing intervals several feet thick are present, one at the base of the Lower Cherty and the other in the middle of the Upper Cherty. The black "intermediate slate" at the base of the Lower Slaty is reportedly an ash-fall tuff containing about 4 to 5.5 % aluminum oxide (Morey, 1992). At the top of the Upper Slaty are several feet of limestone and dolomite. Most of these marker units pinch out to zero in the vicinity of Nashwauk, about 40 mi from the west end of the range (Morey, 1992).

Virginia Formation

There are rare exposures of the Virginia in mines at the east end of the Mesabi Range where it has been metamorphosed by the mafic intrusions of the Mesoproterozoic Duluth Complex. Several holes drilled south of the range to study the underlying iron-formation have been drilled through the Pleistocene cover and have intersected as much as 1443-ft of the preserved lower part of the formation (Lucente and Morey, 1983).

The lower portion of the formation in the drill holes is dominantly black shale. The upper portion of the drill core, while still dominantly shale, contains beds of siltstone and fine-grained feldspathic graywacke comprising thickening- and coarsening-upward turbidite sequences. Ash-fall tuffs, cherty sideritic iron-formation, chert, and limestone are minor rock types low in the formation. The contact with the underlying Biwabik Iron Formation is gradational. The clastic rocks were largely derived from the Archean rocks to the north, with some contributions from lower Proterozoic rocks to the south (Lucente and Morey, 1983).

The Virginia Formation is correlated with the Thomson Formation (Morey and Ojakangas, 1970) that is exposed 60 miles to the south in the vicinity of Carlton and Cloquet, and also with the Rove Formation in northeasternmost Minnesota and adjacent Ontario (Morey, 1967).

Environments of Deposition, Animikie Group

The Pokegama is interpreted to have been deposited in a tidally influenced shallow marine setting near the shoreline, having received clastics from the Archean basement to the north (Ojakangas, 1983). In this model of a transgressing sea, the lower (argillaceous) member was deposited at the shoreline in the upper tidal flat, the middle member of intercalated argillaceous and silty sediment was deposited seaward in the middle tidal flat, and the upper member of quartz sand was deposited still further seaward in a lower tidal flat/subtidal environment. This is illustrated in Figure 4-7. Walther's Law is applicable here, with the vertical facies showing the relationships of the lateral facies.

The lowermost Pokegama contains siltstone beds that contain alternating thicker and thinner laminae that have been interpreted as evidence of the diurnal inequality, and are being investigated further for possible clues to the Paleoproterozoic lunar orbit (G. Ojakangas, 1996).

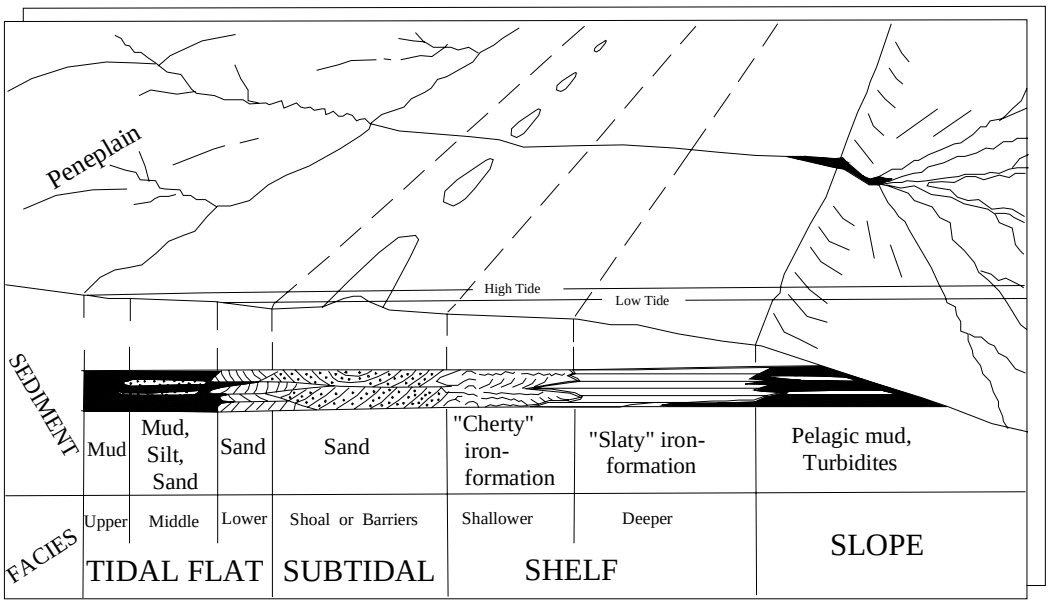


Figure 4-7: Sedimentation model showing lateral relationships of the siliciclastic tidal facies of the Pokegama Formation, the two main facies of the Biwabik Iron Formation, and the Virginia Formation (on the slope?). Thicknesses and geography not to scale. From Ojakangas (1983).

The Biwabik is interpreted to have been deposited seaward of the Pokegama on a shallow marine, tidally dominated shelf (Figure 4-7). Precipitation of iron minerals including iron carbonate, iron silicate, silica, and perhaps some hematite, occurred on the outer shelf in waters below wave base, giving rise to the mud-textured (slaty) iron-formation. These minerals were likely related to upwelling waters from the deeper part of the basin.

The two sand-textured members (Lower Cherty and Upper Cherty) formed in a shallow-water, high-energy environment, as indicated by stromatolites, cross-bedding, and rounded (locally oolitic) grains of iron minerals and chert. Shoreward-moving tidal currents (i.e., flood tides) and/or storms may have disrupted the mud-textured sediment (i.e., precipitates) and transported sand-sized aggregates into shallower water where they were altered by seafloor processes and early diagenetic processes. Thus these granules are “intraclasts” derived from within the basin.

Shallow channels up to a mile wide and tens of feet deep were cut into the Lower Slaty member and filled with sand-textured grains of iron minerals and chert at Minntac where this is called the IBC unit (interbedded chert unit). These grains apparently were derived from shallow water and carried seaward into the deeper water environment in which the iron minerals were precipitating. Ebb-flow tidal currents are interpreted as the erosion and transportation agent.

A plot of 102 cross-bed measurements in the Minorca Mine on the northeast edge of the Virginia Horn (Fig. 4-2) shows 90 % of the readings making a very prominent mode to the north-northeast and a minor broader mode to the south (Fig. 4-8). This distribution is interpreted as the product of a strong flood tide toward the paleogeographically determined northern shoreline and a much weaker ebb tide.

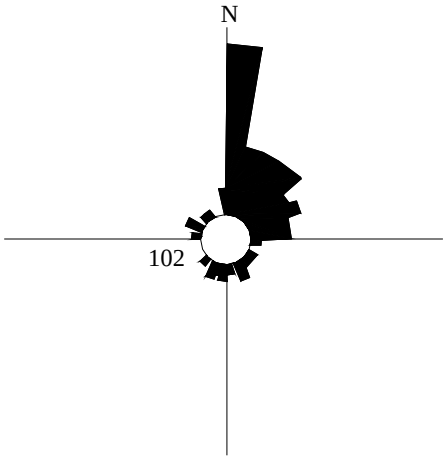


Figure 4-8: Paleocurrent rose diagram of 102 cross-bedded measurements from the Lower Cherty in the Minorca Mine.

A study of the orientations of stromatolite mounds in the stromatolite horizon within the Upper cherty member was conducted by Kevin Boerst (1999) as a UROP (Undergraduate Research Opportunity Project) at the University of Minnesota Duluth. His map is presented here as Fig. 4-9). A paleocurrent plot of mound elongation (Fig. 4- 9) is interpreted as the result of shore-normal tidal currents and shore-parallel longshore currents in shallow water.

The repetition of the cherty and slaty members has long been interpreted as the result of transgression and regression (White, 1954).

The Virginia Formation was deposited seaward of the iron-formation, probably in a slope-type environment (Fig. 4-7) where episodic turbidity currents deposited graded beds. Some volcanic ash falls evidently settled into the basin forming graded beds with a totally volcanic composition. The dominance of black, fissile shale suggests the “raining out” of clay (i.e., settling through the water column) and deposition in deep, anoxic water below wave base. Minor thin sandstone lenses were deposited by bottom currents (Lucente and Morey, 1983).

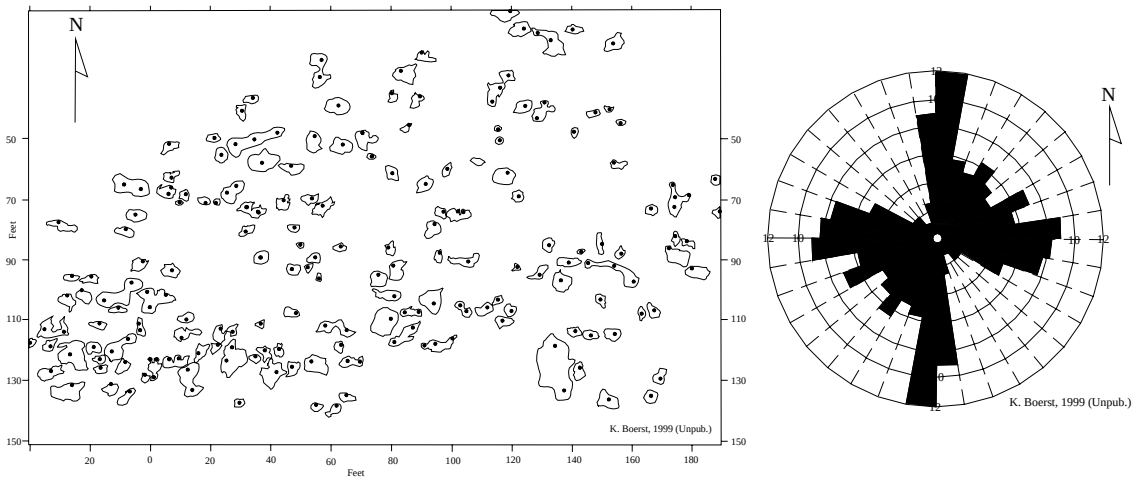


Figure 4-9: Mapped stromatolite mounds in the Algal submember (I submember) in the Upper Cherty of the LTV 2E pit. The rose diagram represents the elongation of the mounds. From unpublished work by Kevin Boerst (1999).

Mineralogy

The detailed origins of the iron minerals are exceedingly complex and are beyond the scope of this introduction. It has to suffice here to say that Eh and pH are major controls on the stability of the various iron minerals in both the depositional and diagenetic environments, and in the easternmost Mesabi, in the metamorphic environment as well. Recrystallization and replacement of the granules during diagenesis has been extensive, and probably consisted of a number of discrete events.

Earlier work on the oxidized taconites of the western Mesabi 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.

Some of the hematite inclusions and crystals in magnetite are similar to those illustrated by Han (1982). He proposed that much of the magnetite formed by the replacement of, and overgrowth on, pre-existing hematite that served as nuclei. Han further suggested that ionic diffusion of ferrous iron was a key process in the formation of the magnetite. Organic carbon may have acted as a reductant in this process.

The nature of the major hydrologic events that removed the 40-60% of the silica and oxidized the iron minerals, thus forming the high-grade (natural) ore bodies, has long been debated. Were they descending cool meteoric waters or ascending hydrothermal waters related to igneous activity? Did this occur during the Cretaceous (the age of conglomerates composed of clasts of high-grade hematite), or prior to that time? Morey (1999) provided an excellent review of the arguments. He then proposed that a large-scale topography-driven hydrothermal groundwater system moved waters northward through the sands of the underlying Pokegama Formation, from the vicinity of the regional Penokean orogenic uplift in northern Wisconsin and east-central Minnesota, 40 to 80 miles to the south.

Production Figures – Iron Ore and Taconite

The annual amount of direct shipped and taconite produced are shown in Figure 4-10. Production and shipping of direct ore started in 1892 and rose steadily until 1953 when a maximum 76 million tons were produced in one year (note the precipitous drop in direct ore production corresponding to the Depression). At around 1955, there was a dramatic decrease in the amount of direct ore as the various mines became depleted. This also corresponds to the initial start-up of taconite mining, using a concentrating and pelletizing method developed by E.W. Davis of the University of Minnesota. Reserve Mining opened the first taconite operations in 1955 (Peter Mitchell Mine) and was shortly followed by Erie Mining in 1957 (old LTV site). Six more taconite operations were added in the 1960s, and by 1967, annual taconite production exceeded direct ore production. The mid-1980s marked a serious depression in the iron ore and steel industry that resulted in the closure of one operation (Butler Taconite) and the bankruptcies of two other taconite producers. More recently, LTV Steel and Eveleth Taconite have closed; Evtac has since reopened as United Taconite.

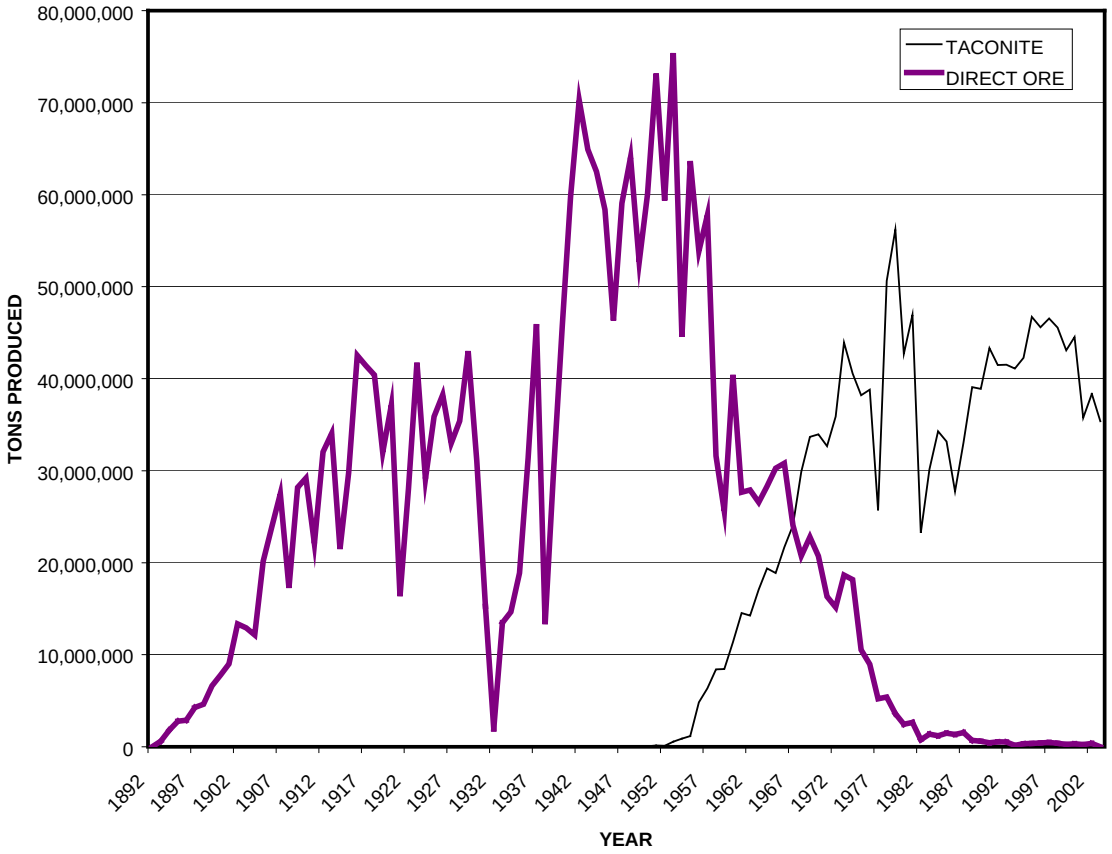


Figure 4-10: Annual production figures for direct ore (includes all forms of direct ore) and taconite for the period 1892-2003. Data and graph from James Sellner, Minnesota Department of Natural Resources, Lands and Minerals Division, Hibbing, MN.

What's in a Name? (Those confusing iron-formation submembers!)

As early as 1917, the Biwabik Iron Formation (BIF) has been informally broken down into four major lithostratigraphic members, or subdivisions, known as (from bottom to top): Lower Cherty, Lower Slaty, Upper Cherty, and Upper Slaty (Wolff, 1917). The cherty iron-formation members are generally thick-bedded and contain round grains (0.5-2.0 mm) of chert that are referred to as granules. These "cherty" members typically contain higher percentages of iron oxides (magnetite, hematite and/or goethite). In contrast, the "slaty" members are thin-bedded (0.5-3.0 mm thick beds) and very fine-grained. They are composed mostly of Fe-silicates and Fe-carbonates. Both cherty and slaty iron-formation types are interlayered at all scales. However, one rock type often predominates in each of the four lithostratigraphic members, and are so-named due to this dominance, i.e., thick-bedded cherty iron-formation is dominant in the cherty members, whereas thin-bedded iron-formation is dominant in the slaty members.

The 1917 four-fold stratigraphy of a Lower and Upper Cherty and a Lower and Upper Slaty members is still used at each of currently operating (and inactive) taconite mines on the Mesabi Range. However, each of the mining companies further subdivides the BIF into several

submembers based on bedding types (Fig. 4-11) and mineral assemblages. It is at this point that the BIF stratigraphy becomes very complicated and at times confusing. This is mainly due to the following reasons:

- There are localized lateral facies changes between mines (and even within a single mine). Some mines reconcile these differences by splitting out numerous submembers (each with a distinct bedding type, texture, ore grade, and/or mineral assemblage), whereas, other mines lump many of these same differences within a single submember.
- There are significant lateral facies changes over several miles between mines. For example a particular horizon may be massive-bedded at one location but is regular-bedded a few miles away. This is particularly troublesome within the Upper Cherty member in the western 2/3rds of the Mesabi Range.
- Not all mines use the same numbering system – some use abbreviations (LC for Lower Cherty, etc.) followed by a number (as in LC-5 at the top of the Lower Cherty). However, other mines use an alphabet system, devised by Gundersen and Schwartz (1962), starting with the A submember at the top of the Upper Slaty (in this system the top of the Lower Cherty corresponds to the R submember). And further still, another mine refers to the Lower Cherty as the number 1 unit and subdivides it into eight submembers, with 1-8 at the top of the Lower Cherty.
- Some mines label downward in their numbering system, whereas other mines label upward in their numbering system.

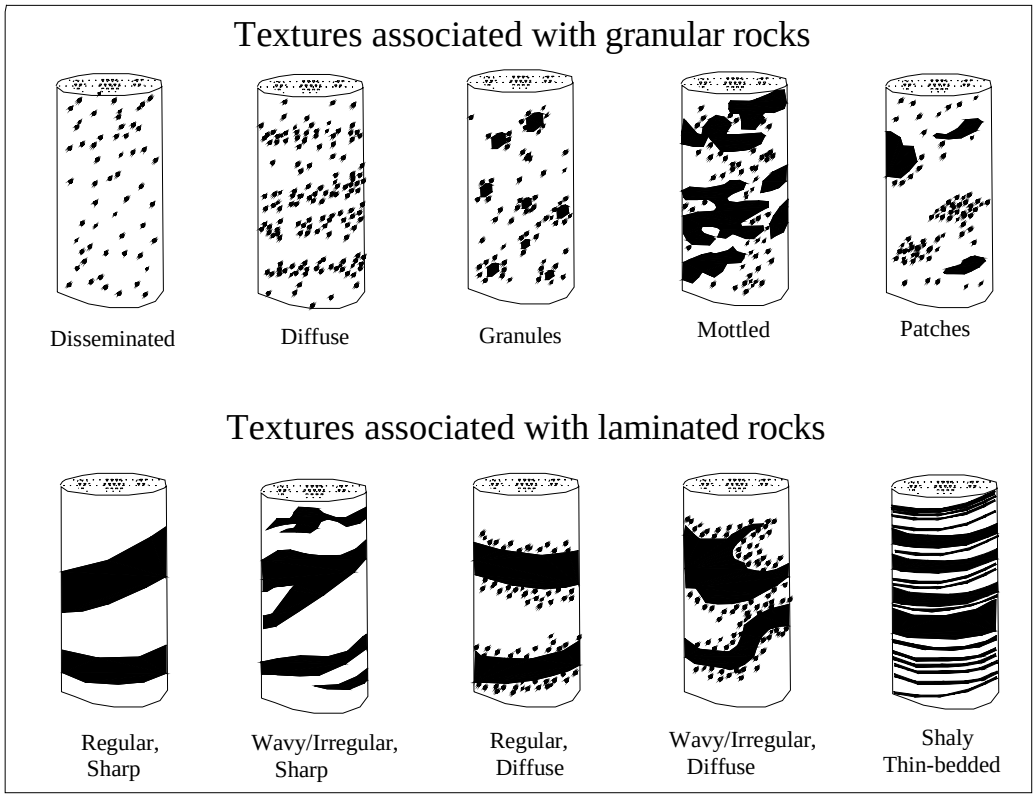


Figure 4-11: Textural characteristics of the Biwabik Iron Formation (modified from Pfleider et al., 1968, from a classification scheme developed by geologists of the Hanna Mining Company).

The submember nomenclature that is used at each of the mines is summarized in Figure 4-12. It can readily be seen on this summary that any particular submember name changes nomenclature from one mine to the next. This is because there are few good marker horizons within the BIF, and even these can exhibit gradual lateral facies changes or pinch-and-swell relationships to each other. A few of the potential marker horizons within the BIF are presented below.

- Top contact of the BIF with the Virginia Formation – In the eastern half of the Mesabi Range a carbonate horizon is present at the very top of the Upper Slaty and the contact between the BIF and Virginia Formation is easily recognized (Gruner, 1924). However to the west of Hibbing, the carbonate layer is absent and lenses of thin-bedded Fe-carbonate iron-formation are present in the Virginia Formation, and the top of the BIF is not easily discerned.
- Submember I/Algal unit – a thin unit containing algal stromatolites and jasper-bearing intraformational conglomerate is present near the top of the Upper Cherty. This submember is easily recognized but is not present to the west of Hibbing.
- Lower Slaty member – The Lower Slaty has a very-well defined "Intermediate Slate" (also referred to as the Q submember or Paint Rock in Fig. 4-12) at its base which is characterized by a black, carbon-rich, thin-bedded, slaty unit that commonly contains pyrite. This unit is readily evident at all of mines on the range. However, the upper contact of the Lower Slaty is "indefinite and a gradual change to other slaty phases takes place" … [which makes] … the dividing line between the two [Upper Cherty and Lower Slaty] somewhat arbitrary" (Gruner, 1924, p.20). The upper contact of the Lower Slaty is particularly troublesome in the Virginia Horn area. Gruner (1946, p. 45) included lenses of cherty and wavy-bedded taconite (referred to as the Interbedded Chert -- IBC unit at Minntac; Fig. 4-12) in the Lower Slaty, whereas White (1954) included these same units in the overlying Upper Cherty member.
- Base of the BIF - The base of the Lower Cherty is generally characterized by thin-bedded iron-formation (also called the "red basal unit") with localized algal stromatolite and basal conglomerate horizons. However, at many localities the base of the BIF exhibits a gradational contact with the underlying Pokegama Formation. In the Virginia Horn area the base of the BIF contains an iron-bearing sandstone (White, 1954) that some mines include with the iron-formation, whereas others lump this type of material with the Pokegama Formation.

From the above description it is readily evident that there are few good marker horizons within the iron-formation; even the upper and lower contacts of the iron-formation are gradational and subject to various interpretations. The "Intermediate Slate" and the algal horizon in the Upper Cherty are the only easily recognizable marker units. However, even using these horizons as markers, one can see from Fig.4-12 that there are problems.

Clearly, much additional work needs to be done in understanding how submembers at one mine correlate with submembers at an adjacent mine. These types of studies could inevitably be important in determining why ore grades, and waste rock characteristics, change between mines and even within a single mine. For example, some of the best ore-grade taconite corresponds to the wavy-bedded or irregular-bedded taconite present in both the Lower or Upper Cherty members. The sedimentary environment that produced this type of taconite ore has not been fully documented nor have any recent detailed sedimentological studies been attempted. A better understanding of the various sedimentological textures in the BIF could ultimately lead to an increased ability to better predict changes in ore grades as they relate to facies changes.

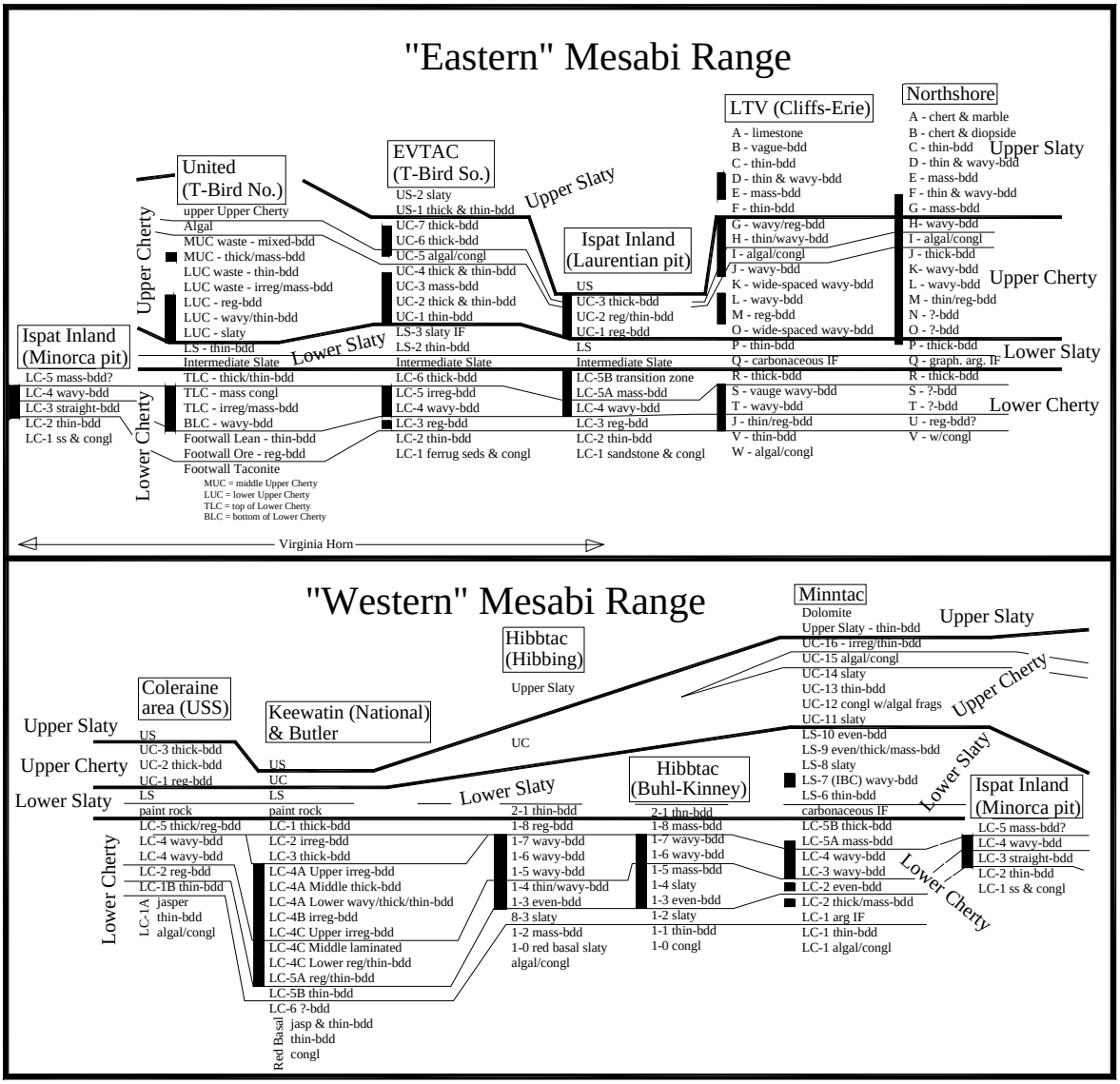


Figure 4-12: Correlation chart of submembers, at each of the mines/areas within the Biwabik Iron Formation, as deciphered from published descriptions and mine handouts (modified from Zanko et al., 2001). All columns are hung on the base of the Lower Slaty ("Intermediate Slate"). It is important to note that this summary is preliminary as it has not been field-checked. No scale is implied and the true thickness of each submember is not portrayed. Bars to the left of the columns indicate mined taconite ore zones. Note that there are several consistent submembers within the Lower Cherty as opposed to very few laterally persistent submembers in the Upper Cherty.

STOP DESCRIPTIONS

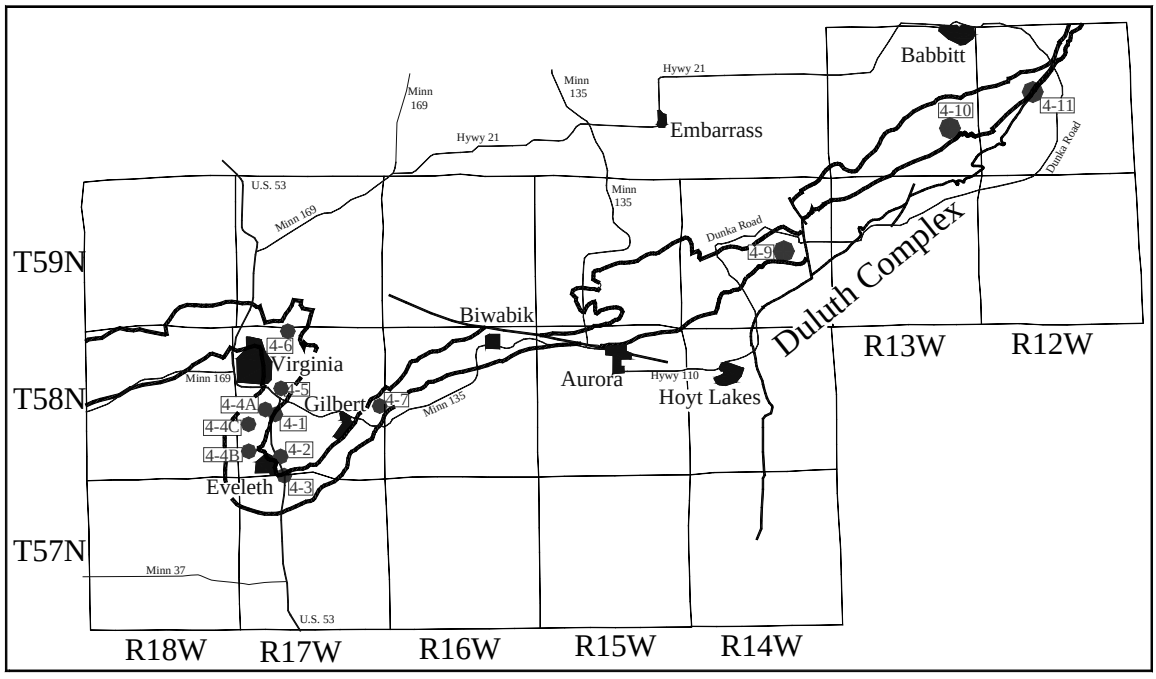


Figure 4-13: Locations of stops that will be visited on the field trip. Generalized contacts of the Biwabik Iron Formation, for the eastern half of the Mesabi Range, and Duluth Complex are shown. Note that in order to observe Stops 4-1, 4-2, and 4-3 in stratigraphic order, we will be back-tracking for short distances.

STOP 4-1A: Pokegama/Archean Unconformity

Location: On U.S. 53 in Eveleth, there is a stoplight at Grant/Industrial Avenue. Proceed northward past the stoplight for about 0.4 mi north to a gas station. Turn right and immediately turn left on the frontage road (Midway Drive). Drive past a church and past the first street on the right (Mesabi Drive). Watch for a low, small outcrop in the trees on the right, just a few feet off the road. Eveleth 7.5' quadrangle, T.58N., R.17W, Sec 20, NW of SW of SE; 535445E, 5259520N (NAD83 UTMs)

Description: This is the only easily accessible exposure of the unconformity. A close examination will reveal the presence of a thin smear of Pokegama conglomerate composed of schist and vein quartz clasts upon Archean schist with a near-vertical foliation. Note that the foliated schist clasts are flat and have a subhorizontal orientation. NO HAMMERING, PLEASE!

STOP 4-1B: Argillaceous lower member of the Pokegama Formation

Location: From stop 4-1A, CAREFULLY walk across both lanes of U.S. 53 to the long, roadcut on west side of highway. Eveleth 7.5' quadrangle, T.58N., R.17W, Sec 20, NW of SW of SE; 535380E, 5259555N (NAD83 UTMs)

Description: This cut is about 500 feet long and 5-10 ft high, and is the only exposure of this member. It consists largely of shale and siltstone with minor fine-grained sandstone. It has been interpreted as having been deposited in a low-energy upper tidal flat environment in a sea that transgressed onto the peneplaned surface of Archean rocks (Ojakangas, 1983). Minor channeling is common at the bases of the thicker sandstone beds, and at one spot, 0.5 m of section has been eroded. Small-scale cross-bedding is present in some siltstone beds, and elongated sole marks are visible on the bottoms of some sandstone and siltstone beds. A total of 57 of these paleocurrent indicators show that the currents were generally oriented in a north-south direction. A few concretions as large as 6 inches in diameter are present, as is soft-sediment deformation. Hemming et al. (1991) illustrated the soft-sediment folding and interpreted it as evidence that the Animikie basin was tectonically active during deposition of the Pokegama and the Biwabik. Alternatively, it is interpreted herein as soft-sediment slumping into tidal channels. Fine laminations and sequences of laminations in the sandstone beds have been interpreted as tidal rhythmites (G. Ojakangas, 1996).

STOP 4-1C: Jaspillite on Archean metaconglomerate

Location: From stop 4-1A, drive a long block northward to Merritt Drive. Turn right. About 100 ft farther, turn right on Mesabi Avenue. About another 100 ft farther, there is a fork in the street. Park on Mesabi Avenue and walk uphill on the street on the left to a broad, flat rock exposure in a driveway. Eveleth 7.5' quadrangle, T.58N., R.17W., Sec 20, SW of SE (No. 7 Mesabi Lane); 535610E, 5259460N (NAD83 UTMs)

Description: This is an excellent exposure of jaspillite resting unconformably on subvertical volcanogenic metaconglomerate. The outcrop is an estimated 30 to 40 ft topographically higher than stop 4-1A, and was also that much higher when the jaspillite was deposited. Is it an erosional remnant of basal Biwabik Iron Formation or is it a local chemical precipitate at the base of the Pokegama Formation? NO HAMMERING, PLEASE!

NOTE: The middle member of the Pokegama Formation (a unit of intercalated beds of sandstone, siltstone, and argillite) was once poorly exposed in the flat area across the highway from the U.S. Hockey Hall of Fame. However, the best part of the poor exposure has since been covered by a frontage road.

STOP 4-2: Upper member of the Pokegama Formation

Location: Continue driving down Mesabi Drive to the frontage road (Midway Drive). Turn right and drive past stop 4-1A to a stop sign. Turn left to intersection with U.S. 53. Cross the northbound lane and turn left on highway. Drive south past stop 4-1B, past the Grant/Industrial Avenue stoplight, past the U.S. Hockey Hall of Fame on the right, and stop at the Rustic Rock Inn (you are now opposite a large roadcut on the east side of the highway). Walk across the highway, CAREFULLY! Eveleth 7.5' quadrangle, T.58N., R.17W, Sec 32, E ½ of SE; 536055E, 5256775N (NAD83 UTMs)

Description: This is the upper member of the Pokegama Formation, composed of silica-cemented quartz sandstone. It is the rock type found immediately beneath the Biwabik Iron Formation; it was penetrated by countless drill holes during mining and exploration, and resulted in the formation being originally named the Pokegama Quartzite. Note the massive beds

separated by thin beds of shale or siltstone. Silica cementation likely obscured some original cross-bedding. This member was interpreted by Ojakangas (1983) as having been deposited in a high-energy, lower tidal or subtidal environment.

STOP 4-3: <u>Lower Cherty member, Biwabik Iron Formation</u>

Location: Proceed south on U.S. 53 for 0.6 miles (passing over an overpass and past a long roadcut of iron-formation on the right), to a left-turn lane. Make a U-turn and proceed north on the highway for about 0.2 miles to roadcut on the off-ramp for Highway 37 (right side of highway). Eveleth 7.5' quadrangle, T.57N., R.17W, Sec 5, E ½ of NE of NE; 536230E, 5256285N (NAD83 UTMs)

Description: Note that this member of the Biwabik Iron Formation overlies the sandy member of the Pokegama Formation of the last stop, and that both units dip gently to the southeast. Observe the thick wavy bedding, the trough cross-bedding, and the sandy texture of the iron minerals and chert grains. The cross-beds are best observed in this eastern roadcut, but are also present at both ends of the longer cut on the other side of the highway. Cross-bedding measurements (Fig. 4-16), although not definitive, are suggestive of a tidally-influenced marine environment. This, coupled with Walther's law of succession of sedimentary facies (i.e., the facies observed vertically are also similarly related laterally), places the deposition of the iron-formation seaward of the Pokegama Formation.

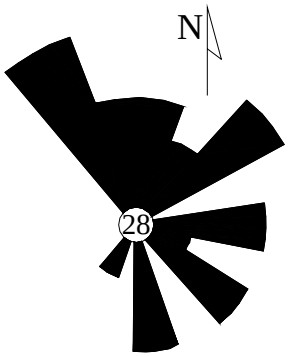


Figure 4-14: Rose diagram of 28 cross-bedded measurements, mostly from the roadcut of Stop 4-3.

Stops 4-4A through 4-4C are at the Thunderbird North Mine of United Taconite (formerly Eveleth Taconite, or Evtac). Simplified stratigraphic submembers of the iron-formation, according to terminology used by Evtac, are portrayed in Figure 4-15. From stop 4-3, proceed north along U.S. 53 past stop 4-2 to the stoplight at Grant/Industrial Avenue. Turn left on Grant Avenue and follow the road past a cemetery to the entrance road to United Taconite (U-Tac).

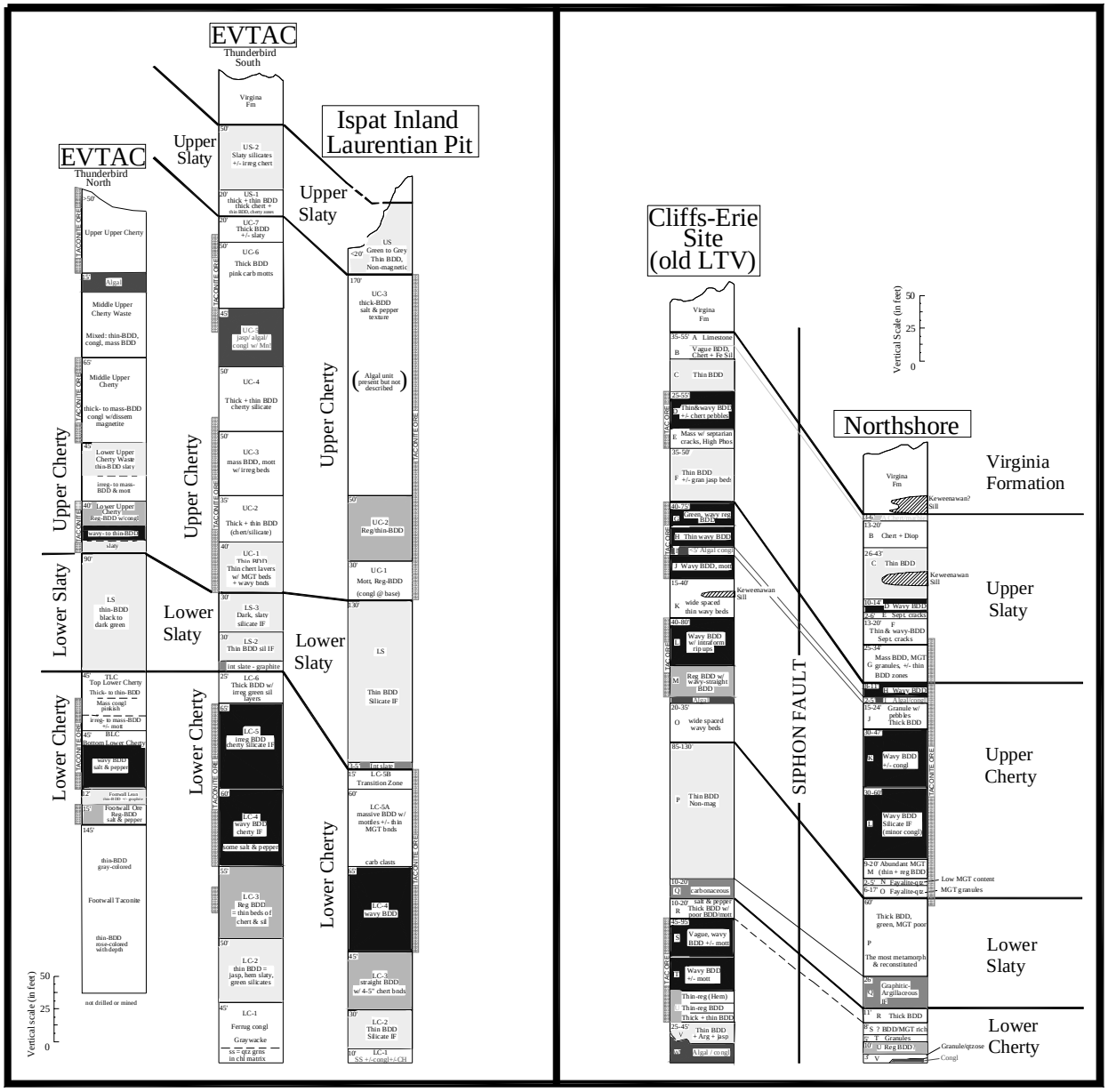


Figure 4-15: Submember nomenclature for the Biwabik Iron Formation as used at the various taconite mines and idled mine sites that will be visited on this field trip. Modified from Plate II in Zanko et al., 2001. Note that the nomenclature portrayed here is preliminary in nature and future correlations of the various submembers used at the different mines still needs field checking and modification. The stratigraphic nomenclature is currently being revised for the Thunderbird North Mine by United Taconite, but was unavailable at press time and is not included in this figure.

STOP 4-4A: Submember LC-3 near the base of the Lower Cherty

Location: Northeast end of the Thunderbird North Mine. Eveleth 7.5' quadrangle T.58N., R.17W., Sec 20 , 535290E, 5260255N (NAD 83)

Description: The LC-3 submember (new U-Tac terminology or "footwall ore" in the old Evtac terminology) is characterized by regular-bedded and wavy-bedded cherty taconite wherein most of the magnetite is contained within dark-colored wavy (irregular) beds that locally show good pinch-and-swell relationships. This is the lowest unit mined at the Thunderbird North Mine and contains approximately 20-23% mag-Fe and 2% Davis Tube Si.

The Auburn Fault can also be viewed at this locality. The fault zone consists of a 20-foot-wide zone of broken-up and highly oxidized vuggy rock. To the northwest is the Auburn Mine that was originally developed as an underground mine by the Minnesota Iron Company in 1894-1902. It was reopened as an open pit by the Oliver Iron Mining Company (a subsidiary of United States Steel Corp.) in 1951 and was essentially closed in the 1960s; scram operations continued intermittently until 1999. The mined ore was located along a main ore trough that coincides with the northwest-trending Auburn Fault. Mined material was obtained from the Upper Cherty and Lower Slaty members, and locally from the Lower Cherty member. Jim Small (personal communication to Peter Jongewaard) reported that he was unable to find any evidence of the Auburn Fault anywhere on the newly-scraped floor of the mine during the final scramming period prior to allowing the pit to fill with water. United Taconite is currently filling the mine with waste rock.

STOP 4-4B: Submember LC-4 near the base of the Lower Cherty

Location: Thunderbird North Mine. Eveleth 7.5' quadrangle T.58N., R.17W., Sec 31, 533930E, 5257810N (NAD 83)

Description: The LC-4 submember (BLC submember – old Evtac terminology) exemplifies typical wavy-bedded taconite (see Fig. 4-16 at end of this report) that is characteristic of much of the taconite ore of the Mesabi Range. This unit varies from the previous stop in that wavy-beds are dominant and locally exhibit cross-bedded features. At this stop the magnetite is in: 1. the wavy beds; 2. disseminated throughout the cherty bands; and 3. within mottles that are generally less than 1 cm in diameter and cored by Fe-carbonate (siderite). The LC-4 is 40-45-ft-thick and contains 20-27% mag-Fe and 1-2% Davis Tube Si. The LC-4 submember is easily recognized in drill core due to the wavy beds and a salt-and-pepper texture that is defined by disseminated magnetite.

STOP 4-4C: Submembers LC-5 through LC-8 near the top of the Lower Cherty

Location: Thunderbird North Mine. Eveleth 7.5' quadrangle T.58N., R.17W., Sec 19; 533915E, 5259540N (NAD 83)

Description: At the base of this pit exposure is the LC-5 submember (according to the new U-Tac terminology, or TLC in the old Evtac terminology) characterized by thick- and irregular-bedded, mottled chert with 23-28% mag-Fe. The mottles are pink Fe-carbonate; they are not readily evident on fresh break but appear after a short weathering period. Overlying the LC-5 is a 1-1.5-ft-thick intraformational conglomerate (LC-6 submember – U-Tac terminology) that

contains abundant red jasper fragments. The next overlying unit is the LC-7 submember (U-Tac terminology) characterized by thin magnetite-rich beds interlayered with green, Fe-silicate-bearing, chert beds (1-6 inches thick). The LC-7 is ore with 20-25% mag-Fe, whereas, the overlying LC-8 submember, which physically resembles the LC-7, is waste as the thin beds no longer contain magnetite – rather Fe-silicates are dominant in the thin beds.

This is the last stop at the Thunderbird North Mine. From here we will return to U.S. 53 and proceed about 1.5 miles north to a gated entrance on the right to the "Mineview in the Sky" – follow the signs to the overlook on top of a waste rock dump.

Stop 4-5: <u>Mineview in the Sky Overlook</u>

Location: Eveleth 7.5' quadrangle, T.58N., R.17W., Sec 17, Ne of NW of SE; 535710E, 5261650N (NAD83 UTMs).

Description: The open pit immediately below the overlook is the Rouchleau Mine that connects northward with additional mines. The Rouchleau Mine produced 52,000,000 tons of high-grade "natural ore" containing 50-55% iron from 1920 to 1976. This area is located on the "Virginia Horn" which is a large Z-shaped bend in the otherwise straight ENE trend of the Mesabi Range. Six miles to the northwest on the horizon is United States Steel's Minntac taconite operation. Steam visible to the northeast of Virginia is from the Ispat Inland taconite plant.

From the Mine View we will proceed to the Minorca Mine of Ispat Inland by driving north on U.S. 53 approximately 4 miles around the town of Virginia. About 1 mile north of the third stop light, make a right turn onto an unnamed gravel road (this road is 0.4 miles north of the road to the golf course). Follow this gravel road across 9th Ave. (paved) and continue to the Minorca mine gate.

Stop 4-6 (optional): <u>Cross-bedded LC-3 submember, Minorca Mine</u>

Location: Near the northwest of the Minorca Mine (UTM coordinates will be acquired at the site during the field trip).

Description: Cross-bedding is exceptionally well exposed at one spot in this mine. Most cross-beds here indicate paleocurrents to the north, but one excellent herringbone cross-bedded set shows paleocurrent trends toward azimuth 0^{o} (lower cross-bed) and azimuth 220^{o} (upper cross-bed). See Figure 4-8.

If stop 4-6 is visited, we will cross the Laurentian Divide on private company mining roads to gain access to the Laurentian Mine of Ispat Inland. If stop 4-6 is not visited we will proceed southward on U.S. 53 from the Mineview overlook and head west on Highway 135 (about 3 miles) to the town of Gilbert. At the junction of highways135and 37, turn left and proceed north to the gated entrance to the Laurentian Mine of Ispat Inland. The stratigraphic terminology of submembers in the Laurentian Mine is portrayed on Figure 4-15.

STOP 4-7A: Submember LC-5a and LC-5b at the top of the Lower Cherty

Location: Laurentian Mine. Gilbert 7.5' quadrangle T.58N., R.17W., Sec 24, NW of SE of NE; 542240E, 5260650N (NAD 83)

Description: This stop picks up where we left off at the last stop at the Thunderbird North Mine. At this locality the LC-5a and LC-5b submembers are exposed. The LC-5a is a wavy-bedded taconite ore unit with pink to brown Fe-carbonate patches, lenses, and mottles. The overlying LC-5b submember consists of green Fe-silicate-rich granular cherty material.

STOP 4-7B: Lower Slaty member and the "Intermediate Slate" at the base of the Lower Slaty

Location: Laurentian Mine. Gilbert 7.5' quadrangle T.58N., R.17W., Sec 24, NW of SE of NE; traverse from 542315E, 5260700N to 542485E, 5260840N (NAD 83)

Description: The entire stratigraphic section of the Lower Slaty (130-ft-thick) can be viewed at this impressive face within the Laurentian Mine. The "Intermediate Slate," at the base of the Lower Slaty, is characterized by thin-bedded, black, organic-rich slate that locally exhibits bright shiny graphitic surfaces with bedding-parallel slickensides. Pyrite is common to this submember and is present as both disseminated fine- to medium-grained cubes and as thin disks (marcasite?) along bedding planes. All of the Lower Slaty constitutes waste rock.

Note that the name "slate" has been applied to all thin-bedded rocks in the BIF, but nearly every writer has pointed out that this term is a misnomer, because these rocks do not have the cleavage of a true slate but merely a parting parallel to bedding (White, 1954). Morey (1993) reported that the "Intermediate Slate" is an ash-fall tuff; however the senior author has yet to see any evidence of volcanic shards in thin-sections collected from this unit.

STOP 4-7C: UC-1 submember at the base of the Upper Cherty

Location: Laurentian Mine. Gilbert 7.5' quadrangle T.58N., R.17W., Sec 24, SE of NE of NE; 542545E, 5260790N (NAD 83)

Description: The UC-1 submember is a regular-bedded (1-2 inches straight/even beds) granular cherty unit that contains magnetite-rich bands, pink Fe-carbonate mottles, and local cross-bedding. This material constitutes ore in that it contains 16-40% mag-Fe (the mag-Fe is extremely variable in this unit) and 2-4% Davis Tube Si. Localized pods of very Fe-rich taconite (up to 40% mag-Fe!) were present but have been recently mined out.

STOP 4-7D: UC-2 submember at the middle of the Upper Cherty

Location: Laurentian Mine. Gilbert 7.5' quadrangle T.58N., R.17W., Sec 24, SE of NE of NE; 542590E, 5260740N (NAD 83)

Description: This submember exhibits both regular-bedded and thin-bedded features and consists of both taconite ore zones (18-20% mag-Fe) and waste zones. At this particular site are numerous bedding-parallel quartz-carbonate veins with chlorite slickensides. The chlorite is

particularly troublesome as it cannot be selectively removed from the blasted ore and ultimately "gums up" in the concentrator. Interbedded with the magnetite-rich beds are red-colored hematite-rich beds that appear to be primary – they are pervasive throughout the area and do not appear to be related to a later oxidation event that produced the direct shipping ores of the Mesabi Range. Also at this locality are localized red jasper beds and intraformational conglomerate lenses with red jasper fragments.

STOP 4-7E: <u>UC-3 submember at the top of the Upper Cherty</u>

Location: Laurentian Mine. Gilbert 7.5' quadrangle T.58N., R.17W., Sec 24, SE of NE of NE; 542635E, 5260785N (NAD 83)

Description: At this locality is a mixed thick-bedded, regular-bedded, and diffuse-bedded granular cherty unit with brown Fe-carbonate-rich beds (irregular to wispy appearance) as well as Fe-carbonate mottles and patches. The basal contact with the UC-2 submember is present at this locality and is marked by a seven-foot-thick thin-bedded zone at the base of the UC-3. A quartz-carbonate vein marks the contact.

Within the middle of this unit is an algal stromatolite-bearing unit which will be viewed at the next stop in LTV pit 2E.

This is the last stop at the Laurentian Mine. Return to the mine entrance (the one that is north of Gilbert), turn left on State Highway 135 and proceed west to Biwabik and Aurora. Follow Co. Road 110 through Aurora to Hoyt Lakes. At the stop sign in Hoyt Lakes take a left and proceed down Co. Road 666 (approximately 4-5 miles) to the guard shack near the office buildings at the Cliffs-Erie site (old LTV offices). After getting permission to enter the property, take an immediate right and follow this road about 3.5 miles to a T-intersection with the Dunka Road (private company road). Follow this road about 2 miles through a sharp right turn and a sharp left turn. At the last turn is a series of closed mining roads angling off to the south – walk down the road that points in the more southerly direction to the brow of the hill (Stop 4-8). The stratigraphic nomenclature for the Cliffs-Erie site is presented in Figure 4-15.

STOP 4-8: <u>Algal submember (I submember) near the top of the Upper Cherty</u>

Location: Cliffs-Erie site (old LTV Pit 2E). Allen 7.5' quadrangle T.59N., R.14W., Sec 23, N1/2 of NW; 568202E, 5270625N (NAD 83). Access to this site is via the Dunka Road which is a private mining company road.

Description: Algal structures were first described by Leith (1903) as "contorted bedding." Grout and Broderick (1919) are the first who assigned an organic origin to them. The algal submember within the Upper Cherty consists of mounds of fossilized algal colonies that are separated by jasper-bearing intraformational conglomerate; both the algal and conglomerate units exhibit a combined thickness of 2-20 feet. This horizon occurs only in the eastern half of the range (not present west of Hibbing) and is only sporadically present between Hibbing and Chisholm.

This locality is an excellent place to view a nearly horizontal portion of the iron-formation that contains abundant individual mounds of algal stromatolite. Stripping of glacial overburden in this area has revealed a dip slope the size of a football field that contains the stromatolite mounds

(Graber, 1993). See Figure 4-9 in this field guide. Internally, the mounds are characterized by many individual columnar finger-like structures that are convex upward. The mounds protrude up through a thin veneer of the overlying thin- to wavy-bedded H submember. Studies of a nearby zone in this horizon showed that all the columnar stromatolites were inclined at 30 degrees to the vertical. Unfortunately, this particular site has been removed by mining activities.

Stromatolite samples will be able to be collected at the extreme eastern edge of this exposure. Also at this locality the J and H submembers locally contain anthraxolite, which is an organic bitumen containing 95% or more carbon that is black with a vitreous luster and conchoidal fracture and resembles obsidian (Morey, 1994). Morey (1994) reported that anthraxolite is present throughout the iron-formation but it is most common beneath the carbon-rich "Intermediate Slate." Furthermore, he suggested it formed via a mechanism of concentrating carbon from a mass-kill phenomenon, followed by later migration of a carbon-rich liquid to form the anthraxolite.

After this site we will proceed eastward about 14 miles down the Dunka Road (through a locked gate) to the Peter Mitchell mine operated by Northshore Mining. The stratigraphic nomenclature of submembers in the Peter Mitchell Mine is portrayed on Figure 4-15.

STOP 4-9: <u>Submembers C, D (Upper Slaty), F, G, H, and I (Upper Cherty)</u>

Location: Peter Mitchell Mine. Babbitt 7.5' quadrangle T.60N., R.13W., Sec 26, S ½ of NE. Several subunits of the Upper Cherty can be viewed within short walking distances at this particular stop (see descriptions and locations listed below).

Description:

<u>Submembers G (wavy-bedded taconite ore), H (wavy-bedded taconite ore) and I (stromatolite-bearing unit)</u> at 578533E, 5278414N (NAD83 UTMs).

The I submember is present at the base of the exposure and is overlain by the wavy-bedded H submember. At the top of H is a 0.5-1.0-ft-thick intraformational conglomerate that separates the H and G subunits (the G submember is actually present within the Lower Slaty according to Gundersen and Schwartz, 1962). Note that both H and G subunits constitute taconite ore and both are characterized by wavy-bedding. This same bedding characteristic was also evident in the taconite ore zones within the Lower Cherty that were seen earlier on this field trip. Note also that the bedding characteristics of the Upper Cherty at this locality are wavy rather than massive as at the Laurentian pit.

<u>Submember F with small septaria-like structures</u> at 578857E, 5278440N (NAD83 UTMs)

The F submember is thin- to wavy-bedded and locally contains small septaria-like structures that consist of whitish quartz-filled subvertical fractures in the granular cherty layers. Even though the F submember contains appreciable magnetite, it is classed as waste material because the magnetite is too fine to be economically concentrated.

<u>Submember F with small septaria-like structures and submember G with minor garnets</u> (optional) at 578841E, 5278503N (NAD83 UTMs)

<u>Submember C (with Keweenawan sill) and submember D (optional) at 578730E, 5278206N (NAD83 UTMs)</u>

Both the C and D submembers are thin-bedded units of the Upper Slaty, however, the D submember is different in that it contains slightly thicker beds and lenses of chert. The "contact"

between the C and D is exposed on this bench. A 20-ft-thick Keweenawan sill is also present at this stop (see below for geologic description) and is positioned about in the middle of the C submember.

STOP 4-10: Submembers A, B, and C within the Upper Slaty, partially-melted Virginia Formation, two Keweenawan sills, and base of the Duluth Complex

Location: Peter Mitchell Mine. Babbitt 7.5' quadrangle T.60N., R.12W., Sec 16, W ½ of SW. Several subunits of the Upper Slaty, the Virginia Formation, and three Keweenawan intrusive rocks can be viewed within short walking distances at this particular stop (see descriptions and locations listed below).

Description:

Keweenawan Sill ("BIFSill") within the C submember at 584051, 5280958N (NAD83 UTMs)
A 2-18-ft-thick sill is present in the middle of the C submember at the Peter Mitchell and Dunka Pit areas, and within the J submember in the LTV 2E pit. The sill is generally fine- to medium-grained with locally very coarse-grained plagioclase phenocrysts and polygonal jointing. A granoblastic texture is evident in thin-section indicating that the sill was emplaced in the early Keweenawan and was later metamorphosed by intrusion of the Duluth Complex. Hauck et al. (1997) noted that this sill is chemically similar to the Logan sills, to the northeast in the Rove Formation, and have informally called this sill a "Logan-type" sill.

C submember at 584129E, 5281033N (NAD83 UTMs)
The C submember is dominated by well-laminated, thin-bedded, slaty iron-formation containing magnetite, fayalite, ferrohypersthene, and chert.

Submembers A, B, and C at 584193E, 5281142N (NAD83 UTMs)
At the very top of the BIF is a 2-6-ft-thick chert and marble unit (A submember) that corresponds to the carbonate horizon that is present in only the eastern half of the Mesabi Range. This unit is locally absent in some areas (non-depositional unconformity) and extremely thick in other areas. The B submember is characterized by alternating chert and diopside bands up to one foot thick; marble layers are locally present. In some areas at this stop, pink granophyric veins locally cut the B submember. These veins exhibit pinch-and-swell relationships in that the veins thicken within the diopside bands and pinch in the chert bands.

Keweenawan Sill ("VIRGSill") at base of the Virginia Formation at 584226E, 5281159N (NAD83 UTMs)
At the very base of the Virginia Formation is a 2-100-ft-thick sill that consists of a fine-grained, granoblastic, rock with varying amounts of plagioclase, clinopyroxene, orthopyroxene, hornblende, olivine, and biotite. The informal term of "Cr-bearing sill" was first used by Hauck et al. (1997) to highlight the relatively high chromium contents (600-1,200 ppm) that are characteristic of this sill. This sill exhibits two varieties: 1. a fine-grained, massive, gray-colored unit (this exposure) that is extremely difficult to distinguish from the hornfelsed Virginia Formation; and 2. a medium- to coarse-grained unit that is olivine- and/or hornblende-rich and is easily recognized.

Partially-melted Virginia Formation in close proximity to the Duluth Complex at 584261E, 5280908N (NAD83 UTMs)

In close proximity to the Duluth Complex, the well-bedded sediments of the Virginia Formation are typically transformed into a rock that at first appearance looks like an intrusive rock due to the presence of randomly oriented biotite. This rock is informally referred to as the "recrystallized unit" but is more properly classed as a diatexite (Sawyer, 1999). During emplacement of the Duluth Complex, the sediments of the Virginia Formation were heated, generating 20-40% pervasive partial melts, that literally enabled these rocks to flow in response to stresses that were applied during emplacement. All bedding planes are obliterated and what remains is a medium-grained recrystallized rock that contains plagioclase, cordierite, orthopyroxene, and decussate biotite. Within this recrystallized matrix are blocks/boudins of more structurally competent siltstone and calc-silicate hornfels (originally limey layers).

Basal contact of the Duluth Complex at 584266E, 5280874N (NAD83 UTMs)

At this locality the basal contact of the South Kawishiwi intrusion (SKI) is irregular with localized "fingers" of the footwall Virginia Formation protruding upward into the intrusive rocks. Rocks of the SKI consist of weakly to moderately mineralized, fine- to medium-grained, ophitic augite troctolite to olivine gabbro. Cu-Ni values are unknown for this exposure.

Return to Duluth via highways 21, 169, and U.S. 53.

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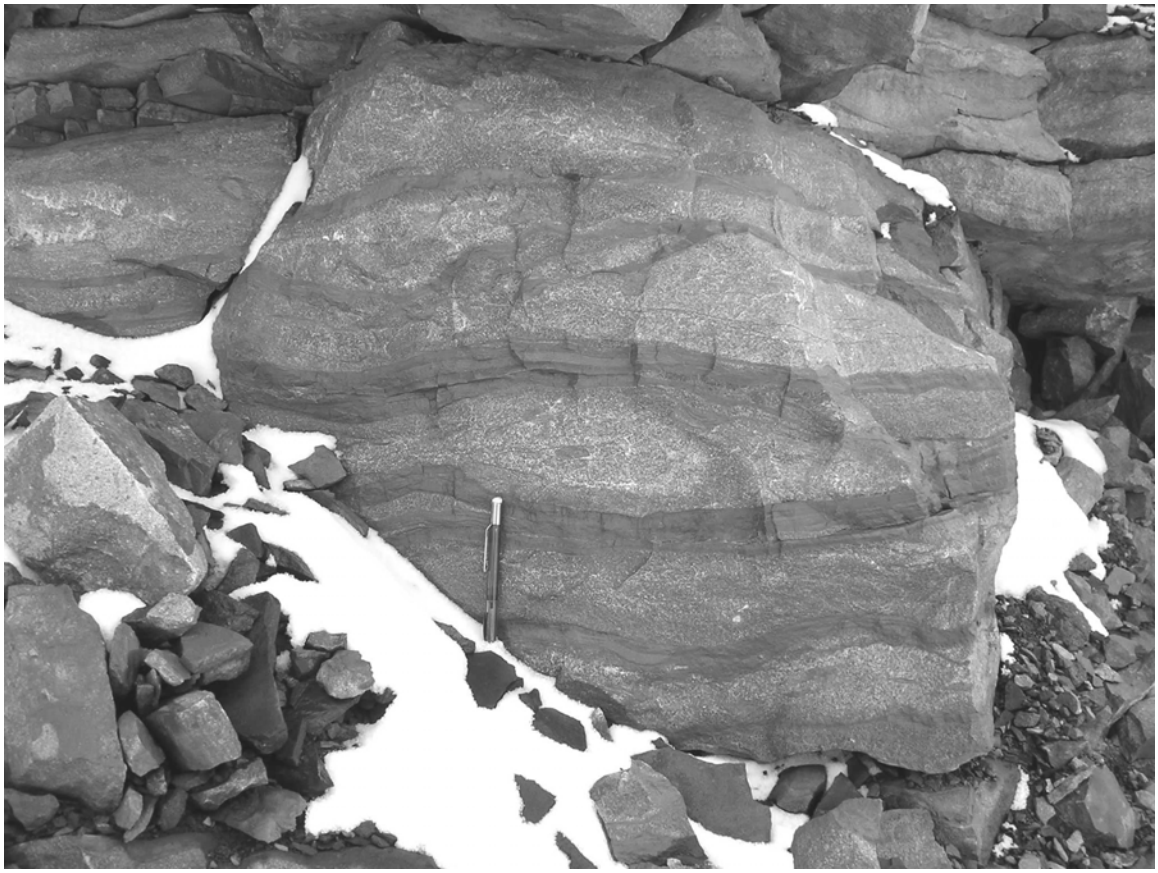


Figure 4-16: Typical example of wavy-bedded taconite (also referred to as irregular-bedded taconite) that is mined from both the Upper Cherty and Lower Cherty members of the Biwabik Iron Formation on the Mesabi Range of northeastern Minnesota. Photo of the LC-4 submember at the Thunderbird North Mine of United Taconite.

FIELD TRIP 5--CLASSIC OUTCROPS OF NORTHEASTERN MINNESOTA

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INTRODUCTION

We propose on this field trip to present the great diversity of Precambrian rock types in northeastern Minnesota using some of the most illustrative and accessible outcrops. We also hope to portray our present understanding of that geologic framework in the context of history—highlighting contrasts between "what we thought we knew" in the earlier years of geologic study, with "what we think we know now." The trip will revisit many of the outcrops on which so many historical discussions, many of them heated, occurred. Geologic controversies were begun, and in some cases resolved, through discussions of the geologic relationships exhibited at some of these localities. This approach is not intended to criticize earlier works or workers; as each has contributed incrementally to the overall geologic picture. Geologic knowledge, like that in most sciences, is "vertical"— that is, each new iteration stands on the shoulders of previous efforts. Obviously, the trip visits only those places that are reasonably accessible from Duluth and from larger roads (Fig. 5.1).

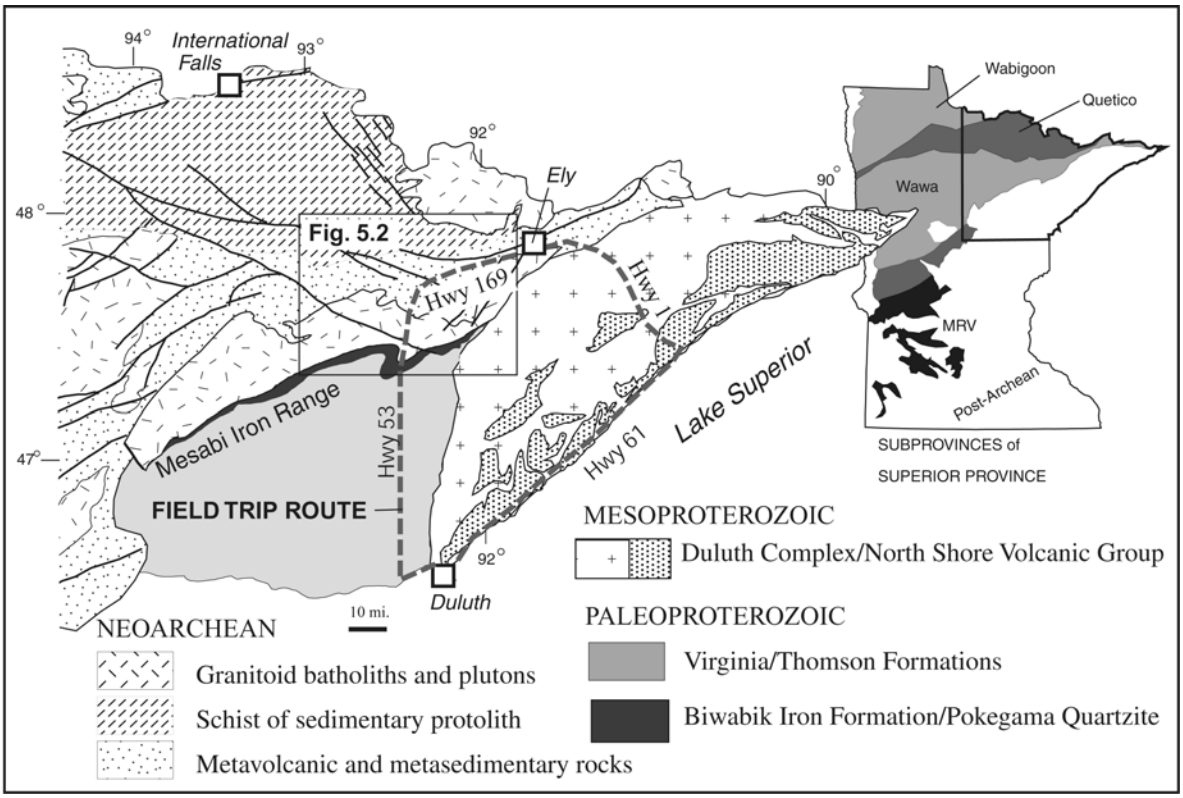


Figure 5.1 Generalized geologic map of northeastern Minnesota showing route of Field Trip #5.

Although we consider many of these to be classic outcrops, it should also be noted that—because of access and location—these are only *some* of the classics; some of the most instructive outcrops are comparatively less accessible. It is likely that we have omitted some favorites, for which we apologize; however, many other fine outcrops can be found in field trip guide books in the accompanying references, including guides of the Institute on Lake Superior Geology, the Minnesota Geological Survey Guidebook Series, and the Geological Society of America (e.g., Biggs, 1987).

HISTORICAL CONTEXT

The Institute on Lake Superior Geology (ILSG) originated in Minnesota—as the "Lake Superior Institute"—with its first meeting held in Minneapolis in 1955, and has met the past 50 years in a variety of localities around the Great Lakes. Including this year's meeting, the Institute has met in Minnesota a total of 13 times; five meetings were headquartered in the Twin Cities, six were held in Duluth, and two in more remote localities such as International Falls and Eveleth. Many, though apparently not all of those meetings, involved field trips demonstrating the diverse geology and mineral resources of the state. Unfortunately, the records of field trips during the earliest meetings are scant. Nevertheless, archives show that field trip topics fall into the following general geologic categories, listed geochronologically (with Minnesota meeting years in parentheses) below:

Quaternary <1.75 Ma
- Twin Cities area (1976, 1998)
- Northern Minnesota (1979, 2004)

Paleozoic <545 Ma
- Southeastern Minnesota (1976, 1979, 1998)

Mesoproterozoic—1600-900 Ma
- North Shore Volcanic Group/Midcontinent rift (1971, 1979, 1989, 1998, 2004)
- Duluth Complex (1965, 1971, 1979, 1993, 2004)

Paleoproterozoic—2500-1600 Ma
- St. Cloud "granite" district/Penokean orogen (1965, 1979, 1989, 1998)
- Virginia/Rove Formations (1963, 1971, 2004)
- Mesabi Iron Range/Biwabik Iron Formation (1963, 1971, 1979, 1993, 2004)
- Gunflint Iron Formation (1971)

Archean—Neoarchean 2900-2500 Ma and Mesoarchean 3400-2900 Ma
- Vermilion and other greenstone belts (1971, 1979, 1989, 1993, 2004)
- Vermilion Granitic Complex/Quetico subprovince (1982, 2004)
- Minnesota River Valley gneiss terrane (1976, 1998)

Although there apparently was no companion field trip for the first ILSG meeting in 1955; the "hot topics," as judged from printed abstracts of that meeting, included the following:

- Age, stratigraphic setting, origin, and lithologic subdivision of Lake Superior iron-formation (presentations by Harold James, Robert Schmidt, and J.F. Wolff)
- Composition and origin of iron ores; addressing competing theories on the creation of "natural ores" by rising hydrothermal vs. descending meteoric waters (Burton Boyum, Stanley Tyler, David White, N.K. Huber, Alan Broderick, Henry Lepp, Tsu Ming Han)
- Geophysical characteristics of the "Lake Superior Syncline" (Edward Thiel, Harold Mooney, George Wollard, Lloyal Bacon)
- Copper-nickel potential in Duluth gabbro, discovered in 1948 (Gerald Anderson and Donald Yardley)

Not surprisingly, most of these same geological topics are important today, and are featured in the various presentations and field trips of this 50th anniversary meeting. Some of the more

significant are represented by this field trip to the "classic" outcrops of northeastern Minnesota geology.

Many of the controversies during early geologic study of the Lake Superior region involve attempts to correlate geologic units, and the consequent application of stratigraphic nomenclature. In Minnesota and adjacent Ontario, this type of controversy is typified by the debate surrounding the "Couchiching series" in the Rainy Lake area, and the stratigraphic and temporal relationships of Precambrian units in the region. Based on field work, largely in Canada, Lawson (1887) proposed that there was a sequence of sedimentary rocks older than Keewatin greenstone, and named these rocks the Couchiching series. Other workers, notably those with the U.S. Geological Survey, strongly disagreed. The question was so fundamental to understanding the regional geology that it was addressed by a special committee of geologists from the U.S. Geological Survey, Geological Survey of Canada, Michigan Geological Survey, and the Ontario Department of Mines, who visited field localities on both sides of the border. The result was a new model which was applied, rightly or wrongly, to the geology of other greenstone sequences. In retrospect, the controversies arose from the fact that geologists working in different areas observe or infer field relationships and attempt (as a natural consequence of human endeavor) to apply those relationships everywhere, including far afield from their immediate areas of knowledge—the "blind man and the elephant" syndrome. Furthermore, lacking geochronologic markers and dating methods, early workers applied temporal constraints based on generalized rock types; for example, implying that all volcanic rocks are "Keewatin," all sedimentary rocks are either Seine (younger than Keewatin greenstone) or Couchiching (older than Keewatin greenstone), and recognizing two distinct periods of granitic rock emplacement as either Laurentian (pre-Seine) or Algoman (post-Seine). We now know from additional field work, advances in the understanding of depositional environments, and geochronologic studies that many periods of volcanism, sedimentation, and pluton emplacement have occurred.

Much of what was known about the Precambrian geology in Minnesota by the 1950s—when the Institute on Lake Superior Geology (ILSG) was in its infancy—was the result of efforts by F.F. Grout, J.W. Gruner, G.M. Schwartz, and G.A. Thiel, all professors of geology at the University of Minnesota, and their graduate students working for the Minnesota Geological Survey. They summarized their views in a long paper entitled "Precambrian stratigraphy of Minnesota" (Grout and others, 1951). This publication served as backbone for a Geological Society of America field trip in 1956 (Schwartz and others, 1956). This was pioneering work that described major rock units and formalized their classification. It divided Precambrian time into a general three-fold classification scheme of Earlier, Medial, and Later—each Era separated by "great unconformities" in the rocks. The efforts established a geologic framework that was tested and refined by many workers, commonly reporting their results during the early years of the ILSG. For example, geochronologic and geologic studies in the 1960s established the Laurentian as an orogenic event, and much of the evidence for that was presented at ILSG meetings and field trips.

FIELD TRIP FACTS

The field stops include rocks ranging in age from Archean to Mesoproterozoic (Fig. 5.1 and Table 5-1). For the most part, the stops are presented in geographic rather than geochronologic order. Temporal settings can be deduced from the various figures, descriptions, and Table 5-1. Five topical areas are represented:

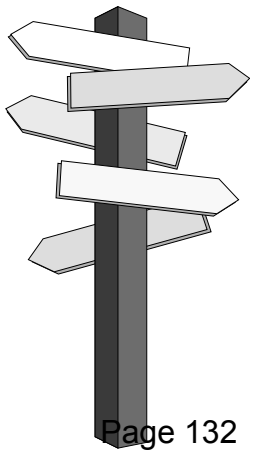
1) Virginia Horn—Archean and Paleoproterozoic (Stops 5-1 to 5-7)
2) Tower-Soudan—Archean (Stops 5-8 to 5-14)
3) Ely area—Archean (Stops 5-15 to 5-18)
4) North Shore and Duluth—Mesoproterozoic (Stops 5-19 to 5-25)
5) West of Duluth—Paleoproterozoic and Mesoproterozoic (5-26 to 5-28)

FORMATION/GROUP/SUCCESSION		INTRUSIONS	EVENT/SETTING	ERA
Keweenawan Supergroup	Hinckley Sandstone Fond du Lac Sandstone (27) North Shore Volcanic Group (19) (20) (22) (23) Puckwunge/Nopeming sandstones (26)	Duluth Complex (24) (25) Beaver Bay Complex (21) Diabase dikes (10)	**Midcontinent Rift System**	900 Ma MESOPROTEROIZOIC 1600 Ma
Animikie Group	Virginia/Thomson/Rove Formations (28) Biwabik/Gunflint Iron Formations (~1880 Ma) (6) Pokegama Quartzite (5) (4)	K-K diabase dikes	**Penokean Orogeny**	PALEOPROTEROZOIC 2500 Ma
Superior Province: **Quetico subprovince**		Vermilion Granitic Complex (18) Giants Range batholith (7)	**D3 Deformation** **D2 Deformation** (13) (~2680 Ma)	NEOARCHEAN
Superior Province: **Wawa subprovince**	Knife Lake Group/Midway (3)/Ogishke cgl. (<2690 Ma) Newton Lake Formation/Bass Lake sequence (16) (17) Lake Vermilion Formation (8) (9) (10) (11) (14) Mud Lake (1) (2)/Ely Greenstone (15) (~2720 Ma) (Upper, Soudan iron-formation, and Lower members) (12)		**D1 Deformation**	

TABLE 5-1—Stratigraphic and temporal classification of Precambrian rocks in northern Minnesota (after Sims, 1972). Ages are referenced in the text. Circles enclose the approximate chronostratigraphic position of field trip #5 stops.

Considerably more stops appear in this guide than can reasonably be visited during a two-day trip. Other stops are included to provide a more thorough view of the regional geology and to facilitate self-led excursions. Some of these stops, as noted below, are described in detail in other field trip texts in this guide book, and not repeated here. Furthermore, much of the regional geologic framework is covered in the introductory material of those field trips, as noted below. The descriptions of many of the stops contain brief discussions of the "Historical Perspective," providing the contrast between early and modern understanding of the geology.

All UTM locations are given in NAD 83, Zone 15 coordinates. Section subdivisions read from smallest to largest quarter; e.g., "NW, SE, SW" should be read "NW quarter of the SE quarter of the SW quarter." The small map insets showing stop locations are taken from USGS 7.5-minute topographic quadrangles listed with each stop.



FIELD TRIP STOPS

Virginia Horn area: Stops 5-1 to 5-7 (refer to Figures 5.2 and 5.3)

Regional geologic context of the Archean and Paleoproterozoic rocks described below is reviewed in the introductory material to Field Trips#1, #4, and #7.

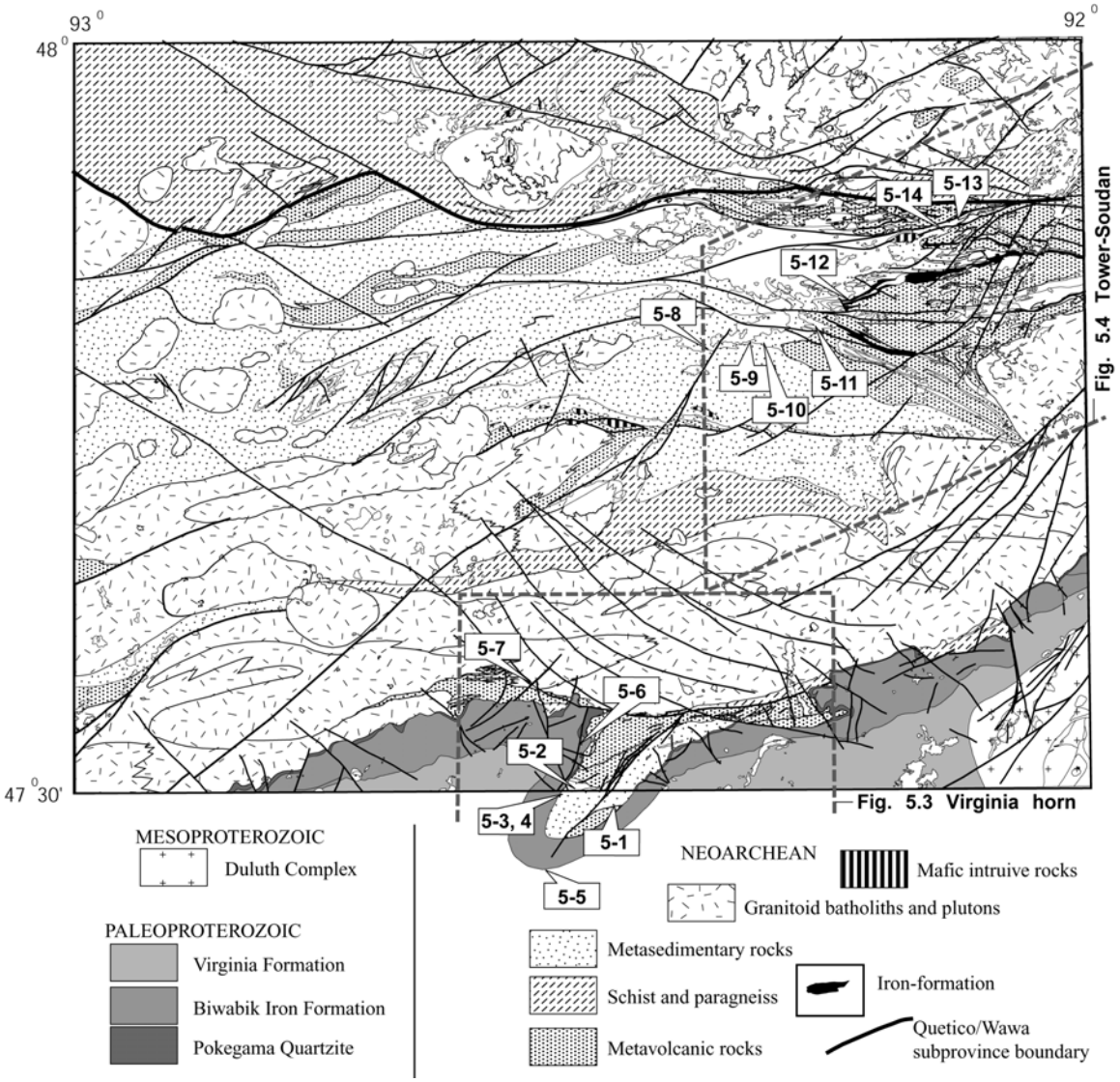


Figure 5.2 Simplified geologic map of Vermilion Lake 30X60' quadrangle (modified from Jirsa and Boerboom, 2003) showing location of field stops 5-1 to 5-13. Insets show location of additional figures.

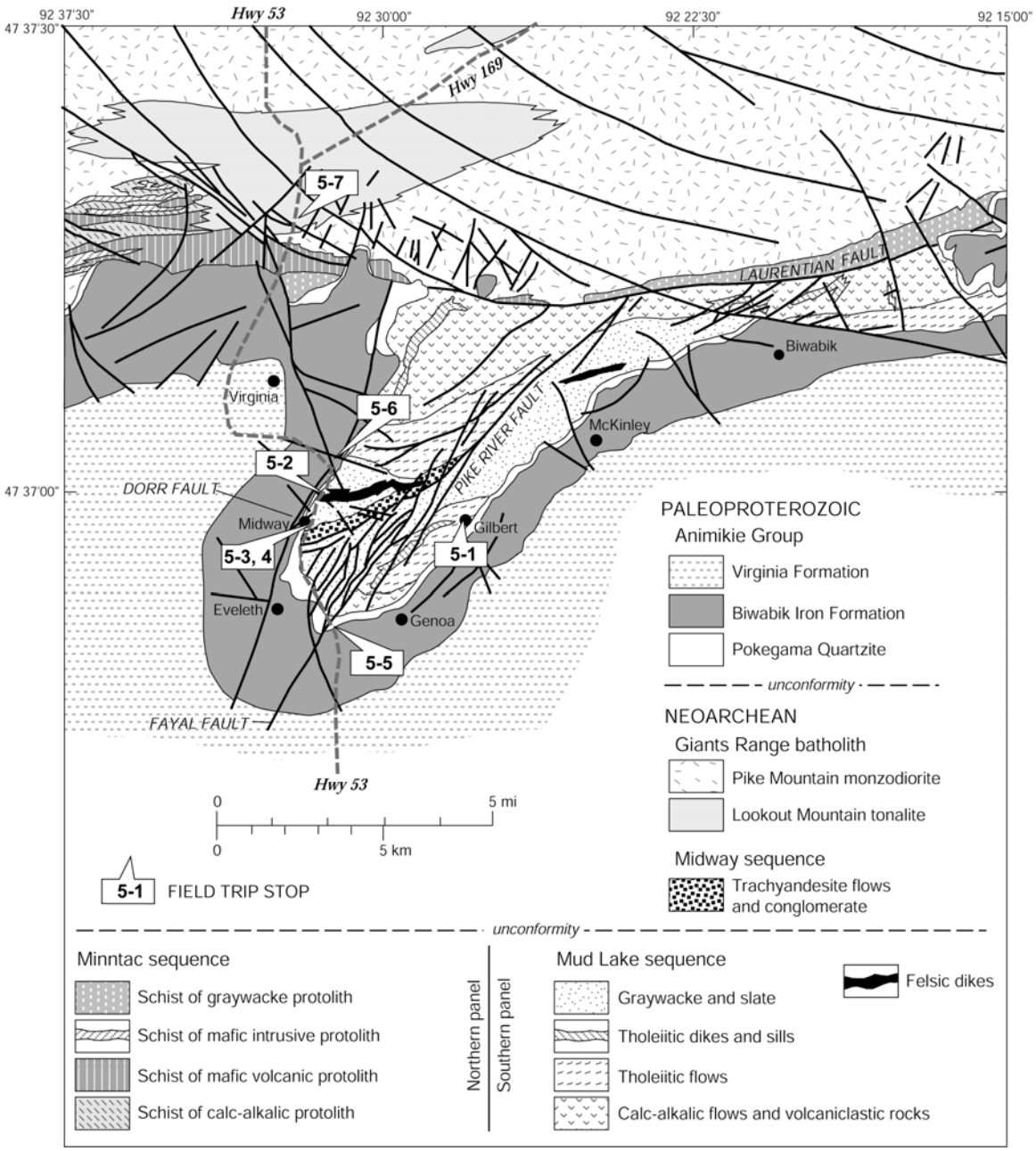
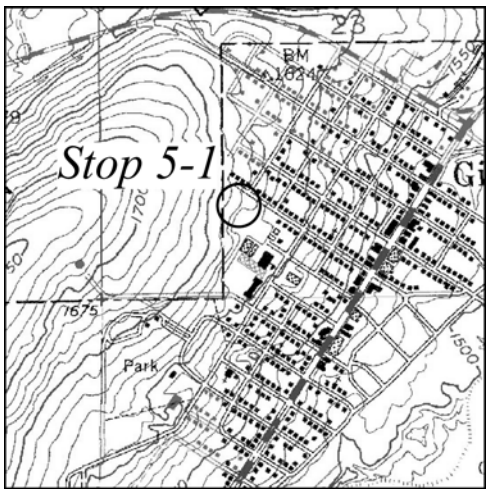


Figure 5.3 Generalized geologic map of the Virginia horn area (modified from Jirsa and others, 1998) showing details of field trip stops 5-1 to 5-7.



STOP 5-1.
<u>Archean pillowed and massive greenstone</u>

Location: T.58N., R.17W., sec.23 NW, SE, SW; north edge of athletic fields, Gilbert Junior High School.

Gilbert 7.5-minute quadrangle

UTM: 539,820E/5,259,750N

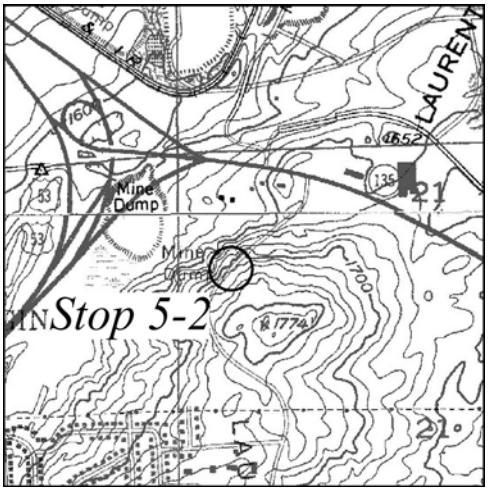
DESCRIPTION:
Outcrop of pillowed and massive basalt is part of the Archean Mud Lake sequence, metamorphosed to low greenschist-grade. Pillow shapes indicate stratigraphic facing is to the northwest, which places this outcrop on the south side of a major D_1 structure known as the Mud Lake syncline. Note also the presence locally of fractures filled with reddish jasper, presumably deposited in depressions on the rock surface by overstepping of Paleoproterozoic seas during deposition of the Biwabik Iron Formation.

HISTORICAL PERSPECTIVE STOP 5-1:

Early work by Gruner (1941) and later by Sutton (1963) interpreted the Archean volcanic and graywacke succession now known as the Mud Lake sequence as an anticline, based on few stratigraphic top indicators, and perhaps influenced by early discussions of the "Couchiching series" (Lawson, 1887) that placed metasedimentary rocks of the Rainy Lake area stratigraphically beneath "Keewatin" volcanic rocks. Detailed structural study by Jirsa and others, (1998) and Jirsa and Boerboom (2003) demonstrate that tholeiitic (Stop 5-1) and calc-alkalic volcanic rocks and tholeiitic intrusions are conformably <u>overlain</u> by graywacke and slate (Stop 5-2). In detail, the succession forms a broad, twice-deformed syncline that has been segmented by faults of several generations.

NEXT:

Return to Highway 37, travel northwest to Hwy 135, thence west on 135 approximately 2.5 miles to Bourgin Road. Turn left (south) on Bourgin Road and continue about 0.4 mile to large cut on left (east) side of road (Stop 5-2).



STOP 5-2.
<u>Archean graywacke and slate, intruded by quartzofeldspathic porphyry</u>.

Location: T.58N., R.17W., sec.21 SW, SW; road cuts on east side of the Bourgin Road.

Eveleth 7.5-minute quadrangle.

UTM: 536,311E/5,260,659N

DESCRIPTION:

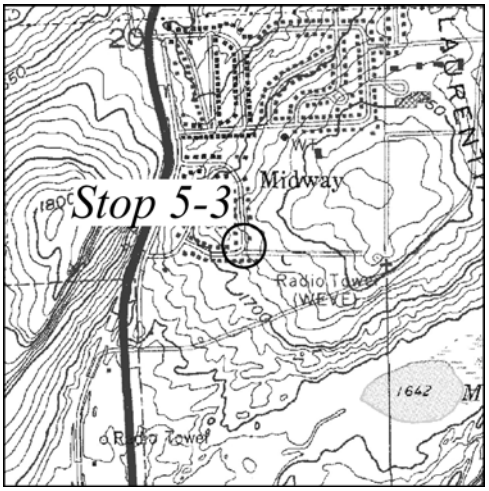
Outcrops along this side of the road expose quartzofeldspathic porphyry (QFP) intruded into variably deformed graywacke, siltstone, and slate of the Mud Lake sequence. The sedimentary rocks here are moderately deformed, but much of that deformation is inferred to predate the main cleavage-forming event D_2, and some may be soft-sediment in origin. The QFP is large and continuous to the east, but at this locality it appears to be segmented into a zone of multiple 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 QFP intrusions—presumably because the QFP 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 surprisingly barren.

HISTORICAL PERSPECTIVE STOP 5.2:

One of the earliest gold discoveries in Minnesota was made by J.W. Gruner (in Grout, 1937) in a railroad cut not far from stop 5-2. The cut exposes graywacke intruded by quartzofeldspathic porphyry, having visible gold associated with small quartz veins. Despite several episodes of prospecting and systematic study of the region, no economic gold deposits have been discovered.

NEXT:

Follow Bourgin road to the south and west to a frontage road on the east side of Hwy 53. Turn north (right) on the frontage road and travel about 0.2 miles to first road to right, turn up-hill and continue to #7 Mesabi Lane (Stop 5-3).



STOP 5-3.
Archean conglomerate

Private driveway!

Location: T.58N., R.17W., sec.20 SW, SE, No. 7 Mesabi Lane; village of Midway.

Eveleth 7.5-minute quadrangle.

UTM: 535.713E/5.259.459N

DESCRIPTION:

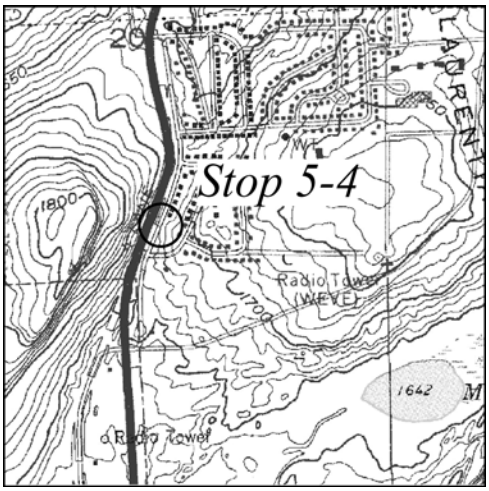
Archean conglomerate and lithic sandstone that form the driveway here are part of the northeast-trending Midway sequence, containing these strata types locally interbedded with subaerially deposited, calc-alkalic (trachyandesitic) volcanic rocks. The sequence is inferred to have formed after earliest deformation (D_1) of the enclosing graywacke and basaltic rocks of the Mud Lake sequence, but before the cleavage-forming D_2 deformation that affected both sequences. The conglomerate contains clasts of basalt, graywacke, porphyritic trachyandesite, and quartzofeldspathic porphyry (QFP). This provenance indicates that the older Archean rocks of the Mud Lake sequence were intruded by QFP, deformed, and uplifted, to provide detritus to what was probably a successor or "pull-apart" basin developed along a major structure now occupied by the Pike River fault zone. Note also the presence of remnant skins of red jasper as at stop 5-1.

HISTORICAL PERSPECTIVE:

Midway sequence conglomerate has previously been interpreted as a basal sediment (Sutton, 1963), and as a proximal turbidite fan deposit (Levy, 1991), depositionally transitional with graywacke and slate of the Mud Lake sequence. Subsequent work (Jirsa, 2000) indicates that the conglomerate is part of a Timiskaming-type clastic and volcanic sequence that unconformably overlies the older volcanic strata. Deposition of the Midway sequence required uplift, subaerial erosion, continental volcanism, and deposition in isolated basins along a major structural break. Interestingly, Gruner, 1941, inferred that these rocks may be equivalent to similar units of conglomerate in the Knife Lake area, which are now also considered to represent successor-basin deposits that post-date deposition of older greenstone by as much as 30 million years (Corfu and Stott, 1998). The rocks share many attributes with other Timiskaming-type sequences, including the namesake Timiskaming Group near Timmins and the Shebandowan Group near Thunder Bay. The use of the term Timiskaming in this context is quite different from that of Goldich and others, (1961), who applied it as a system term to include the Seine, Knife Lake, and other presumably correlative metasedimentary units.

NEXT:

Return to frontage road paralleling Hwy 53; turn north and travel about 500 feet to low outcrop (Stop 5-4) on east side of road.



STOP 5-4.
Archean/Paleoproterozoic unconformity

Location: T.58N., R.17W., sec.20 SW, SE; hamlet of Midway.

Eveleth 7.5-minute quadrangle.

UTM: 535,441E/5,259,520N

DESCRIPTION:

In the brush on the east side of the frontage road is a small exposure of Archean conglomerate, having nearly vertical foliation and lithologic content much like that at stop 5-3, capped by a thin and discontinuous "skin" of subhorizontally foliated conglomerate. The latter represents basal deposition of the Paleoproterozoic Pokegama Quartzite, the lowest of a tripartite sequence of formations that constitute the Animikie Group. Age dates are somewhat speculative; however, the Archean rocks of the various sequences in the horn probably are about 2.7 Ga (e.g., Peterson and others, 2001). Based on a date from tuffaceous rocks in the Gunflint Iron Formation (Fralick and others, 2002) that is inferred to be equivalent to the Biwabik Iron Formation, the Animikie Group is about 1.8 Ga. Thus, this unconformity represents a geological hiatus of approximately 900 million-years— almost twice that of all Phanerozoic time.

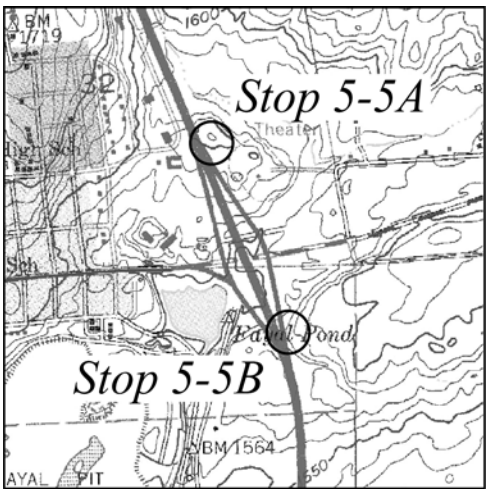
The large road cut visible on the west side of Highway 53 is the argillaceous lower member of the Pokegama Quartzite. The road cut exposes shale, siltstone, and minor sandstone, which has been interpreted by Ojakangas (1993) as having been deposited in a low-energy upper tidal flat environment in a sea that peneplaned the Archean surface. Minor channeling is common at the base of thicker sandstone beds, and small-scale cross-bedding occurs in some siltstone beds. Local soft-sediment deformation may be the product of syn-depositional tectonism, or alternatively may represent localized collapse near tidal channels.

HISTORICAL PERSPECTIVE STOPS 5-4 and 5-5:

The Paleoproterozoic rocks exposed at Stops 5-4 and 5-5, and discussed at stop 5-6 were previously considered to represent a "geosynclinal sequence," the precise tectonic mechanism of which was poorly understood. Work by a number of authors (notably Southwick and others, 1988) demonstrates that the tripartite formations exposed along the Mesabi Iron Range were deposited along the leading edge of a foredeep known as the Animikie basin that transgressed north over the Archean craton during the Penokean orogeny. In detail, deposition of basal quartzite (Pokegama), medial chemical (Biwabik) and upper turbiditic (Virginia) sediments represents a transgression of near-shore, shelf, and slope environments, respectively (Ojakangas, 1993).

NEXT:

Travel south on Highway 53 about 1.5 miles; stop on Highway and cross (carefully) to large outcrop on east side (Stop 5-5A)



STOP 5-5.
Paleoproterozoic Pokegama Quartzite (A) and Biwabik Iron Formation (B)

Location: T.58N., R.17W., sec.32 SE, SE, and adjacent, junction of Highways 37 and 53.

Eveleth 7.5-minute quadrangle.

UTM: scattered outcrops extend from 535,956E/5,256,913N on the north (stop 5-5A), to 536,263E/5,256,200N on the south (stop 5-5B).

DESCRIPTION **5-5A**:

This is the sandy, upper member of the Pokegama Quartzite. It is characterized by coarse grain size and massive beds as thick as 1.5 m. Massive beds are separated by thin beds of shale and siltstone. Ojakangas (1993) interpreted the deposition of this facies as within high-energy, lower tidal or subtidal environment.

NEXT:

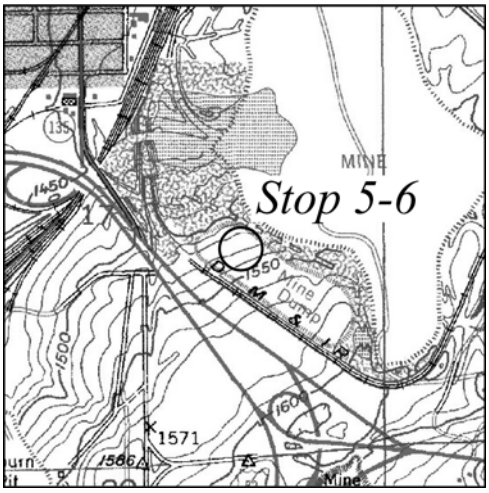
Return to vehicle and continue south on Highway 53 past Highway 37. Turn around and drive north to large outcrop near the SE corner of the junction of Highways 53 and 37 (stop 5-5B)

DESCRIPTION **5-5B**:

This exposure of gently southeast-dipping strata is part of the Lower Cherty member of the Biwabik Iron Formation. It overlies and is generally in transition with the Pokegama Quartzite at stop 5-5A. Notice that both formations have sandy textures and cross-bedding, implying a moderately high-energy depositional environment. The most significant difference between these two units is the abrupt change in sediment source from the extrabasinal quartz grains in the Pokegama, to recycled, chemically precipitated chert in the Biwabik. Measurements of cross-bedding in the iron-formation are bimodal, implying deposition in a tidally influenced marine environment (Ojakangas, 1993).

NEXT:

Drive north on Highway 53 approximately 4 miles to entrance to Mineview overlook (Stop 5-6), just northwest of the junction of Highways 53 and 135. Follow driveway to overlook.



STOP 5-6.
Mineview in the Sky Overlook

Location: T.58N., R.17W., sec.17 NW SE. North of the junction of hwys 53 and 135.

Virginia 7.5-minute quadrangle.

UTM: Top of overlook approximately 535,710E/5,261,650N

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. In detail, there were some 15 separately named mines within view that shipped ore during the period 1893-1986. All of them, and nearly 500 more along the 150-mile long Mesabi Iron Range, extracted oxidized (hematite- or goethite-rich) and leached iron-formation generally referred to as "natural ore." The ore deposits here are localized along a set of fault zones (Fig. 5.3) that presumably provided the plumbing system for fluids that first oxidized the formation and produced permeability; and secondly, leached silica from the porous zones. Natural ores contained as much as 50 percent iron and less than 10 percent silica. By contrast, the mammoth open-pit mine in the distance to the northwest, US Steel's Minntac mine, is developed in unoxidized, magnetite-rich ore containing about 30 percent iron, and 50 percent silica. This type of ore at Minntac, and five other open-pit mines currently in operation along the range, is the source of the iron concentrate known as taconite. The name taconite has also been applied generally to unoxidized iron-formation containing sufficient iron to be mined for a profit using today's technology.

HISTORICAL PERSPECTIVE STOP 5-6:

Nearly 70 percent of the 3.6 billion metric tons of iron ore produced on the Mesabi range was extracted as natural ores. Although it has been generally accepted that these ores formed along fracture, bedding, and fault planes by processes of oxidation and leaching, the source of altering solutions has been the subject of considerable debate among economic geologists for nearly 70 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 no oxidized (and subsequently metamorphosed) iron-formation exists on the eastern-most Mesabi range that was metamorphosed by the Mesoproterozoic Duluth Complex.

The precise origin of the structural bend in iron-formation known as the Virginia horn has long been questioned. It has been explained as a paired anticline-syncline vs. a modified warping

around a fault-bounded horst (Morey, 2003). A topic for future research involves the extent to which vertical tectonism may have been operating *during* deposition of the iron-formation.

NEXT:

Return to Highway 53 and travel north through the city of Virginia to wayside rest approximately 2 miles north of town (Stop 5-7).



STOP 5-7.
Archean Giants Range batholith at "Confusion Hill," Laurentian Divide

Location: T.59N., R.17W., sec.19 SE, SE; wayside off Highway 53.

Virginia 7.5-minute quadrangle.

UTM: 534,337E/5,269,458N

DESCRIPTION:

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. Dikes of tonalite that cut the adjacent high-grade supracrustal rocks of the Minntac sequence contain metamorphic fabrics, yet little evidence of metamorphic origin can be seen in the interior of the body, implying it is syntectonic with respect to D_2 deformation. U-Pb zircon dates (Boerboom and Zartman, 1993) of two components of the batholith exposed to the north bracket the age of D_2 deformation between about 2674 and 2682 Ma. 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 40-mile wide 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 Tower-Soudan area—making stratigraphic correlation between the two districts speculative.

HISTORICAL PERSPECTIVE STOP 5-7:

It was once generally thought that the mafic component of this mixed rock represented xenolithic rafts of subjacent amphibolite grade metavolcanic strata. In the process of regional mapping, Boerboom (in Jirsa and Boerboom, 2003) recognized that the intimately co-mingled tonalitic and mafic phases are the product of magma mixing.

NEXT:

Travel north on Highway 53 to the junction with Highway 169; follow the latter to the northeast approximately 28 miles to County Road 77. Turn left (northwest) on 77 and proceed about 0.5 mi. to Pike River (Stop 5.8).

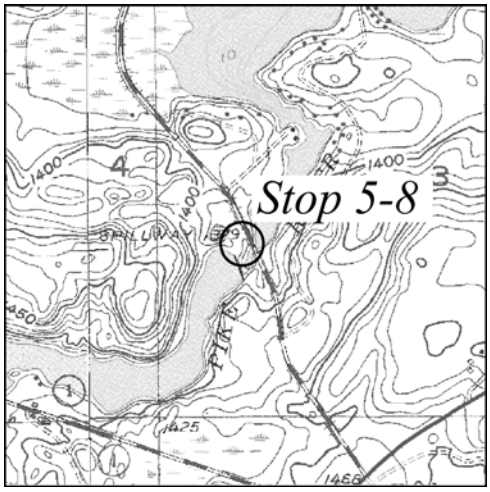
Tower-Soudan Area (refer to Figures 5.2 and 5.4)

Much of the framework geology for the Vermilion district, including the Tower-Soudan area, is covered in the prelude to other trips in this guide, notably field trips #1 and #7.

HISTORICAL PERSPECTIVE STOPS 5-8 to 5-18; Vermilion District

The application of formal stratigraphic nomenclature to these rocks that are poorly constrained by incomplete exposure and scant geochronologic control produced considerable confusion over the years. For example, early workers considered all metabasaltic sequences equivalent, and applied the name "Ely Greenstone" to similar rocks exposed in a broad area of northern Minnesota. It is clear from subsequent mapping and geochronologic work that "greenstones" in the various belts are not all correlative. Use of the term continues today; but is restricted to specific sequences in the Ely and Tower-Soudan areas.

In his publication of 1887, Lawson interpreted rocks of the International Falls area as depicting "Couchiching" sedimentary rocks to lie beneath "Keewatin" volcanic strata, and these interpretations were extended to explain stratigraphic relationships throughout much of Minnesota including the Vermilion district. Lawson's criteria were based in part on stratigraphic younging in metasedimentary sequences. More recent interpretations recognize that these strata have experienced at least two deformations, the earliest of which (D_1) involved inversion of large packages of strata into great thrust nappes prior to the major cleavage-forming event (D_2). Thus, the boundaries between sedimentary and volcanic rocks in many areas are more structural than stratigraphic, and the use of younging directions in these strata requires detailed structural mapping to fully evaluate. In addition to structural mapping (e.g., Hudleston, 1976; Hudleston and others, 1988; Schultz-Ela and Hudleston, 1991; Jirsa and others, 1992), the study of geochemistry (e.g., Arth and Hanson, 1975; Schulz, 1980; Southwick and others, 1998) and U-Pb zircon geochronology (e.g., Boerboom and Zartman, 1993; Peterson and others, 2001) in Minnesota and adjacent Ontario are beginning to resolve some of the temporal issues. Much work remains in this endeavor.



STOP 5-8.
Archean graywacke at Pike River Dam

Location: T.61N., R.16W., sec. 3, NW, SW;
west side of County Road 77, on N side of river.

[Note that Fortune Bay Casino—the overnight hotel—lies to the north off of CR 77].

Tower 7.5-minute quadrangle.

UTM: 547,300E/5,293,340N

HISTORICAL PERSPECTIVE STOP 5-8

Prior to about the 1950s, no depositional mechanism could satisfactorily explain the coincidence in graywacke of 1) coarse sand derived from a source many kilometers distant and having an altered clayey matrix; 2) interbedded black slate; and 3) the lack of evidence for reworking in shallow water (indicative of deposition below wave base). This was changed when the concept of turbidity currents was introduced to the geological profession by Kuenen and Migliorini (1950). Despite widespread publication on turbidites in more modern geologic

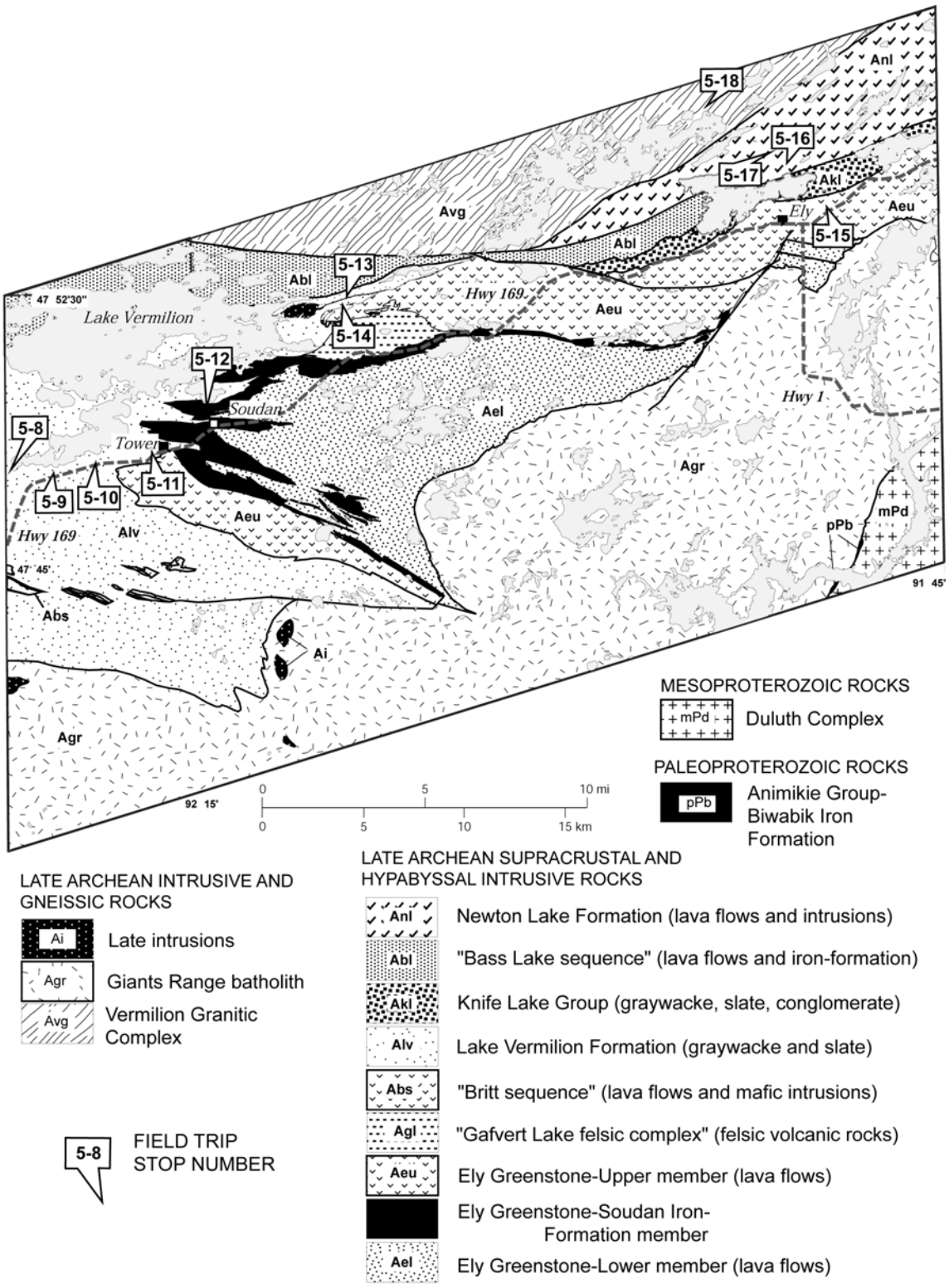
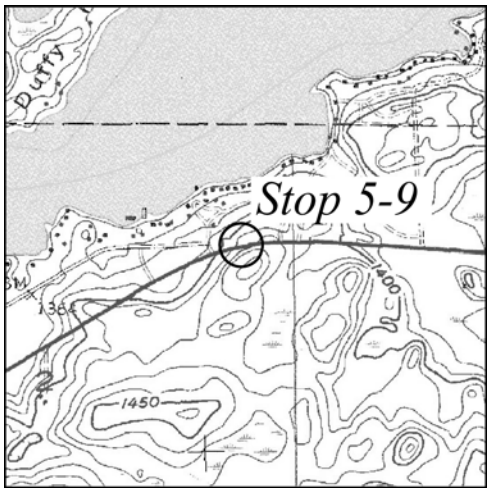


Figure 5.4. Geologic map of the western Vermilion district (modified from Peterson and Jirsa, 1999), including the Tower-Soudan and Ely areas, and showing details of field trip stops 5-8 to 5-18.

settings through the 1950s and 1960s, the facies model was not refined and applied to Archean and Proterozoic strata in the Lake Superior region until somewhat later (Morey, 1965; Ojakangas, 1966).

NEXT:

Return to Highway 169 and turn east. Continue approximately 1.7 miles to just east of the junction of 169 and County Road 526.



STOP 5-9.
Multiply folded Archean graywacke

Location: T.61N., R.16W., sec. 2, NE, NE; south side of Highway 169 just east of CR 526.

Tower 7.5-minute quadrangle.

UTM: 550,050E/5,294,000N

DESCRIPTION:

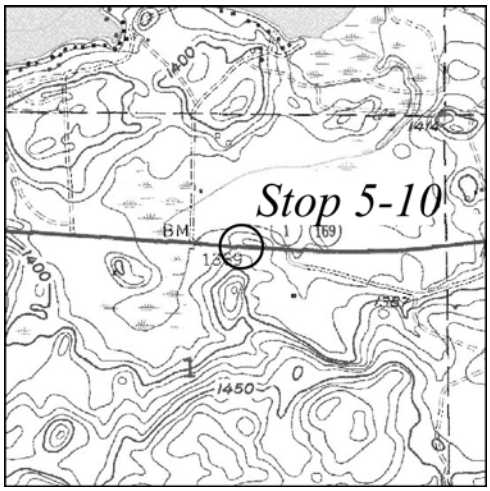
This outcrop at the road and several smaller ones in the bush nearby show the superposition of two generations of folds in thin-bedded, well-graded graywacke of the Lake Vermilion Formation. The second-generation folds (F_2) are associated with a regional axial plane cleavage in which sedimentary clasts are flattened. The earlier F_1 folds have no associated cleavage and tend to be erratic in form, trend, and distribution. Folds display "eye" and "mushroom" shapes that locally are interpreted to be sheath folds (Hudleston and others, 1987). These characteristics are consistent with deformation of poorly lithified sediment. The superposition of deformation events is manifest in the transection of F_1 folds by cleavage related to D_2. In this area and to the west, one can find anticlinal synclines and synclinal anticlines, indicating stratigraphic inversion prior to D_2 folding.

HISTORICAL PERSPECTIVE STOP 5-9:

The complex structure of the Vermilion district was poorly understood in the 1950s and 1960s. Subsequent structural study (Hooper and Ojakangas, 1971; Hudleston, 1976; Ojakangas and others, 1978; Jirsa and others, 1993) demonstrated two distinct periods of deformation: D_1 that includes largely soft-sediment deformation represented at Stop 5-9, and a D_2 transpressive deformation and metamorphic cleavage-forming event. Much of the early D_1 deformation produced broad areas of down-facing strata formed by large thrust nappes (Jirsa and others, 1992). Prior to this recognition, there was much confusion about temporal relationships of individual volcanic and clastic sequences.

NEXT:

Continue east on Highway 169 approximately 0.7 mile to Stop5-10.



STOP 5-10. [Optional] <u>Archean dacitic tuff/ Paleoproterozoic or Mesoproterozoic diabase dike</u>

Location: T.61N., R.16W., sec. 1, SW, NE; Highway 169 road cut.

Tower 7.5-minute quadrangle.

UTM: 551,160E/5,293,960N

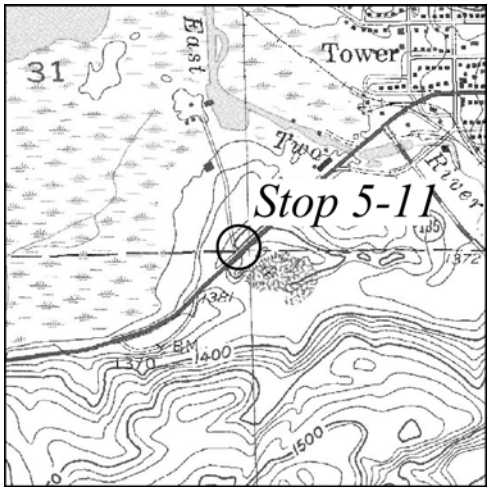
DESCRIPTION:

These road cuts expose outcrops of white, dacitic tuffaceous sedimentary rock, a component of the Lake Vermilion Formation. Regionally, the formation consists of all compositional gradations between what appears to be first-cycle tuff, tuffaceous graywacke, and mixed-source graywacke, interbedded on all scales. In a general way, the tuffaceous component increases in proportion to the east toward the Tower-Soudan anticline, presumably the source region of volcanic detritus. Rocks of the anticline are interpreted to represent the core of what was probably a large, composite volcanic shield complex, bordered by irregular basins composed of detritus shed from the shield, now represented by the Lake Vermilion Formation.

The northeast-trending, steeply dipping, seven-meter wide diabase dike that cuts tuffaceous rocks has been the source of considerable debate. It's petrographic (olivine-bearing) and geochemical (silica undersaturated) composition is similar to Mesoproterozoic dikes (Schmitz, 1994); yet it lies nearly along strike with, though east of, dikes of the Paleoproterozoic Kenora-Kabetogama dike swarm.

NEXT:

Continue east on Highway 169 approximately 1.5 miles to road cut.



STOP 5-11 [Optional]
<u>Archean fragmental volcanic rocks</u>

Location: T.62N., R.15W., sec. 32, SW, SW; Highway 169 road cut, west edge village of Tower.

Tower 7.5-minute quadrangle.

UTM: 553,380E/5,294,430N

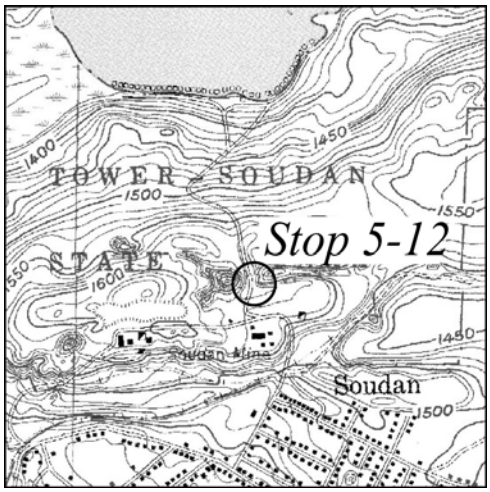
DESCRIPTION:

This outcrop consists of fragmental, variably reworked volcanic conglomerate and tuffaceous rocks of the Lake Vermilion Formation. The presence of this rock type demonstrates the

eastward-increasing proximity to volcanic source rocks from stop 5-9 to here. The rock is composed of about 85-95 percent dacitic detritus, 3-5 percent gray clasts of graywacke, slate, and basaltic andesite, and a small percentage of magnetic and sulfidic fragments. Fragments range in size from a few millimeters to 20 cm. The generally poorly developed sorting and bedding, together with varied clast composition, implies a debris-flow origin.

NEXT:

Travel east on Highway 169, through the city of Tower, and continue east to the village of Soudan. Follow signs to left toward Soudan State Park for approximately 0.3 mi.—at this point, State Park entrance is on your left. Do not follow Park signs, but rather, continue on roadway turning toward the right and follow the road to junction about 0.4 mi. to east. Turn left and follow road for about 0.5 mi., more or less straight past the Soudan Fire Station and up the hill toward the back side of the Park and Stuntz Bay. Disembark at mine buildings and walk about 150 feet north and up-hill to outcrop on the right.



STOP 5-12.
Archean Soudan iron-formation member of Ely Greenstone

No hammering please!

Location: T.62N., R.15W., sec. 27, NE, NE; Soudan Mine State Park.

Soudan 7.5-minute quadrangle.

UTM: 557,120E/5,296,660N

DESCRIPTION:

This classic exposure of the Soudan iron-formation member of the Ely Greenstone lies on the north limb of the Tower-Soudan anticline, and at the stratigraphic top of the volcanic sequences known collectively as the Lower member of the Ely Greenstone. The outcrop displays two generations of tight folding in delicate laminae of chert (creamy white), chert-hematite jasper (red), and magnetite-chert (black to silver-colored). The second generation of folds (F_2) is tectonic in origin, having subvertical axial surfaces that trend east, and steeply plunging axes. Most display Z-asymmetry. The earlier folds (F_{0-1}) appear to have been sharply refolded to produce complex interference patterns. Lundy (1985) studied folding at this locality and concluded that some of the apparent interference structures are the product of early-formed sheath folds that did not involve refolding by D_2. The F_1 structures are predominantly intrafolial, and exhibit a great variety of style and orientation; implying they formed by layer-parallel, soft-sediment slumping.

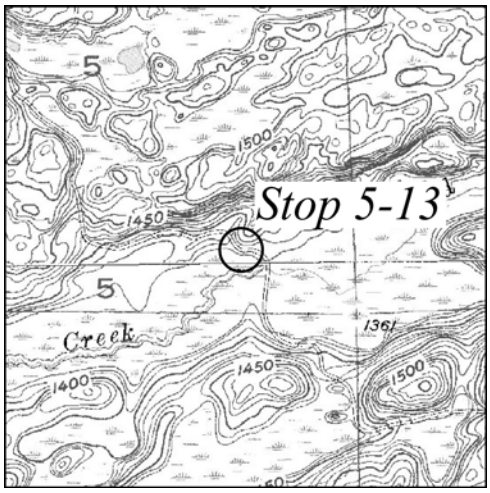
It is interesting to observe the rhythmic microlaminae (1 mm or so thick) in various cherty beds exposed here and speculate about the paleoenvironment—that is, whether these represent daily heating/cooling, tidal, climatic, annual, or some other repetitive influence in the depositional environment. What is known about units of iron-formation in the Ely Greenstone, of which there are many, is that deposition occurred during periods of relative volcanic and tectonic quiescence by the slow subaqueous "rain" of chemical precipitates.

The deep excavations in this area are the early workings of the Soudan iron mine, the first in Minnesota. The mine produced about 16 mt of high-grade hematite ore (60-63 percent iron

content) from 1884 until 1962, when the land was deeded to the State of Minnesota and converted to a park. Most of the production came from underground workings that began here in 1900, and which now can be visited on guided tours. The mine also houses an underground physics research facility at 2340 feet below the surface. A massive expansion of that facility is under consideration to create a national underground laboratory at considerably greater depths (Peterson and Patelke, 2003).

NEXT:

Return to vehicle and drive to east-bound Highway 169. Travel east approximately 7.5 miles to Mud Creek Road. The Highway follows in a general way, the strike of the nearly vertically dipping Soudan iron-formation, bounded by volcanic rocks of the upper member of the Ely Greenstone on the north and the lower member on the south. Turn north (left) on the Mud Creek Road and follow it for approximately 3.8 miles to Mud Creek.



STOP 5-13.
Mud Creek shear zone

Location: T.62N., R.14W., sec. 5, SE, SE; Outcrop just northwest of Mud Creek near road.

Chad Lake 7.5-minute quadrangle.

UTM: 564,230E/5,302,800N

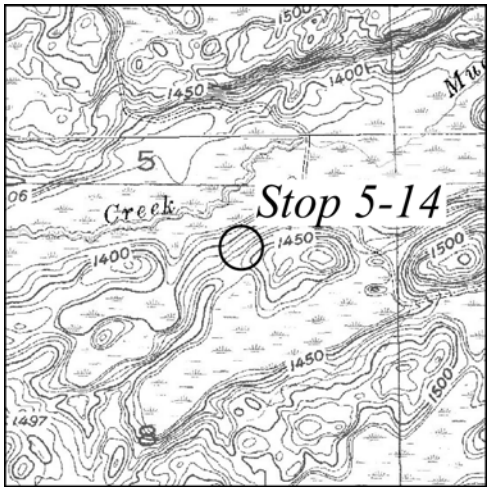
DESCRIPTION:

This outcrop shows highly strained rocks in the Mud Creek shear zone. The rock type is a quartz-iron carbonate-sericite schist, having quartz and tourmaline knots, abundant pyrite, and trace amounts of gold. Its protolith is unknown, because of the intense deformation, but could be any of several rock types in the region, including quartzofeldspathic porphyry, metavolcanic rock, or graywacke. The shear fabric trends east-northeast, and lineations plunge at shallow angles to the east. Development of this shear zone, which occupies most of the valley of Mud Creek, is a product of largely dextral transpressive deformation that has been partitioned into discrete zones, presumably late in D_2 deformation. It is generally believed that gold-bearing mineralization was introduced during these later deformation events, and the Mud Creek shear zone and environs continue to attract considerable attention as a gold target (Field Trip #7).

The Mud Creek shear is a broad, anastomosing zone that forms the boundary between rocks of the Ely Greenstone and Lake Vermilion Formation on the south, and volcanic and iron-formation-bearing rocks known informally as the Bass Lake sequence on the north. The Bass Lake rocks may be equivalent to parts of the Newton Lake Formation exposed north of Ely, but a complex series of faults in the intervening area makes this correlation speculative.

NEXT:

Drive south from Mud Creek to first gravel trail to the west; turn west and proceed about 0.2 to gravel pit.



STOP 5-14 [Optional]
Archean graywacke and slate of Lake Vermilion Formation

Location: T.62N., R.14W., sec. 8, NW, NE.

Soudan 7.5-minute quadrangle.

UTM: 564,100E/5,302,400N

DESCRIPTION:

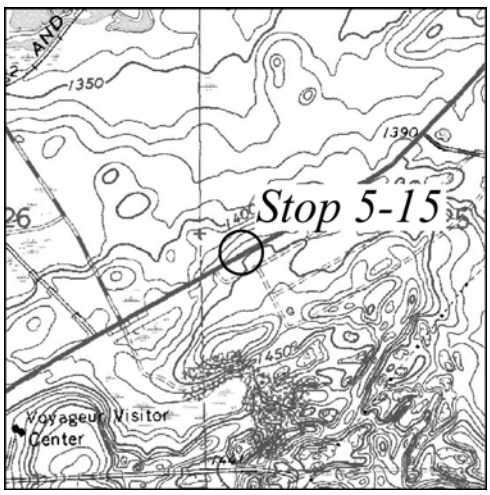
This recently exposed outcrop of the Lake Vermilion Formation consists of siliceous slate having a 1-2 meter thick unit of thinly bedded tuffaceous graywacke. The slate carries a strong S_2 cleavage and is cut by shears; both trending to the east-northeast. Bedding in the graywacke trends north-northeast and youngs to the southeast—this contrasts with the regional setting which strikes to the east-northeast and youngs to the north. The reason for this anomalous trend is unclear, but it may be the result of fault drag along the Mud Creek shear zone.

NEXT:

Return to Mud Creek Road, travel south to Highway 169, then turn east on 169 and continue approximately 12 miles to Ely. [Enroute, one might visit a spectacular outcrop of pillowed flows described in Field Trip #1, Stop 1-13; Hilltop S of Highway 169 at SE, SW, S19, T.62N, R14W, UTM:056200E, 5297800N]. Pass through Ely and continue approximately 1 mile east of town to junction of 169 and County Road 88; then proceed 0.3 mi farther east to road cut.

Ely and vicinity (refer to Figures 5.1 and 5.4)

Geologic framework and outcrop locations are shown on Green and Schulz, 1982.



STOP 5-15 [Optional]
Archean Ely Greenstone

Location: T.63N., R.12W., sec. 25, NW, SW; Road cut on Highway 169 approximately 0.3 mi. east of junction with County Road 88 (Spaulding Bay Road).

Ely 7.5-minute quadrangle.

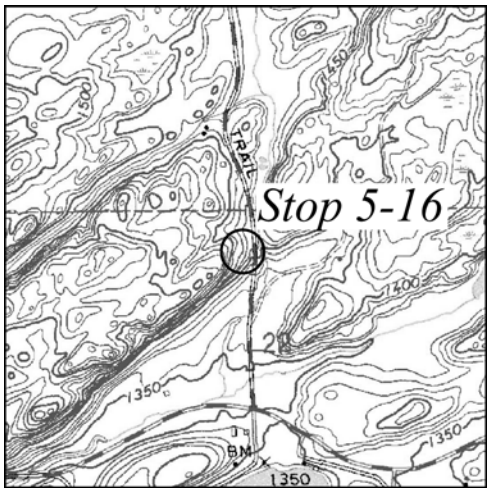
UTM: 588,350E/5,307,075N

DESCRIPTION

Pillowed basalt of the Ely Greenstone. Flows and complexly interlayered iron-formation in the area are strongly folded into anticlines and synclines striking predominantly to the northeast. Stratigraphic younging to the north can be established from pillow structures in this exposure and others nearby. High grade iron ore was mined from iron-formations in the noses or axes of many of the folds. The productive Zenith, Chandler, and Pioneer mines, for example, lie within one such fold, the Ely trough.

NEXT:

Return to junction of Highway 169 and County Road 88, turn north on 88 and travel approximately 2 miles to Echo Trail. Turn right (north) on Echo trail and proceed 0.3 miles north to road cut on west side of the road.



STOP 5-16.
<u>Archean Newton Lake Formation—variolitic basalt</u>

Location: T.63N., R.12W., sec. 22, NE, NW; road cut on Echo Trail.

Ely 7.5-minute quadrangle.

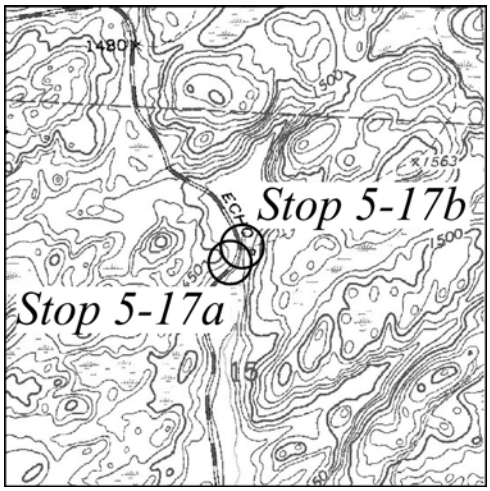
UTM: 585,750E/5,309,360N

DESCRIPTION

This road cut exposes pillowed and variolitic flows and hyaloclastite breccia of the Newton Lake Formation. Pillow sizes vary from "normal" to "mattress", and their shapes indicate younging to the south. The Newton Lake is separated from the Ely Greenstone to the south by a complex zone of faulting (Shagawa Lake and Sibley faults) developed within sedimentary rocks of the Knife Lake Group. Although relatively undeformed conglomerate and sedimentary rocks of the Knife Lake Group are exposed just a few miles to the east, they are typically so sheared and altered in this area as to obscure lithologic and sedimentary interpretations, and will not be visited. The Newton Lake Formation differs from the Ely in that the former contains abundant diabasic sills and rare iron-formation. In addition, lava flows of the Newton Lake typically have larger MgO and incompatible element contents than those of the Ely, and some are classed as komatiites (Schulz, 1980). The Newton Lake Formation (and possibly equivalent Bass Lake sequence) is considered to be the youngest Archean supracrustal sequence in the Vermilion district. Rocks having nearly identical composition and stratigraphic/structural setting occur in Itasca County some 80 kilometers to the west (Jirsa, 1990).

NEXT:

Continue north on Echo Trail for 0.8 mi. to outcrop 5-17A on left (west) side of road.



STOP 5-17.
Archean Newton Lake Formation—felsic tuff and mafic intrusions

Location: T.63N., R.12W., sec. 15 SE, NW; road cuts on Echo Trail.

Ely 7.5-minute quadrangle.

UTM: **5-17A**=585,775E/5,310,660N
5-17B=585,810E/5,310,700N

DESCRIPTION:
5-17A Road cut on west side of Echo Trail is aphanitic to fine-grained, laminated, light-colored felsic tuff and probable reworked tuff, cut by diabase. Minor graded bedding indicates younging to the north; however, it is likely at this exposure that the tuff represents a large block rafted in diabasic intrusions.

NEXT:
Cross to east side of Echo Trail and walk north to first outcrop.

5-17B This road cut exposes layered metadiabase intruded by hornblende-pyroxene-biotite-bearing lamprophyre; both presumed to be intrusions of the Newton Lake Formation. Lamprophyre is coarse- to very coarse-grained and contains inclusions of the diabase. Overall, the Newton Lake Formation contains abundant and locally differentiated sills that vary in composition from diabase and gabbro to pyroxenite and peridotite.

NEXT:
Return to vehicle and proceed north on Echo Trail, which winds around the east shore of Burntside Lake, for approximately 3.5 miles to junction with County road 803 (Passi Road). Turn left (west) on Passi Road and stop.



STOP 5-18.
Archean schist (metagraywacke), paragneiss, and trondhjemite of Quetico subprovince

Location: T.63N., R.12W., sec. 8, SW, NW; outcrop in SW corner of junction Echo Trail and Passi road.

Shagawa Lake 7.5-minute quadrangle.

UTM: 581,660E/5,312,050N

DESCRIPTION:

About one mile south en route to this stop, we passed a sharp, narrow valley that marks the trace of the Burntside Lake Fault. The fault forms a major boundary between comparatively low (greenschist) grade rocks on the south that typify the Wawa subprovince, with high-grade (amphibolite) rocks of the Quetico subprovince on the north. Rocks here include biotite schist, migmatite, and trondhjemite, a part of what's known as the Vermilion Granitic Complex. Despite the amphibolite grade of metamorphism, the schist retains some evidence of graded bedding locally that indicates a graywacke protolith.

NEXT:

Backtrack to Highway 169, turn west and continue to Highway 1 at the east side of Ely. Follow Highway 1 south toward the north shore of Lake Superior, a distance of approximately 50 miles. Though we will not stop, this route takes us out of the Archean Wawa subprovince, crossing parts of the Mesoproterozoic Keweenawan Supergroup, including the Duluth Complex and North Shore Volcanic Group.

North shore and Duluth areas (refer to Figures 5.1 and 5.5)

For regional framework geology, refer to introductory material to Field Trip #2.

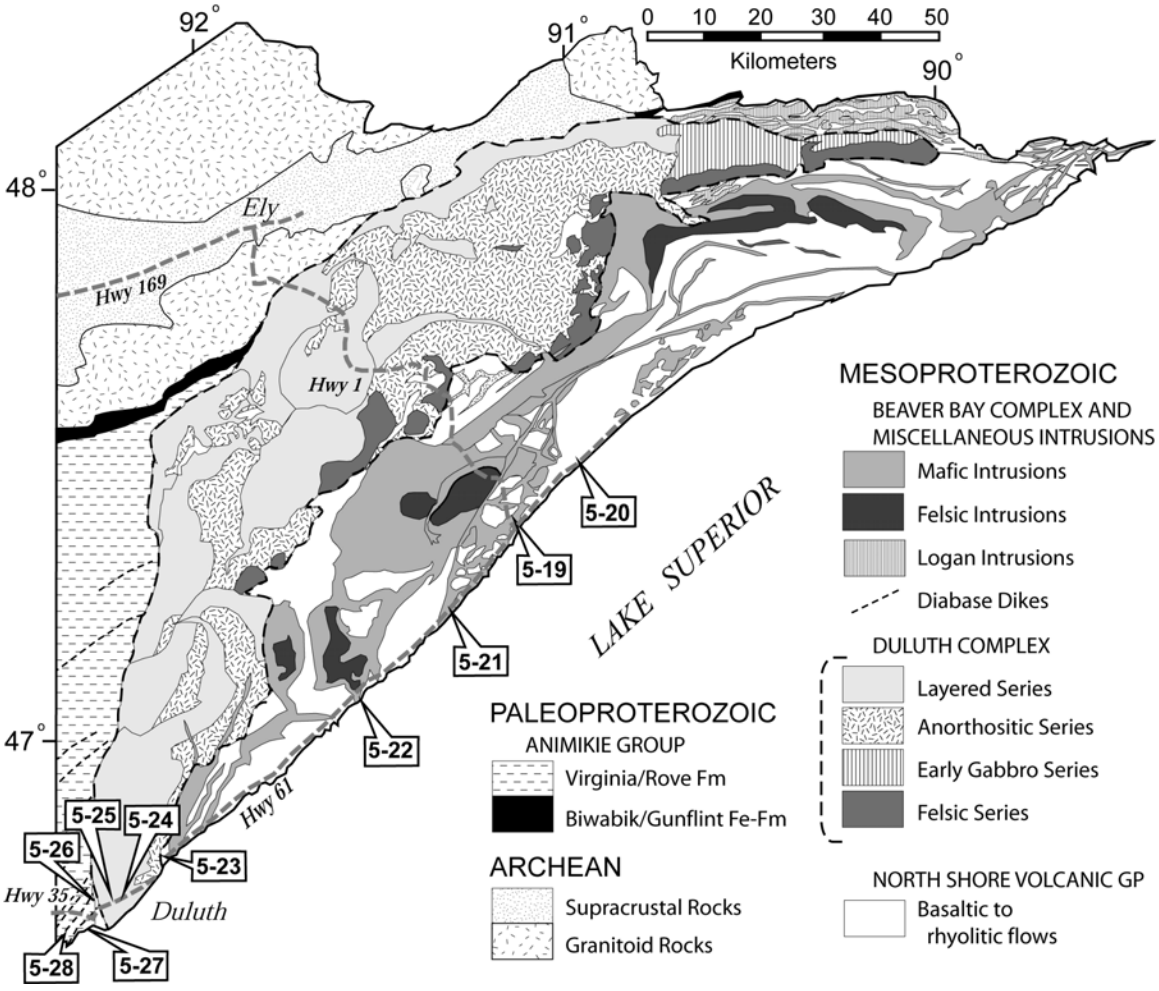
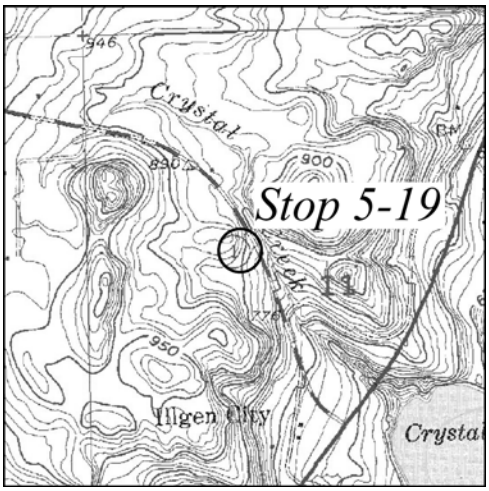


Figure 5.5 Geologic map of northeastern Minnesota (modified from Miller and others, 2002) showing field trip stops on the north shore and in the Duluth area



STOP 5-19.
Top of Mesoproterozoic Palisade rhyolite flow (North Shore Volcanic Group)

Location: T.56N., R.7W., SEC. 11 NE, SW; 0.2 mi. N of Highway 61 on Highway 1.

Illgen City 7.5-minute quadrangle.

UTM: 636,600E/5,245,850N

DESCRIPTION:

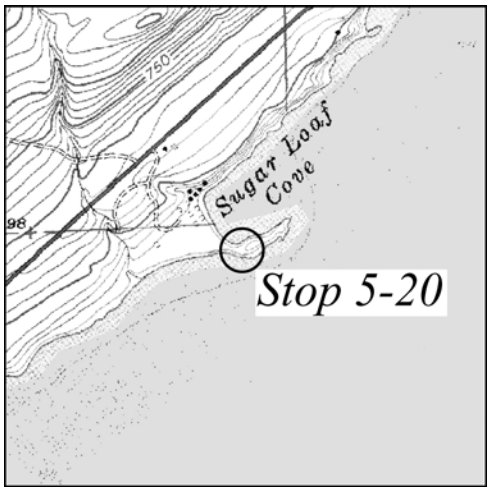
Road cuts on the west side of Highway 1 expose the upper zone of the Palisade rhyolite, a very large and extensive porphyritic flow that is part of the North Shore Volcanic Group. Much of the flow, seen for example on Palisade Head and Shovel Point and the Hwy 61 cut just SW of the junction with Hwy 1, has the appearance of a normal lava flow, but features here at the top, and at the inaccessible base of the cliff at Palisade head, show clearly that it was erupted explosively. The hot ash settled and welded, and the resulting hot, devolatilized mass flowed as "reconstituted lava" or ignimbrite. This road cut in the rapidly chilled top part of the flow is made of flow-brecciated, flow-folded welded tuff; microscopic study reveals deformed shards and fiamme in some samples. (More obvious shards are visible in the chilled welded tuff at the flow's base). Pneumatolytic or hydrothermal alteration has kaolinitized much of the feldspar at this locality. In exposures of the massive flow interior, flow-aligned platy quartz after tridymite is abundant in the groundmass, implying a high temperature of eruption (greater than 870 degrees C).

HISTORICAL PERSPECTIVE:

Little attention was paid to the rhyolites in the North Shore Volcanic Group (NSVG) in earlier accounts. In some of the earliest work by N.H. Winchell (1889), he attributed all north shore rhyolites to the process of melting of sedimentary rocks. More recently, the importance of these rocks has been increasingly recognized because of their abundance compared to intermediate compositions (bimodal tholeiitic suite), and the opportunity they present for geochronologic study (Goldich and others, 1961). With the further development of the U/Pb zircon dating method, many rhyolites have provided age calibration for magmatism in the Midcontinent Rift. The Palisade rhyolite, for instance, has been dated at 1098.6 ±1.7 my (Davis and Green, 1997). Whereas Bowen-inspired models of magma evolution assumed rhyolites such as this to be the products of crystal fractionation of the dominant basaltic parent, their high abundance in the NSVG, along with recent Sm/Nd analyses (Vervoort and Green, 1997) show this and many others to have been produced by partial melting of Archean basement. Finally, detailed field studies and petrography of this and other large felsic flows has indicated that they were erupted at unusually high temperatures for rhyolite, were highly mobile, and that many are rheoignimbrites (Green and Fitz, 1993).

NEXT:

Continue south on Highway 1 for 0.3 miles to Highway 61. Turn northeast (left) on 61 and drive approximately 14 miles to turn-in for Sugarloaf Cove Scientific and Natural Area. Walk to Sugarloaf Point.



STOP 5-20.
Mesoproterozoic basalt at Sugarloaf Cove

Location: T.58N., R.5W., Sec. 29 NE, NE.

Little Marais 7.5-minute quadrangle.

UTM: outcrops begin about
652,020E/5,261,120N

Note: This stop is on property of the Sugarloaf interpretive Center Association (a non-profit) and the State of Minnesota. The beach and point are part of a Scientific and Natural Area (MN DNR) established to preserve these features. Please do not use hammers, or remove samples of any kind. Thanks.

DESCRIPTION:

Many of the attributes of olivine tholeiite lava flows of the NSVG are well exposed on the beach, mainland shore, and point. Flows range from several tens of meters thick to thin flow units of a meter or less. Massive interiors are ophitic: each "ophite" is made of an augite oikocryst enclosing many small plagioclase tablets. Abundant tiny olivine grains occur with plagioclase between the ophites. Pipe amygdules are common in the flow bases, and small vesicle cylinders occur in the massive interior of the thick flow that caps the sequence (The Sugarloaf) on the point. These are all pahoehoe flows, and ropy surfaces are well displayed, as are columnar joints and red sandstone-filled fractures ("clastic dikes") in the interiors.

HISTORICAL PERSPECTIVE:

Although considerable petrographic study and some geochemistry had been done on the basalts of the Cu-bearing lavas of the Keweenaw Peninsula of Michigan (e.g., Butler and Burbank, 1929; Broderick, 1935), the NSVG only began to receive such attention in the mid- to late 1960's, as the world geochemistry community began to focus on basalts and their parent ultramafic rocks of the mantle. The evolution of the plate-tectonics paradigm soon provided a model for the *raison d'etre* for these basalts, and the concept of the Midcontinent Rift System was born (White, 1972; Green, 1977). Further attention to the physical and geochemical aspects of basalts has generated more data on the NSVG (e.g., Nicholson and others, 1997, and references therein); olivine tholeiites such as these are interpreted as being mainly mantle plume melts, with minor lithospheric contribution.

NEXT:

Return southwest on Highway 61 approximately 28 miles to entrance to Split Rock Lighthouse State Park. Turn into the park and walk to base of lighthouse. Note that traveling southwest through this western limb of the NSVG, one traverses down-stratigraphic section.



STOP 5-21. <u>Mesoproterozoic (Keweenawan) Beaver Bay Complex at Split Rock Lighthouse</u>

Location: T.55N., R.8W., Sec. 33 SW, SW; Split Rock Point Lighthouse History Center and State Park.

Split Rock Point NE 7.5-minute quadrangle.

UTM: base of lighthouse at 623,640E/5,228,680N

DESCRIPTION **5-21A**:

Outcrops of fine-grained, ophitic olivine diabase of the Beaver River diabase, which hosts the majority of anorthosite inclusions from Split Rock to Grand Marais, are exposed just northeast of the lighthouse atop a sheer 30-m-high sea cliff. The centimeter-wide augite oikocrysts obvious here are typical of the lower portion of this diabase sill, which dips gently (<15°) into the lake. The prominent point just to the northeast (Rusty Point) is held up by a very large (>200 m) inclusion of medium-grained granophyric granite lying at the base of the sill.

Making our way to the base of the lighthouse, we come upon a good exposure of the large anorthosite inclusion that holds up the entire sea cliff. The coarse-grained (± 1 cm) anorthosite here displays meter-scale modal layering of noritic anorthosite (20% En_{72} hypersthene; 80% An_{60-80} plagioclase) and anorthosite (>99% Pl). This inclusion is altered and displays a moderate to severe cataclastic and granulated texture indicative of having been recrystallized and/or tectonized (Morrison and others, 1983). The steeply dipping layers are cut by thin dikes of medium-grained, igneous-textured augite leuconorite (Pl-An_{56}; Opx-En_{75}; Cpx-En_{66}).

DESCRIPTION **5-21B**.

Follow the footpath south to the pump house on the lakeshore. Here, black, massive basalt is exposed in wave-washed outcrops. The brecciated, amygdaloidal AA flow top of this basalt flow is exposed in the rubbly bluff to the northeast in the direction of the lighthouse. Laminated siltstone in the matrix of the flowtop dips gently toward the lake. Carefully scrambling over some large boulders of ophitic diabase to the northeast, we can view the inclusion-rich base of the diabase sill conformably overlying the basalt flow top. Note that the diabase is chilled against the basalt, but not against the anorthosite inclusions. The abundance of inclusions disrupts the development of regular columnar jointing in the diabase. Numerous boulders along the base of the cliff display the textural and mineralogical varieties of anorthosite inclusions present. Looking to the northeast, one can see that the vertically layered inclusion beneath the lighthouse extends to lake level.

HISTORICAL PERSPECTIVE:

Occurrences of nearly pure plagioclase rock holding up various points of high ground between Split Rock Point and Grand Marias have held a special fascination for Lake Superior geologists for over 150 years. These enormous (up to 500m across), light-colored, coarse-grained masses embedded in dark, fine-grained mafic rock were first recognized by J.G. Norwood in 1849 (Fig. 5.6) as part of the federal Owen Survey (Owen, 1852). During the initial state geological survey (1872-1900), N.H. Winchell noted the resemblance of the feldspar rock, as he called it, to the Rice Point Gabbro at Duluth (anorthositic series of the Duluth Complex). He

suggested that the anorthosites were blocks plucked from older feldspathic phases of the Duluth gabbro by later outpourings of the great gabbro flood, which he fancied to be an enormous extrusive (Winchell, 1900). He noted that the inclusions became somewhat smaller and more dispersed toward the lake and interpreted this to be due to disaggregation as they traveled down-slope (to the southeast) from their source. Midway through the survey, Winchell commissioned A.C. Lawson to study these masses in greater detail. Lawson (1893) concluded that the anorthosite represented the peaks of Archean mountains that where inundated and largely buried by outpouring of the great gabbro flood. The inclusion character evident in most anorthosite occurrences was attributed to spalling of large blocks from the mountain peaks.

In the early 1900's the search for corundum in northern Minnesota followed well publicized discoveries of the mineral in Canada. In 1902, several property owners mistook the light green anorthosite on the north shore for corundum. They formed a company, attracted investors, built a plant complete with processing, storage, and transport facilities---all without ever analyzing the rock! The first sale of mined product was also the last when it was discovered that the rock contained no corundum, but instead is composed of the much softer feldspar mineral, anorthite. Despite the setback, the investors regrouped and eventually formed the Minnesota Mining and Manufacturing company (now 3M); demonstrating an important principal: the lessons learned from small failures sometimes lead to great successes.

One of the most complete inventories of the anorthosites of the North Shore was reported in Minnesota Geological Survey Bulletin 28 "The Geology of the Anorthosites of the Minnesota Coast of Lake Superior" by Grout and Schwartz (1939). In this report, they documented most known anorthosite occurrences in six 1:31,500 scale township maps. In considering the origin of the anorthosite, Grout and Schwartz clearly recognized them as inclusions of pre-existing rock that were derived from a deeper crustal source and were transported in diabasic magma to their present site by virtue of their low density. They agreed with Winchell that the source was likely the anorthositic rocks of the Duluth Complex.

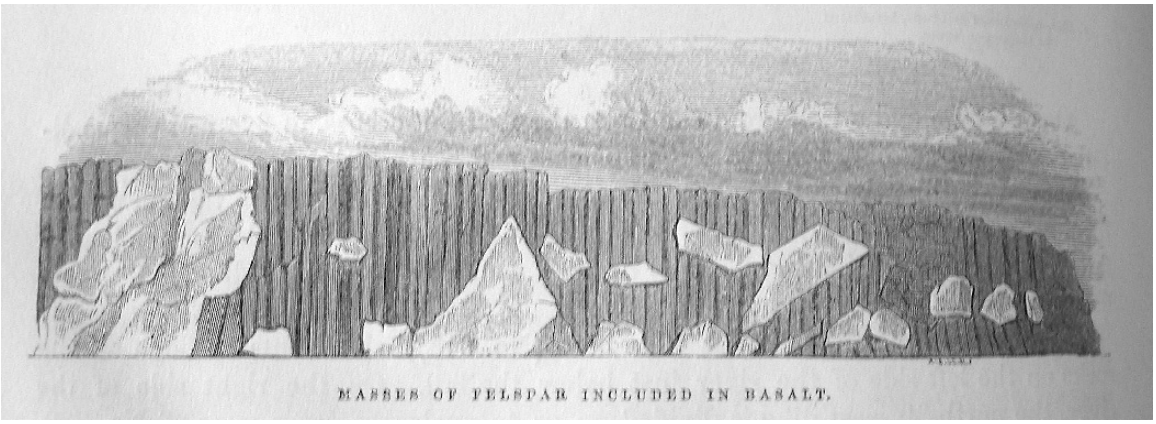


Figure 5.6—Woodcut print from J.G. Norwood's report on the geology of the north shore (in Owen 1852) showing massive anorthosite inclusions in well-jointed basalt.

More detailed considerations of the petrographic and geochemical attributes of the anorthosite inclusions led Phinney (1968) to conclude that Duluth Complex gabbroic anorthosites were an unlikely source for the inclusions. The inclusions, which commonly show signs of severe recrystallization and cataclasis, were more plagioclase-rich (>95%), coarser-grained, and more anorthitic (An_{70-75}) than most anorthositic rocks of the complex. Also, Duluth Complex anorthositic rocks rarely contain orthopyroxene, a common mafic phase in the inclusions. Instead, he speculated that the inclusions were derived from deeper anorthosite bodies in the mid-

to lower crust. Proterozoic anorthosite massifs seemed an attractive source by their depth of occurrence and plagioclase-rich composition; however, massif anorthosites are less anorthitic (An_{45-55}) than even Duluth Complex anorthositic rocks (An_{55-65}) and tend to be strongly recrystallized. In a follow-up petrographic and geochemical (Rb-Sr, Sm-Nd, and REE) study (Morrison and others, 1983), Phinney and his coworkers uncovered evidence of crustal contamination in the anorthositic inclusions from a source older than 1.9 Ma. They concluded from this that the inclusions were derived from early Proterozoic anorthosite bodies formed in the lower crust. As a corollary to a model for the origin of plagioclase crystal mush parent magmas to the anorthositic series of the Duluth Complex, Miller and Weiblen (1990) speculated that these anorthositic inclusions may be Keweenawan in age. In their model, plagioclase-rich magmas were generated by ponding of mantle-derived melts in the lower crust. Under the high pressures of the lower crust, plagioclase will be significantly buoyant in mafic magma and would likely give rise to anorthositic cupolas in these deep magma chambers. Early lower crustal intrusions may have been contaminated by interaction with the lower crust and thus imparted crustal isotopic signatures into these early anorthosite cumulates. As rifting progressed and the lower crust was displaced, later magmas may have picked up these early Keweenawan anorthosites and transported them tens of kilometers into subvolcanic dikes and sills of the Beaver River diabase.
NEXT:

Return to Highway 61, turn southwest (left), and continue approximately 15 miles to Silver Cliff Tunnel.



STOP 5-22 [Optional]
Mesoproterozoic (Keweenawan) diabase and andesitic flows at Silver Cliff Tunnel

Location: T.53N., R.10W., Sec. 15 and 22.

Castle Danger 7.5-minute quadrangle.

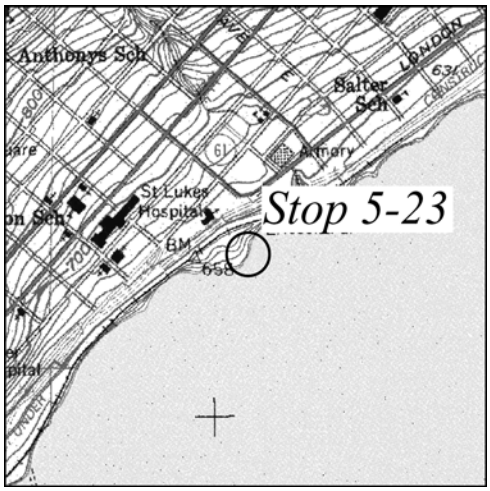
UTM: 606,993E/5,213,952N to
606,755E/5,213,486N;

DESCRIPTION:

The Silver Creek diabase is an irregular, subcordant, subhorizontal intrusion at least 200 feet thick. The diabase forms a prominent highland that projects inland several miles from Silver Cliff at Lake Superior. Excavation of the Highway 61 tunnel has created excellent exposures of the contact between volcanic rocks and both the top and bottom of the diabase, and has exposed a north striking, 55-degree east-dipping fault that cuts the base of the diabase.
Refer to Field Trip 2, stop 2-6 for details.

NEXT:

Continue southwest on Highway 61 for approximately 30 miles. Near the City of Duluth, 61 becomes London Road; continue southwest on London Road. Do NOT turn onto Interstate Highway 35, but clock mileage and continue approximately 1.25 miles to parking area for Leif Ericson City Park near 12th Avenue East. Walk into the park and to the shoreline.



STOP 5-23.
Mesoproterozoic (Keweenawan) interflow sedimentary rocks at Leif Ericson Park

Location: T.50N., R.14W., Sec. 23 SW.

Duluth 7.5-Minute quadrangle.

UTM: 570,050E/5,182,950N

DESCRIPTION:

This classic location exposes an estimated 110' of interflow sandstone, lying on slightly eroded amygdaloidal flows of the gently southeast-dipping Leif Ericson Park Lavas. The sandstone is fine- to medium-grained, moderately well sorted, and derived almost totally from underlying and adjacent lava flows—no extrabasinal detritus has been detected (Jirsa, 1984). The presence of planar-tabular and trough cross-bedding, together with the lenticular distribution of interflow units, implies the strata represent occasional stream flow deposits in depressions on the evolving lava surfaces. This is one of many interflow sedimentary units, most of which are "red bed-like" in appearance (considered in the early works of N.H. Winchell as Potsdam equivalents). By contrast, this interflow lies very near the base of the North Shore Volcanic Group, and thus has experienced burial metamorphism to nearly greenschist facies—though it is possible that some of the elevated grade above the zeolite facies typical of rocks just to the northeast, is an aureole effect of the adjacent Endion and other diabasic sills.

NEXT:

Follow signs to Interstate Highway 35, travel southwest on 35 to Highway 53. Turn north on 53 (Piedmont Avenue) and continue about 1 mile to intersection with Skyline Parkway. Turn left (southwest) on Skyline and travel 0.4 mi. to crossing with 28th Ave. W; continue 0.2 mi. farther southwest on Skyline to outcrop on the right (Stop 5-24).



STOP 5-24 [Optional]
Mesoproterozoic Duluth Complex; Layered Series "Chill", Granophyre, and Anorthositic Series

Location : T50N, R14W, Sec 32, SW, SW;
Skyline Parkway about 1/5 mi. SW of 28th Ave.

Duluth Heights 7.5-minute quadrangle.

UTM: 564,880E/5,179,640N

DESCRIPTION:

Exposed at the northeast end of this roadcut is a fine-grained mafic rock with intermingled granophyre that together cut coarse-grained olivine gabbroic anorthosite of the anorthositic series. Since early mapping by Grout in the Duluth area in 1911, the relationships displayed by this rather innocuous exposure has been critical in interpreting the intrusive history of the Duluth Complex. The distinct, but complexly intermingled domains of mafic and felsic rock displayed here and commonly observed in the upper parts of the Duluth Complex led Grout (1918c) to the interpretation that these rocks formed by liquid immiscibility. This idea was strongly at odds with the newly introduced notion of granite formation by differentiation of mafic magma touted by Bowen (1928) and led to years of debate between these two famous petrologists.

Another significant feature displayed by this exposure of more parochial interest is the relationship between the fine-grained gabbro and the coarse-grained olivine gabbroic anorthosite. The fine-grained mafic rock exposed here can be traced up over the ridge to the west where it merges into the upper contact zone of Layered Series (see Green and Miller, this abstract volume for general summary of the geology of the Duluth Complex at Duluth). Grout and later Taylor (1964) saw this exposure as evidence that the anorthositic series had cooled considerably when the layered series was intruded and thus was considerably older. This paradigm was accepted by all subsequent workers on the Duluth Complex up through the 1990's. Consequently, it came as something of a shock when high resolution U-Pb ages (Paces and Miller, 1993) revealed that the anorthositic series and the layered series were virtually identical in age (within 0.5 Ma relative to the 22 Ma span of Midcontinent Rift magmatic activity). This precipitated a major paradigm shift in the perception of the intrusive relationships between these two series here and throughout the Duluth Complex (Miller, 1992).

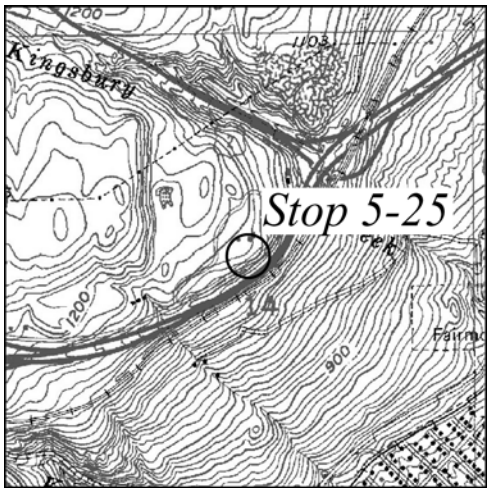
A closer look at this layered series "chill" reveals that it is not a thermal quench of the Layered Series at Duluth (DLS) at all. This rock type is found at the contact with the anorthositic series throughout most of this area and has a remarkably homogeneous, evolved composition (MgO/(MgO+FeO) = ~ 37; see Table 2 in Miller and Ripley, 1996). In thin section, it is a subprismatic biotitic oxide ferrodiorite. Phase equilibrium modeling of its composition indicates that it should be in equilibrium with evolved compositions of augite, ilmenite and plagioclase - phases that comprise gabbroic cumulates found in the cyclic zone and gabbro zones of the DLS. In sum, this rock is much too evolved to have produced the entire cumulate pile of the layered series. Rather than this being a thermal quench, Miller and Ripley (1996) have interpreted this rock to represent decompression quenching of an evolved, water-saturated layered series magma during venting at a time when the cyclic zone was crystallizing. Whereas decompression of less than water saturated magma will result in superheating and a suspension of crystallization (or at least a significant change in phase equilibrium), decompression under water saturated conditions will cause supercooling and quenching. This model fits nicely with the explanation for the cyclic zone with which this composition is apparently comagmatic (see Stop 25).

How does this scenario of venting of a hydrous magma relate to the mafic-felsic duplex which so intrigued Grout? The lobate contacts between the irregular masses of medium-grained granophyre and the fine-grained ferrodiorite host clearly gives the appearance of two-magma mixing. Grout's (1918a) idea that these two magmas formed by silicate liquid immiscibility of hydrous mafic magmas in the roof zone of the Duluth Complex seems more plausible today than when he first proposed it. The concept of silicate liquid immiscibility, which had come to be totally disparaged during Grout's day, has regained some respectability as an plausible, albeit, uncommon petrologic process (Roedder, 1979). However, alternative explanations for this compositional dichotomy can be suggested. One is that these felsic magmas were derived from anatectic melting of various inclusions carried into the layered series chamber. Because of their high silica content and low density, these felsic melts did not readily mix or assimilate with mafic melt, but rather rose to the roof zone where they ponded beneath the anorthositic series cupola.

During magma venting from the chamber, the felsic melts became entrained and irregularly mixed with the mafic magmas. While decompression under water saturated conditions caused rapid crystallization of the mafic magma, the felsic melt became irregularly entrapped in the quenched mafic host and cooled more slowly to a medium-grained texture.

NEXT:

Return to Interstate 35 and follow it southwest to exit for Thomson Hill Rest Area near the crest of the hill west of Duluth. Follow signs to rest area (Stop 5-25).



STOP 5-25. <u>Mesoproterozoic (Keweenawan) Duluth Complex at Duluth</u>

Location: T.49N., R.15W., Sec. 14 SE, NW;
Thomson Hill Rest Area on Skyline Parkway.

West Duluth 7.5-minute quadrangle.

UTM: 560,700E/5,175,380N

HISTORICAL PERSPECTIVE:

The well-exposed gabbroic rocks forming the escarpment above the city of Duluth have long been recognized as the type section of the Duluth Complex. While early surveys recognized the presence of two distinct rock types in the Duluth area - normal gabbros and feldspathic gabbros (Winchell, 1900), Grout was the first to interpret the layered gabbros as a product of convection and magma differentiation (Grout, 1918a-c). Taylor (1964) produced the first detailed-scale (1:24,000) geologic map of the complex in the Duluth area wherein he distinguished the stratiform gabbro cumulates of the layered series from the older, structurally complex anorthositic series. He interpreted the structural complexity of the anorthositic series as having formed from multiple injections of a plagioclase crystal mush. He also recognized the differentiated character of the layered series and pointed out that it was basically similar to the Skaergaard Intrusion - the classic example of a well-differentiated intrusion formed by fractional crystallization of a tholeiitic magma (Wager and Deer, 1939). More recently, detailed mapping and petrologic studies in the Duluth area have added substantially to our understanding of emplacement and crystallization history of these two series here at the type locality of the Duluth Complex (Miller and Ripley, 1996, Miller and others, 2002). See Green and Miller, this abstract volume for more details about the geology of the Duluth Complex at Duluth.

In two outcrops at this stop, we can view exposures that demonstrate the inclusion/intrusion relationship between of anorthositic and layered series rocks and that illustrate the macrocyclic nature of the cumulate rocks which characterizes the medial part of the Layered Series at Duluth, an interval called the cyclic zone.

DESCRIPTION **5-25A**:

In the road cut north of the rest area parking lot is an exposure of intermittently layered, well-foliated olivine oxide gabbro hosting several meter-sized, flattened inclusions of olivine anorthosite. The gabbro is a four-phase cumulate of plagioclase, augite, Fe-Ti oxide and olivine. This cumulate typifies the upper gabbroic parts of troctolite→gabbro macrocycles that

characterize the cyclic zone. In this three-dimensional view, the anorthositic inclusions clearly have a pancake shape that is conformable to layering and foliation in the host gabbro. This may represent the original shape of the inclusions or it may indicate compaction of a partially molten blocks. In either case, the occurrence of anorthositic inclusions in layered series rocks is a ubiquitous feature throughout the Duluth Complex. That the opposite relationship is rarely if ever observed has reinforced the longstanding interpretation that the anorthositic series is older than layered series. However, high precision U-Pb dating (Paces and Miller, 1993) has shown that this age difference is less than 1 million years relative to the 20-million year magmatic history of the Midcontinent Rift.

DESCRIPTION **5-25B**:

In the layered sequence of gabbroic rocks exposed along the 200-m-long roadcut below the observation deck (Fig. 5.7), changes in cumulus mineral assemblages are displayed that characterize the cyclic zone of the Duluth Layered Series. The approximately 1-km thick cyclic zone is composed of five to six major macrocycles, ranging in thickness between 50 and 200 m, within which troctolitic cumulates (Pl+Ol) grade upward into gabbroic cumulates (Pl+Cpx+FeOx±Ol). Boundaries between the macrocycles are defined by abrupt cumulate reversals back to troctolitic cumulates and are also commonly marked by the occurrence of anorthosite inclusions and microgabbro cumulates at the top of the lower cycle. This exposure traverses the boundary between the third and fourth macrocycles.

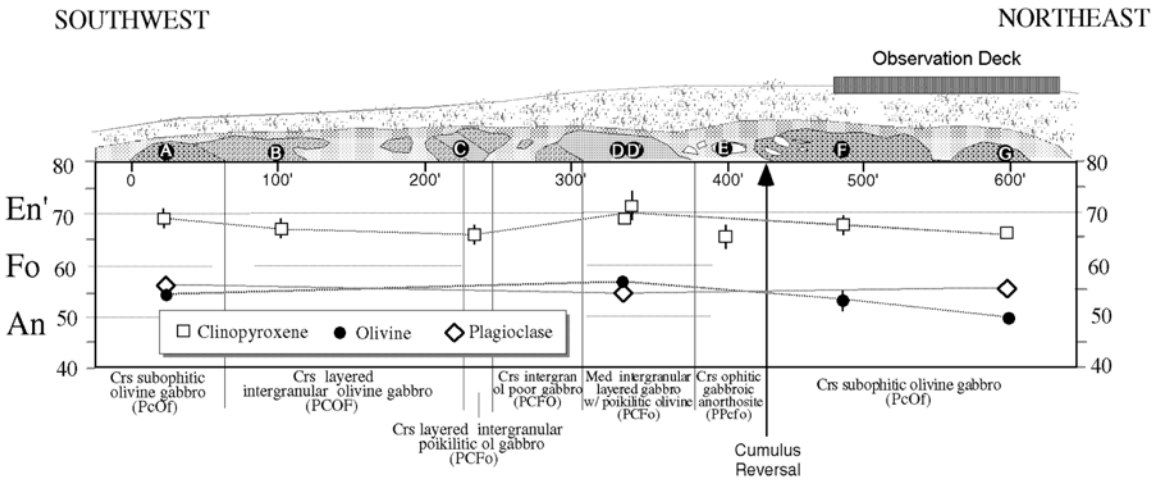


Figure 5.7. Geology and cryptic variation of mineral compositions along Skyline Parkway road cut near Thompson Hill rest area, Stop 5-24B. View is to the north. Dip of lamination and layering is exaggerated; it averages about 20° to the east. Contacts between units are gradational over thicknesses of 10 cm to 1 m. Mineral composition parameters are An in plagioclase = $CaO/(CaO+Na_2O+K_2O)$, En' in augite and Fo in olivine = $MgO/(MgO+FeO)$; all in mole%. Cumulus rock codes (in parentheses) list abbreviations of minerals in decreasing order of abundance (P/p plagioclase, C/c-clinopyroxene (augite); F/f -Fe-Ti oxide; O/o-olivine; upper case = cumulus mineral, lower case = intercumulus minerals). Samples A-G are noted by dark circles.

The west end of the road cut begins with a coarse-grained, moderately laminated, subophitic olivine gabbro, which locally is intergranular and elsewhere is leucocratic (sample A, Fig. 5.7). Augite in this rock is marginally cumulus but becomes definitely so quickly up-section where it consistently has an anhedral granular to subprismatic habit (sample B). Beyond a poorly exposed interval, this gabbro is interlayered with a 2-m-thick interval in which minor olivine becomes

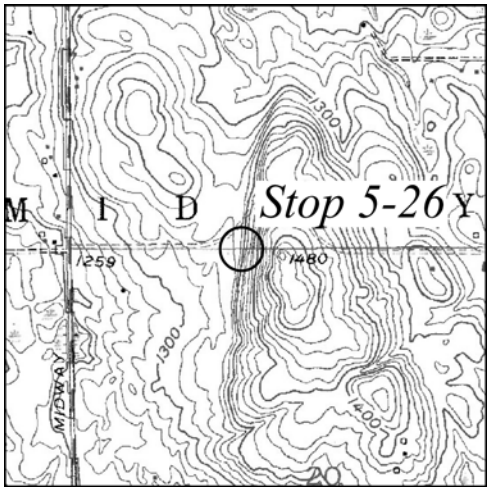
subpoikilitic and concentrated in layers (sample C). Another 3 m above this, the coarse gabbro passes into a medium- to medium fine-grained, well-laminated, subpoikilitic olivine-bearing oxide gabbro (samples D and D') which displays layering of olivine oikocryst concentration and elsewhere isomodal layers rich in Fe-Ti oxide and pyroxene. The very strong foliation and subhedral to euhedral habit of cumulus phases (plagioclase, pyroxene, and ilmenite) impart an adcumulate texture to this rock. Over a poorly exposed interval about 15 m long is an altered, coarse-grained, ophitic gabbroic anorthosite (sample E) that is texturally and mineralogically identical to rocks in the anorthositic series. At the beginning of the next well-exposed section of roadcut, several similar gabbroic anorthosite inclusions are found in a coarse-grained, subophitic to intergranular olivine gabbro (sample F), which gradually grades upward into a more consistently subophitic texture over the remainder of the roadcut (sample G). This rock type closely resembles that at the west end of the roadcut and indicates a downgrading in the cumulus status of pyroxene (and oxide?) and a reemergence of cumulus olivine.

At first glance, a sensible interpretation of this cumulus regression is that it represents a magma recharge event. However, the cryptic variation of En and Fo across the cumulus regression exposed in this section (Fig. 5.7) is the reverse of what would be expected from magma recharge. An alternative explanation to magma recharge is that the textural and compositional variations across this interval reflect decompression of the chamber due to eruption to the surface. Decompression of a volatile-enriched magma would cause supercooling and multiple saturation of the magma and thereby explain the abrupt decreases in grain size and cumulus phase changes without much compositional variation. Magma expulsion through the roof of the layered series would also explain the occurrence of a gabbroic anorthosite inclusion at the cumulate regression. The hydrothermally altered nature of the gabbroic anorthosite is consistent with a volatile-rich environment in the anorthositic series cupola of the Duluth Layered Series magma chamber. The cumulus reversal to ophitic olivine gabbro without an increase in mg# could be explained by repressurization of a devolatilized magma.

NEXT:

Return to Interstate Highway 35 and travel southwest on it approximately 3 miles to Highway 13, Midway Road. Turn north on Midway road and continue 0.8 mi. to road to east (right). Follow road east approximately 0.2 mi., disembark and walk east to base of bluff.

West of Duluth (refer to Figures 5.1, 5.5, and 5.8)



STOP 5-26.
Mesoproterozoic /Paleoproterozoic Unconformity

Location: T.49N., R.15W., Sec. 17 SE, SW.

Esko 7.5-minute quadrangle.

UTM: 555,580E/5,174,380N

DESCRIPTION:

This area contains exposures showing the unconformable relationship between the Paleoproterozoic Thomson Formation (will see at Stop 5-28), and the Mesoproterozoic (Keweenawan) North Shore Volcanic Group. Scattered outcrops in the brushy area just west of the bluff are folded and metamorphosed feldspathic graywacke, siltstone, and mudstone or slate of the Thomson Formation. By contrast the Keweenawan strata exposed in the bluff have been little affected by tectonism since deposition, as shown by their gentle dip to the east. The unconformity is not exposed; but, inferring from the two sets of exposures, it appears to be an angular one. It represents a nearly 800 million-year hiatus, based on an age of roughly 1100 Ma for the Lakeside lavas northeast of here (Davis and Green, 1997) versus a date of about 1880 Ma for tuff interbedded with iron-formation in the Animikie Group, of which the Thomson Formation is a component (Fralick and others, 2002).

The Keweenawan stratigraphic section in this area consists of a basal sedimentary unit, the Nopeming formation, which is conformably overlain by the Ely's Peak basalts. The latter comprise the oldest volcanic strata of the North Shore Volcanic Group. The Nopeming consists of approximately 30 feet (10 m) of interbedded conglomerate and quartz arenite, with minor siltstone beds in the uppermost part of the unit. Much of the sandstone is medium- to coarse-grained, well-sorted and rounded, quartz arenite. The uppermost, silty parts of the Nopeming contain load casts and other structures indicating soft-sediment deformation. The overlying Ely's Peak basalts locally contain well-developed pillow structures, indicating subaqueous deposition. Interestingly, a nearly identical stratigraphic sequence—the Puckwunge Formation—is observed 150 miles to the northeast near Grand Portage (Mattis, 1972). Together, these sequences are representative of a broad depositional and tectonic setting at the onset of Keweenawan deposition in this part of the Lake Superior region.

HISTORICAL PERSPECTIVE:

The stratigraphic position of the Nopeming at the base of the Keweenawan and its correlation with the Puckwunge Formation in the Grand Portage area and the Bessemer Quartzite in the Ironwood, Michigan area has long been accepted. However, a curious complication to this simple interpretation is presented by the 120 year old record of a 500' deep well drilled about 3 miles due south of this site. The log of the Short Line Park well, as reported by Winchell (1889), encountered sandstone and conglomerate, very similar in appearance to the Nopeming that was interlayered with basaltic lavas over a 140-foot interval. Surprisingly, the lower 84 feet of the hole is reported to be exclusively in basalt. This has led some to question whether the Nopeming is the basal unit or just an interflow sedimentary unit (Mattis, 1972). If the latter is correct, it implies a north-south fault along the base of the slope at this site.

NEXT:

Return to Interstate 35, cross under it, and continue south to join with Becks Road. Follow Becks Road southeast to Gary-New Duluth and Highway 23. Turn south (right) on 23 and follow it through town. At the south edge of town, Highway 23 turns west (right); follow it into the city of Fond du Lac. At the west edge of Fond du Lac, turn northwest (right) on Highway 210 and turn very shortly afterward to left into city park. Walk to northwest to edge of bluff (Stop 5-27).

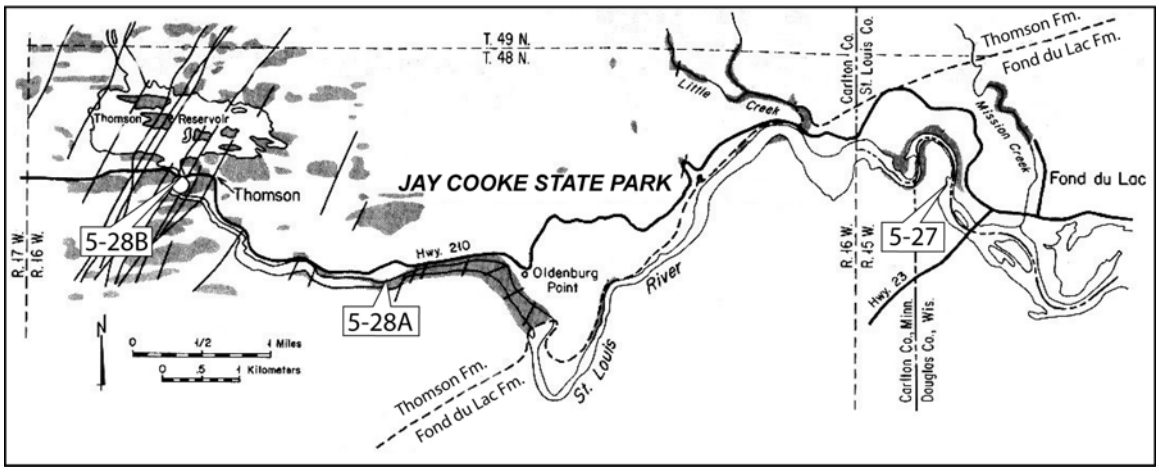
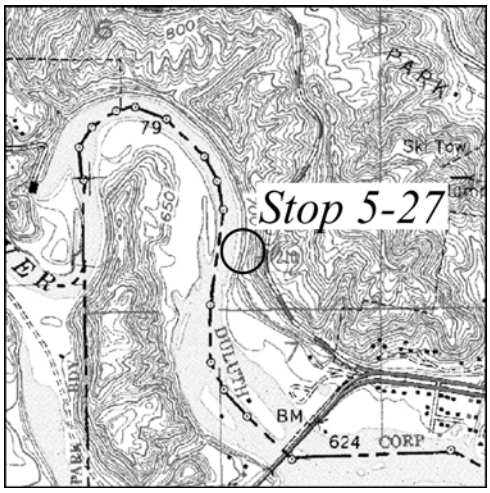


Figure 5.8. Geology and location of stops in the Fond du Lac, Jay Cooke State Park, and Thomson Dam areas (modified from Jirsa and Morey, 1987).



STOP 5-27.
Mesoproterozoic Fond du Lac Formation

Location: T.48N., R.15W., sec. 6 SE, SE along west shore of St. Louis River near city park.

Esko 7.5-minute quadrangle (SEE Fig. 5.8 for location).

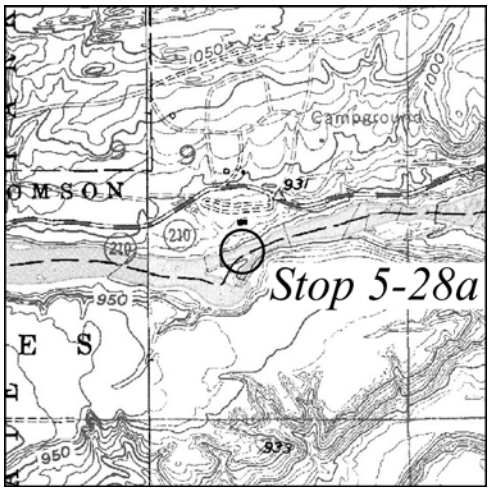
UTM: 554,670E/5,168,140N

DESCRIPTION:

The Fond du Lac Formation consists of nearly 800 feet of red sandstone, shale, and minor basal conglomerate, of which only a few hundred feet are exposed here along the lower St. Louis River. It is part of an accumulation of as much as 20,000 feet of clastic sediments deposited in the central Lake Superior region following Keweenawan volcanism. The Fond du Lac is considered to be equivalent to the Orienta Sandstone of the Bayfield Group in Wisconsin, and the Jacobsville Sandstone of Michigan. In this area (though not visible at this locality), the Fond du Lac unconformably overlies rocks of the Paleoproterozoic Thomson Formation, Ely's Peak basalts, and the Duluth Complex (Morey, 1967). The sandstone consists of poorly sorted arkose to subarkose, containing clasts of quartz, chert, microcline, micas, and minor volcanic rocks; implying a source from both outside and inside of the Midcontinent rift. Depositional structures, primarily trough cross-bedding, indicate transport to the east. These sedimentary structures, and the presence of mud cracks and rain imprints, indicate fluvial deposition by streams that meandered across a broad alluvial plane (Morey and Ojakangas, 1982; see also papers in Ojakangas and others, 1997).

NEXT:

Return to northwest-bound Highway 210 and follow it for about 5 miles to entrance to Jay Cooke State Park. Turn in at main park building area on south side of 210.



STOP 5-28, Location 5-28A
Paleoproterozoic Thomson Formation

Location 5-28A: T.48N., R.16W., Sec. 9 NW, SE; Jay Cooke State Park.

Esko 7.5-minute quadrangle.

UTM: 548,120/5,166,870N

DESCRIPTION 5-28A:

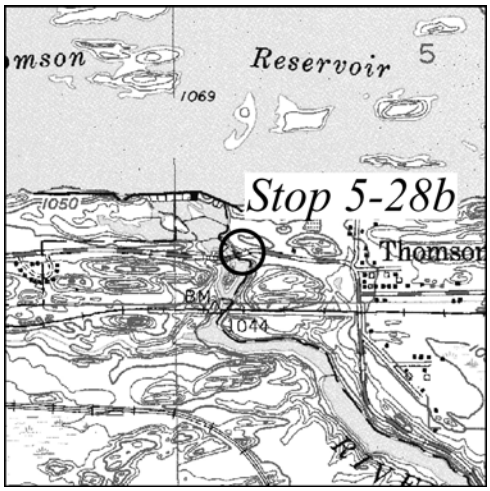
The walking bridge over the river south of the park buildings provides a vantage (depending on river level) to steeply dipping beds of graywacke, siltstone, and slate of the Thomson Formation. The Thomson is the upper stratigraphic unit of the Paleoproterozoic Animikie Group, presumably equivalent with the Virginia Formation discussed in stops 5-4, 5, and 6. Geochronologic data indicate that the Thomson Formation was deposited approximately 1870-1880 Ma (broadly inferred from Fralick and others, 2002), and was deformed during the Penokean orogeny (ca 1850 Ma). As shown in Figure 5.8, the Thomson is unconformably overlain by redbeds of the Mesoproterozoic Fond du Lac Formation, exposed discontinuously just to the east. The unconformity, marked by a basal quartz-pebble conglomerate, can be seen with some difficulty in the river bed at low-water levels near Oldenburg Point, and up Little Creek to the northeast.

HISTORICAL PERSPECTIVE

In the 1950's, the Thomson Formation was considered by a number of authors (e.g., Grout and others, 1951) to be equivalent with the Knife Lake Group in northeastern Minnesota. The two rocks share many of the same structural and lithologic attributes. Most significantly, they both unconformably overlie Archean granitic rocks locally. For clarity, the Thomson is part of a sequence of strata that was known to unconformably overlie "Laurentian" granite. As compelling as this argument is, the Thomson is broadly equivalent to the Virginia Formation on the Mesabi range, with which continuity can now be established from geophysical and drill hole data, and the Rove Formation in the Gunflint district. Thus, the Thomson is now assigned as the uppermost unit of the Paleoproterozoic Animikie Group; and the Knife Lake is interpreted to have formed in part within a successor basin during latest Archean.

NEXT:

Return to Highway 210 and continue northwest approximately 1.6 miles to Thomson Dam.



STOP 5-28, Location 5-28B
<u>Paleoproterozoic Thomson Formation</u>

T.48N., R.16W., sec. 5 SW SW; Thomson Dam.

Cloquet 7.5-minute quadrangle.

UTM: 545,610E/5,168,100N

DESCRIPTION 5-28B:

This is the type locality of the Thomson Formation, represented by approximately 650 feet of strata exposed between the dam north of Highway 210 and the RR bridge south of it. These exposures contain about equal proportions of graywacke, siltstone, and slate; metamorphosed to the greenschist facies. The formation contains abundant carbonate-rich concretions that locally are useful for delineating bedding in otherwise massive graywacke beds. Graywacke units range in thickness from 1 inch (2 cm) to 14 feet (4 m), and commonly display sedimentary structures indicative of turbidite deposition, included graded bedding, cross-bedding, sole marks, flute casts, and flame and ball structures (Morey and Ojakangas, 1970). Cross-bedding indicates flow to the south; though the trends of other structures imply more diverse and localized paleoslope directions.

Structures in the Thomson Formation include gentle to open folds on varied scales, presumably related to deformation during the Penokean orogeny (Holst, 1984). The fold axes trend east, have vertical to steep south dips, and plunge gently east and west. A well developed axial-planar cleavage is present in the slaty beds, and concretions and mud chips are flattened in the plane of cleavage. The cleavage is deformed locally by kink bands. Quartz veins ranging in width from several cm to 3 meters are common in the formation. One of the largest, just north of the Highway 210 bridge, occupies an extensional fracture near the crest of a large anticline. Smaller veins are more contorted, and presumably were folded along with the adjacent rock.

Several dikes of ophitic microgabbro ranging in width from a few inches to 200 feet (65 m) form a dike swarm that occupies northeast-trending joints (Fig. 5.8). The dikes are generally fine- to medium-grained and have chilled, fine-grained margins. Their precise age is unknown; however, their northeast trend and composition implies that they are related to the Mesoproterozoic Midcontinent rift.

END OF TRIP

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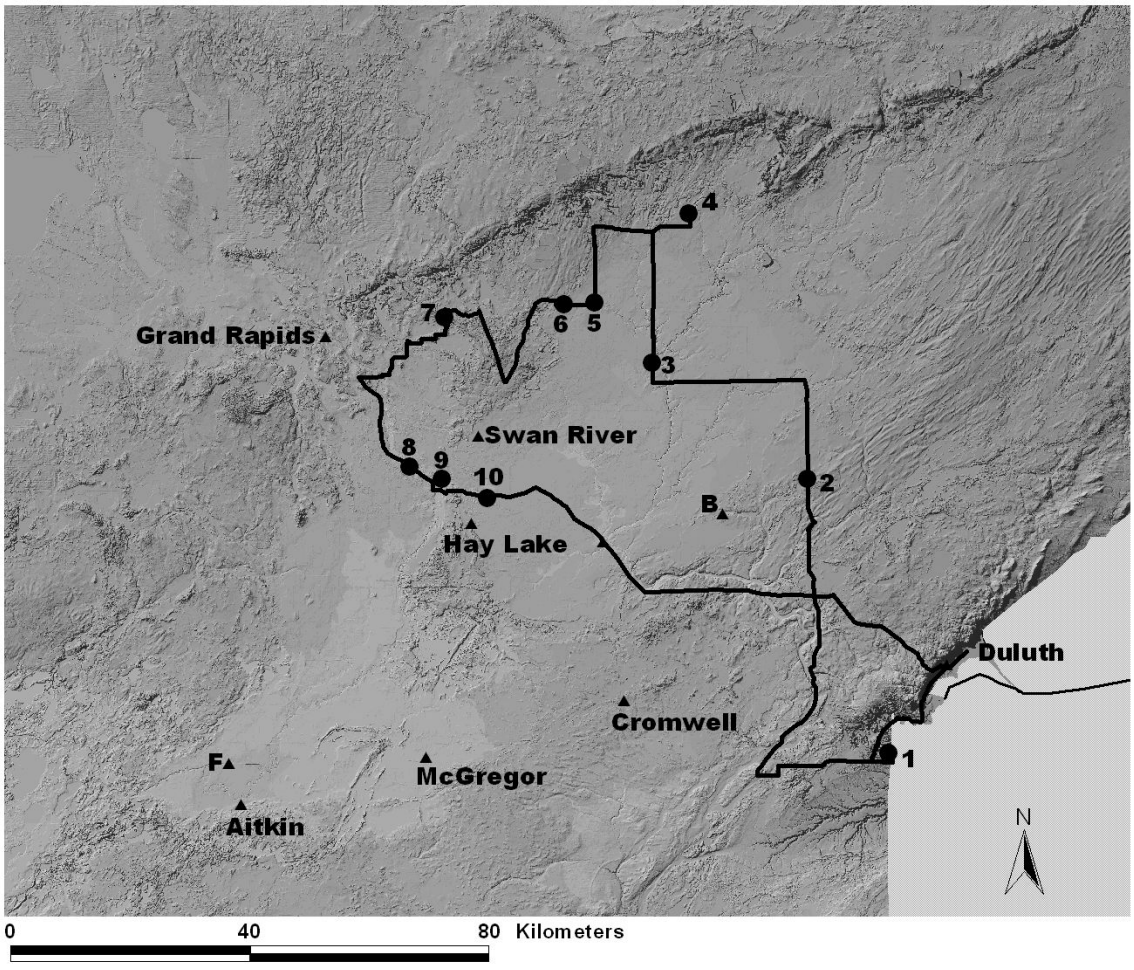
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FIELD TRIP 6

GLACIAL AND POSTGLACIAL LANDSCAPE EVOLUTION IN THE GLACIAL LAKE AITKIN AND UPHAM BASIN, NORTHERN MINNESOTA

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Glacial Lake Aitkin and Upham Road Log with numbered field trip stops, geographic markers and discussion sites: F: Farnum site, B: Baker Spider Creek site, and Hay Lake. Image from 30 meter Digital Elevation Model of Minnesota

GLACIAL AND POSTGLACIAL LANDSCAPE EVOLUTION IN THE GLACIAL LAKE AITKIN AND UPHAM BASIN, NORTHERN MINNESOTA

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INTRODUCTION

Recent investigations into the glacial geology of northeastern Minnesota have considerably revised the glacial chronology. This field trip examines features (landforms and stratigraphy) of the Glacial Lakes Aitkin and Upham basins that record a complex sequence of events related to advance and retreat of the Laurentide Ice Sheet. Deglaciation and formation of glacial Lakes Aitkin and Upham were followed by drainage of the lakes and eolian activity on exposed lacustrine sediments.

Glacial Lakes Aitkin and Upham occupied a basin bounded by the Giants Range to the north and moraines of the Superior and Rainy Lobes and St. Louis Sublobe to the south, east, and west; respectively. As the Rainy Lobe retreated during the Late Wisconsin, meltwater ponded in front of the ice forming proglacial Lakes Aitkin and Upham I. As the Rainy Lobe continued to retreat northeastward across the basin to a margin roughly coincident with the Giants Range, the St. Louis sublobe advanced into the basin (Fig. 6-1). Stagnation and melting of the St. Louis sublobe resulted in formation of Glacial Lakes Aitkin and Upham II (Hobbs 1983). Continued retreat of the Rainy Lobe northward from the Giants Range resulted in formation of Glacial Lakes Norwood and Koochiching, both of which drained southward into Aitkin and Upham II. The evolution of the lakes by a combination of downcutting of the outlets and isostatic rebound is recorded by a series of beaches, wave-cut scarps, and multiple outlets.

Dune formation in the Glacial Lakes Aitkin and Upham II basin was strongly controlled by sediment availability. Dune clusters show a strong correlation with the presence of underlying fine-grained nearshore lacustrine sand. Dune formation was likely episodic, coinciding with periods of rapid lake level lowering and exposure of nearshore sands. Cessation of eolian activity resulted from more gradual stabilization of dunefields. Peaks in eolian activity are indicated at 9.8, 9.3 and 7.4 kyr BP (all dates

given in C^{14} years BP) by the magnetic susceptibility record of Hay Lake, a small lake near the Glacial Lake Upham II shoreline (Figs. 6-2 and 6-3). These peaks in eolian activity may coincide with episodes of drainage of Upham II and Aitkin II.

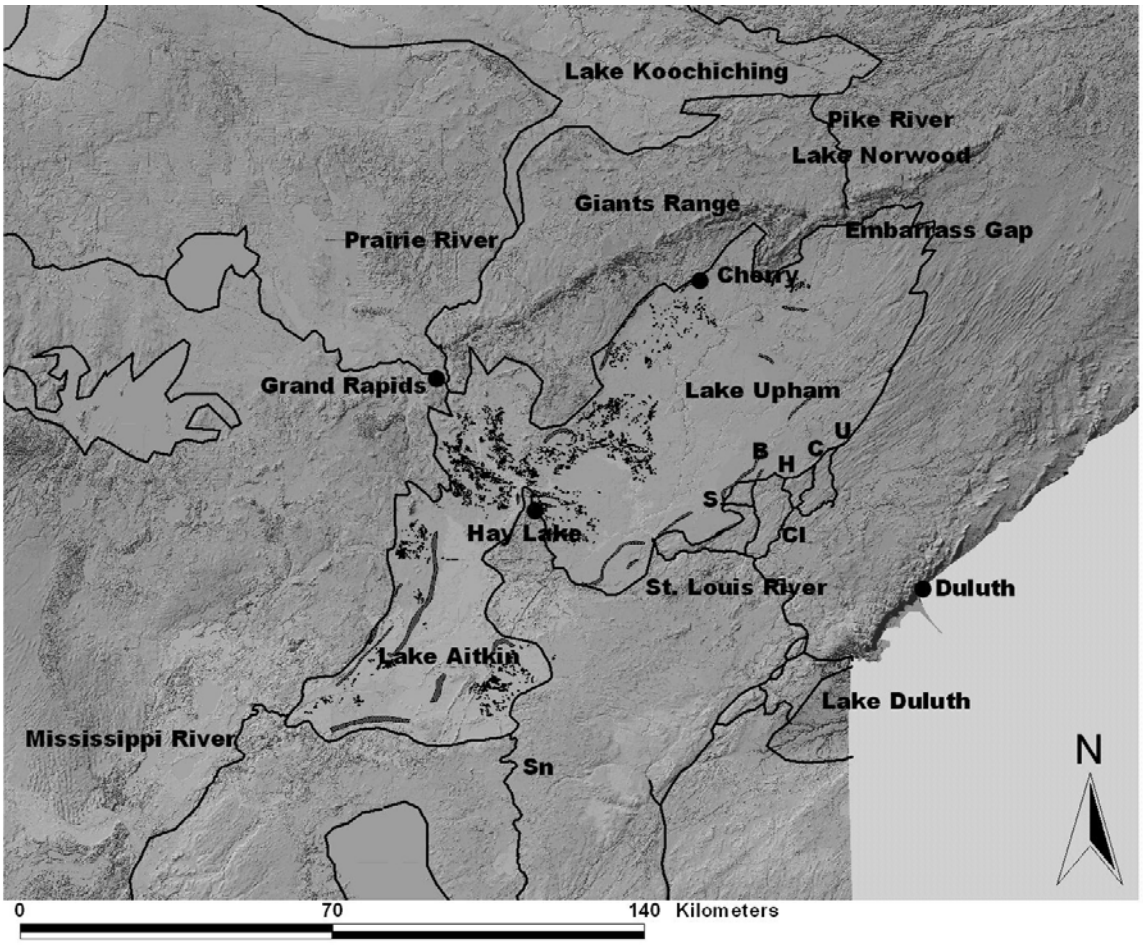


Figure 6-1. The Glacial Lake Aitkin and Upham basin. Beaches are noted by shaded polygons; dunes appear as black dots in the basin; boundaries and channels noted as black lines. Field trip stops will illustrate examples of the following: beach, dune, underflow fan, rhythmites, alluvial fan, Lake Upham sediments in Rainy lobe outwash and a wave-washed esker. Map is modified from Hobbs (1983).

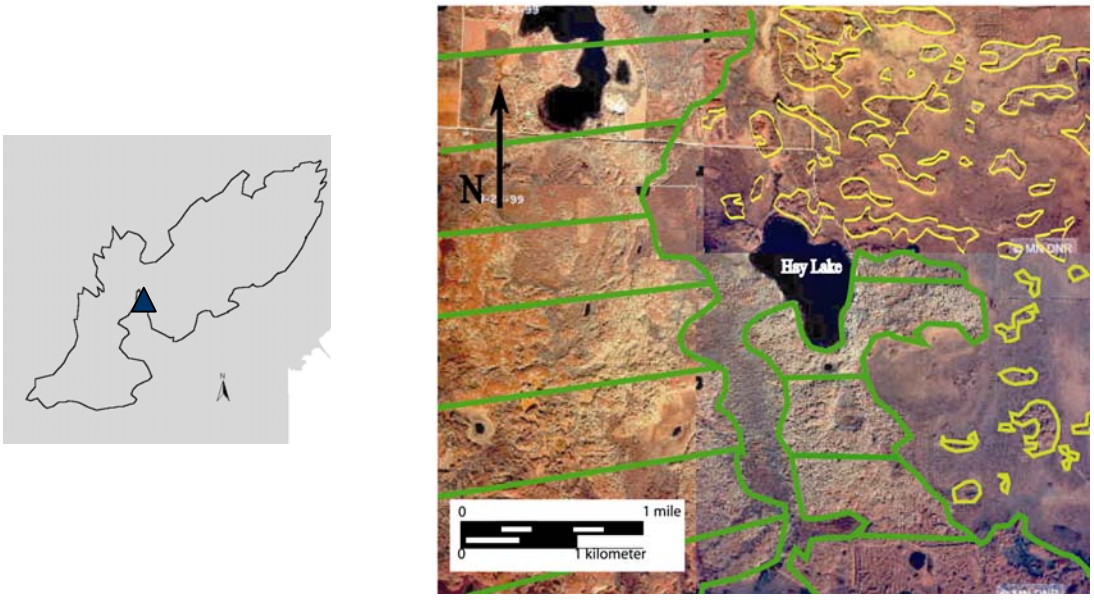


Figure 6-2. Location of Hay Lake. The lined region indicates the Lake Upham upland area; white polygons are dunes. (Vanduse 7.5' USGS quadrangle)

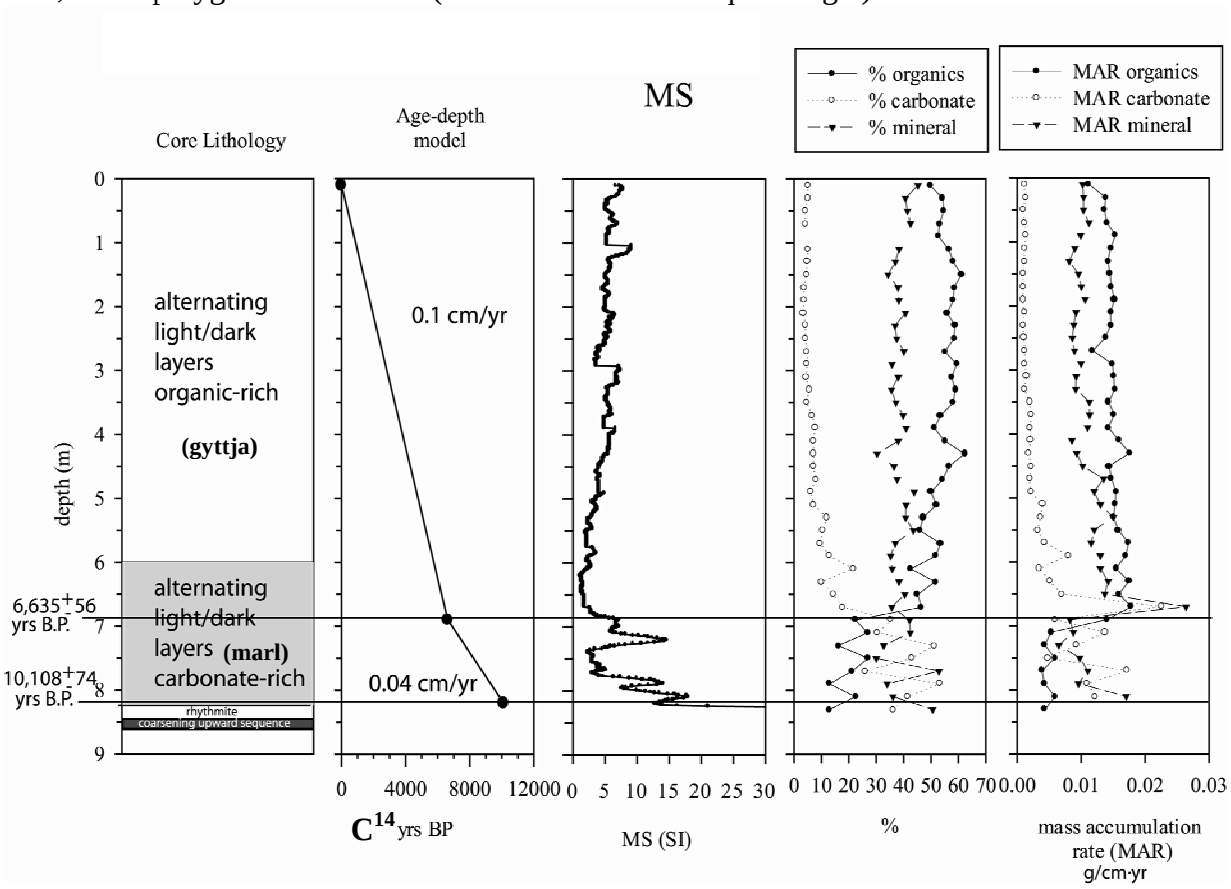


Figure 6-3. Hay Lake sediment core (8.57m): 1. Lithology 2. Age-depth model 3. Magnetic susceptibility 4. Composition 5. Mass accumulation rate.

The dunefields in the Glacial Lakes Aitkin and Upham II basin display similarity to other Quaternary eolian deposits throughout Minnesota and the Midwest (Figs. 6-4 and 6-5). Despite well over a century of research on Minnesota's glacial landscape, these paleodunefields have received relatively little attention. Although mentioned by Upham (1896), Winchell (1896), Hall and Sardeson (1898), Elftman (1898), Leverett and Sardeson (1919) and Leverett (1932), the first comprehensive work on Minnesota's eolian landscape was that of Cooper (1935, 1938).



Figure 6-4. North American dune and loess distribution. (Brady and Weil, 2000)

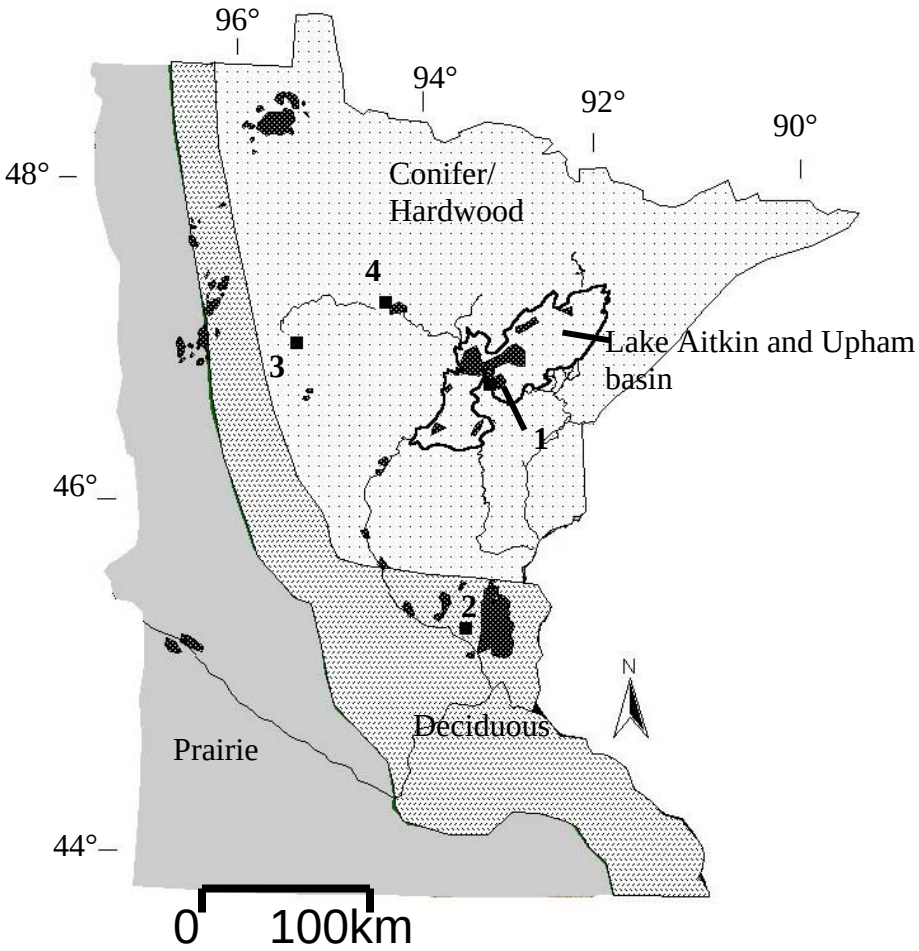


Figure 6-5. Location of sites in Minnesota associated with eolian sedimentation referenced in the text. 1) Hay Lake 2) Lake Ann 3) Elk Lake 4) Lake Winnibigoshish. Filled areas are prominent dune colonies. (Modified from Keen and Shane 1990)

More recently, Grigal and others (1976) examined a dunefield exposed along the southeast shore of Lake Winnibigoshish (Fig. 6-5). They described several buried soil horizons, which yielded radiocarbon dates in the range of 7910 to at least 5040 yr BP suggesting episodic dune formation during the mid-Holocene The formation of these dunes was related to lake-level instability and episodic exposure of littoral areas to wind erosion during the mid-Holocene (Larson and Mooers 2003).

The mid-Holocene climate in the upper Midwest was considerably warmer and drier than the modern climate (Webb et al. 1983; Bartlein et al. 1984; Dean et al. 1984; COHMAP 1988; Dean et al. 1996). The warmer, drier conditions resulted in an eastward shift of prairie ecotone (Fig. 6-5). Lower groundwater levels were accompanied by lake level lowering, particularly in groundwater-dominant closed-basin lakes and ponds, and increased fire frequency and vegetational stress locally resulted in landscape destabilization.

Grigal and others (1976) speculated that the dunes of the Anoka Sand Plain may be mid-Holocene in age, an idea Keen and Shane (1990) explored in detail in their paleoenvironmental investigation of Ann Lake (Fig. 6-5). They concluded that there was a relatively continuous record of eolian activity throughout the mid-Holocene, a reversal of the ideas of earlier workers (Hall and Sardeson 1898; Cooper 1935). Keen and Shane's (1990) whole-core magnetic susceptibility record from Ann Lake records increased silt and fine sand input to the lake from about 8.0 until 4.5 kyr BP with discrete peaks in clastic input at ca. 7.4, 5.8, and 4.9 kyr BP. However, it is difficult to distinguish whether this sediment was exported to the lake directly from active dunes on the shore, eroded from the shoreline due to lake level instability, or deposited suspended airborne sediment. Despite these ambiguities, they concluded that the increased clastic input to Ann Lake was due to landscape destabilization and widespread dune activity on the Anoka Sand Plain.

The distinction between lake level lowering and landscape destabilization as triggers of eolian activity is subtle but important. Although the prairie ecotone shifted eastward during the mid-Holocene, this in itself did not create an environment conducive to widespread landscape destabilization. At present in the central Dakotas there are numerous locations with ample sediment availability, but no widespread dune formation because of the stabilizing vegetation of the prairie environment. Landscape destabilization and dune formation in the existing prairie

environment occurs by excessive drying such as in the Nebraska Sand Hills (Winspear and Pye 1996). Viewed in this context, while Grigal and others (1976) obtained dates from paleosols buried by eolian events, the Keen and Shane (1990) and Dean (1997) magnetic susceptibility records may merely be a record of dryer, dustier conditions during the mid-Holocene rather than pervasive local eolian activity.

Today the Glacial Lakes Aitkin and Upham basin is patchwork of small 'islands' occupied by upland vegetation formed on stabilized dunes. The 'islands' are interrupted by vast areas of peatland developed on the poorly drained, low-relief lakebed composed of fine-grained lacustrine sediment. Locally, scattered topographic highs are underlain by older glacial sediments predating formation of the lakes.

STOP 6 – 1. Glacial Lake Duluth underflow fan. (NW¼ SW¼ Sec. 30, T48N, R15W, Carlton County, *Frogner 7.5' USGS quadrangle; 553728E, 5161950N, NAD83*)

This location preserves the record of several glacial events that affected the western Lake Superior region and provides insight into their relative chronologies. Here, red Glacial Lake Duluth clays are overlain by a subaqueous fan deposited by meltwater from Glacial Lake Aitkin and Upham II (Fig. 6-6). These in turn are overlain by Superior Lobe till deposited by ~~of~~ the Marquette advance. These relationships indicate meltwater drained from Glacial Lake Upham into Glacial Lake Duluth (Table 1) (Mooers and Lehr 1997; Clayton and Moran 1982; Wright 1972; Wright et al. 1973).

During the last phase of Glacial Lakes Aitkin and Upham II, the two lakes were thought to have separated and drained through different outlets; Upham through the St. Louis River and Aitkin through the Mississippi River (Hobbs 1983). A digital elevation model (DEM) reconstruction correcting the basin for isostatic rebound confirms the separation of the two lakes (Marlow 2004). The Mississippi River outlet to Aitkin is a relatively narrow passage through collapsed ice-cored terrain. It is unlikely this underdeveloped outlet carried significant meltwater. Final drainage of Aitkin therefore occurred after significant meltwater contributions to the lake basins had ceased.

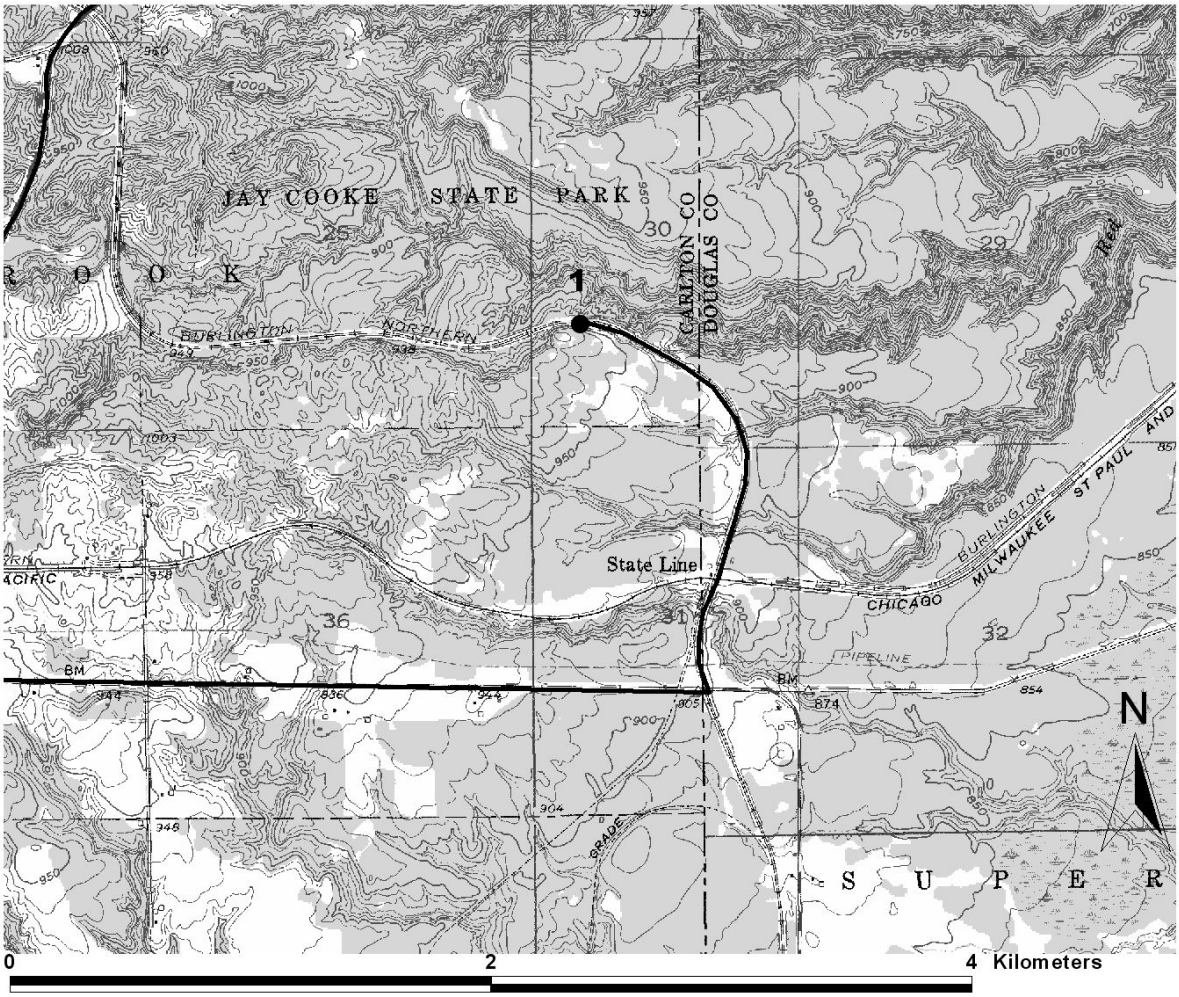


Figure 6-6. Stop 6-1. Underflow fan at the mouth of the St. Louis River drainage in Glacial Lake Duluth basin. The sediments were deposited by meltwater from Glacial Lake Upham II.

Table 6-1. The glacial chronology of northeastern Minnesota outlined resulted from a compilation of several resources (Mooers & Lehr, 1997; Larson (unpublished); Marlow, 2004; Bjork, 1990; Wright, 1972; Clayton & Moran, 1982; Hobbs, 1983; Fenton, 1983; Clayton, 1983; Lehr & Hobbs, 1992)

Phase	Moraine	Lobe	Glacial Lakes			Other events	C^{14} Date (kyr BP)
			Aitkin II			drainage of Aitkin II	
Marquette		Superior					9.9
				Upham II	Duluth	drainage of Upham II	
						cessation of meltwater inflow to Aitkin and Upham II	~11.7
Vermilion	Vermilion	Rainy				L. Duluth underflow fan	12.0
						Inflow from Prairie River	
Big Rice	Big Rice/ Wampus	Rainy					
						St. Louis River outlet	
						Inflow from Embarrass gap	
Nickerson	Nickerson	Superior				Us-Kab-Wan-Ka/Chicken/ Hellwig/Birch/Spider outlets	
						formation of Goodland esker	
North of Nashwauk	Allen	Rainy					~13.0
Alborn	Culver	St. Louis					
Split Rock	Cloquet	Superior	Aitkin I	Upham I			
	Sandy Lake	Rainy					
Automba	Outing	Rainy					
St. Croix	St. Croix	Rainy					~15.5-16

STOP 6 – 2. <u>Hellwig Creek outlet and Glacial Lake Upham II basin overlook</u>. (SE ¼ SE ¼ Sec. 34, T53N, R17W, St. Louis County, *Canyon 7.5' USGS quadrangle*; *540174E, 5207618N, NAD83)*

This stop is near beaches and outlets that existed during the earliest phases of Glacial Lake Upham II, and provides a good overlook from the Upham shore into the basin (Fig. 6-7).

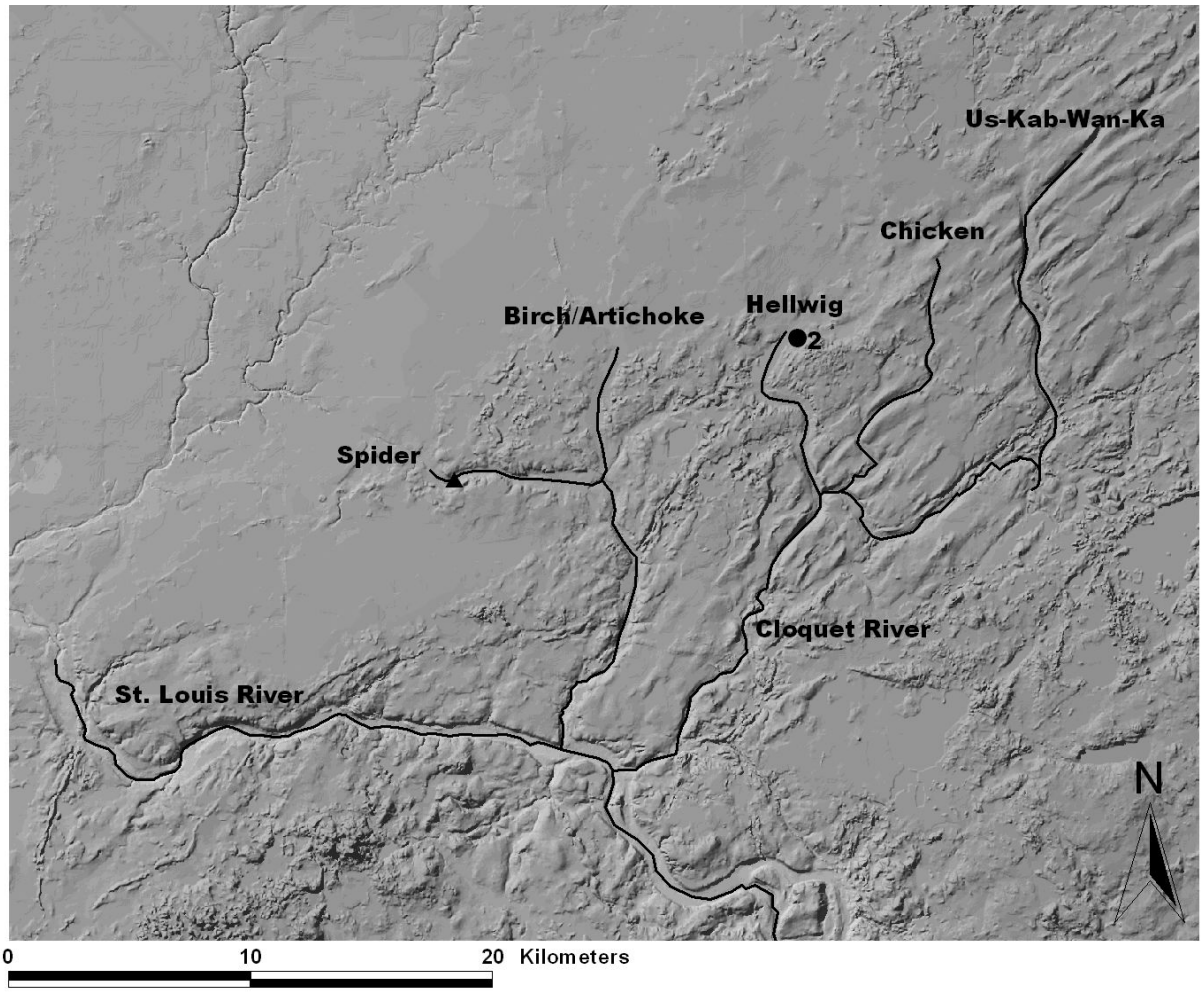


Figure 6-7. Stop 6-2. Location of early outlets to Lake Upham; Triangle marks location of core taken by Baker (1965) containing marl. Circle indicates stop 2 site.

Following stagnation of St. Louis Sublobe ice meltwater flowed through a combination of surface and englacial channels toward the ice margin then along the ice margin through a chain of small proglacial lakes.

Among the earliest and highest outlets in the Upham basin from which these small lakes drained were the Us-Kab-Wan-Ka River (427 m; 1400') and Chicken Creek (434 m; 1425') channels. These channels drained to the southeast along the margin of the Superior Lobe eventually reaching the St. Croix River (Hobbs, 1983). Proglacial lakes and subglacial meltwater in the Aitkin basin drained southward through the Snake channel (381 m; 1250') to the St. Croix River and perhaps southwest to the Mississippi River (378 m; 1240') (Hobbs 1983). All of these outlets lie above the highest main beaches of Glacial Lakes Aitkin and Upham II, and the DEM correction for isostatic rebound indicates ice must have been present in the basin for them to function (Marlow 2004). Therefore they were likely associated with small proglacial lakes rather than the main stages of the lakes.

After the St. Louis Sublobe advance, the margin of the Rainy Lobe retreated north of the Giants Range resulting in the ponding of water between the retreating ice and the Giants Range, forming Glacial Lake Norwood (Fig. 6-8) (Winchell 1901; Hobbs 1983). Glacial Lake Norwood drained south through the Embarrass Gap entering Glacial Lake Upham II at an elevation of 436 m (1450') (Hobbs 1983). This was the first major meltwater inlet to Glacial Lakes Aitkin and Upham II. During this time a significant amount of stagnant ice remained in the basin, and it is possible that Aitkin and Upham were separated by stagnant ice along the Swan River Sill (Fig. 6-8).

As the Rainy Lobe continued to retreat northward, Glacial Lake Norwood increased in extent. Downcutting in the Embarrass Gap lowered the level first to 436 m (1430') and eventually to 427 m (1400'). This lower stage is known as Glacial Lake Koochiching (Nikiforoff 1947; Hobbs 1983; Lehr and Hobbs 1992; Leverett 1932).

While Glacial Lake Norwood was in existence, the upper strandlines (411 to 421 m; 1350-1380') in the extreme northeastern part of Glacial Lake Upham II formed. These levels of Upham likely correspond to outlets at Hellwig Creek (409 m; 1330'), Birch (406 m; 1320'), and Spider Creek (400 m; 1300'). A combination of isostatic rebound and downcutting eventually led

to successive abandonment of the Hellwig Creek, Birch, and Spider Creek outlets and lake level drop.

Baker (1965) reported a bulk radiocarbon date of 13,000±400 (W-1234) yr B.P from a sequence of lacustrine marl within the Spider Creek outlet (Fig. 6-7; and "s" in Fig. 6-8). The marl must post-date the cessation of drainage through the channel since 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 guide. 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 kyr B.P.

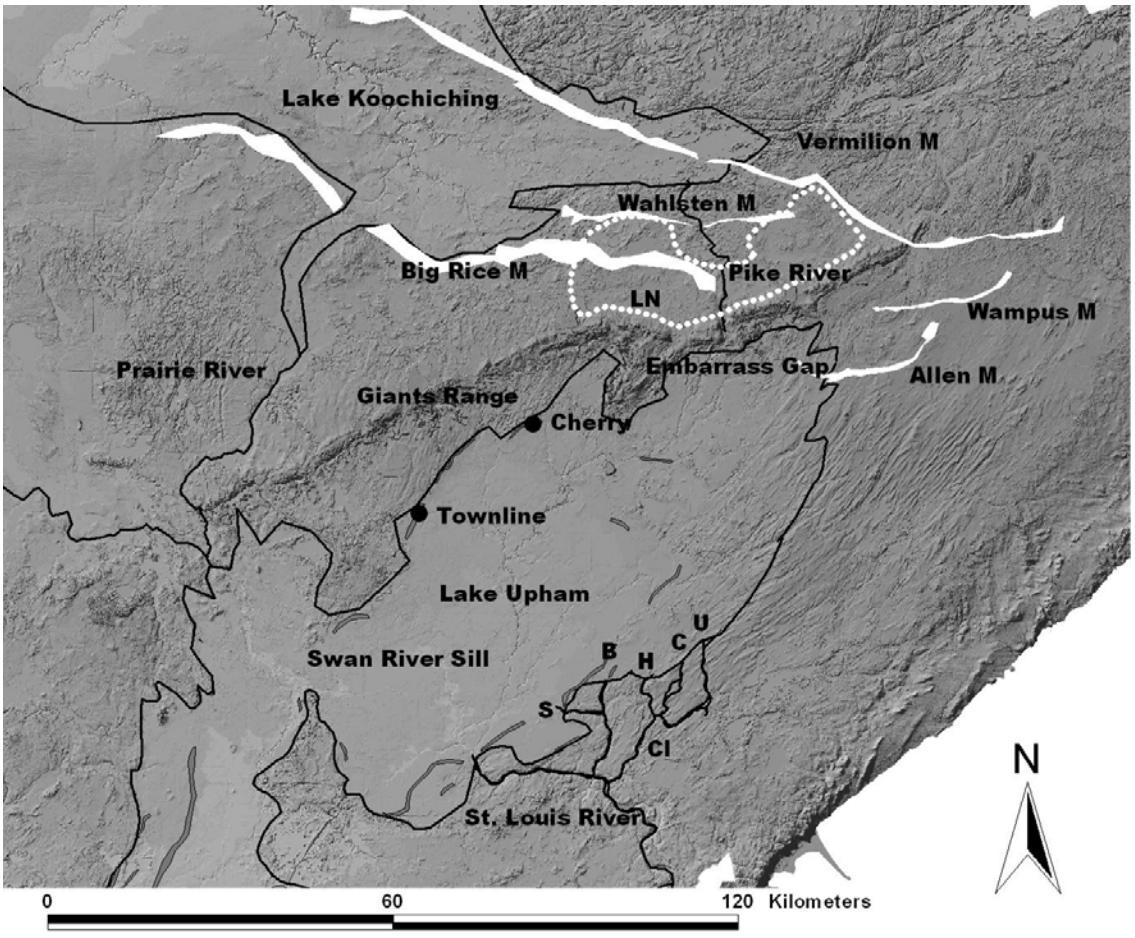


Figure 6-8. Significant features associated with the recession of the Rainy Lobe after the St. Louis sublobe advance. Glacial Lake Norwood (LN) extent is indicated by a dashed line.

STOP 6 – 3. <u>Wave-washed esker island</u> (NE¼ NE¼ Sec. 1, T54N, R20W, St. Louis County, *Toivola 7.5' USGS quadrangle, 514367E, 5226982N, NAD83*).

There are many landforms within the Glacial Lakes Aitkin and Upham II basins that predate development of the lakes. This exposure is an example of an esker deposited during retreat of the Rainy Lobe that was later modified by waves. (Fig. 6-9). 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 gravels. The gravels contain abundant northeast-provenance material and locally derived mudstone and shale from the Paleoproterozoic Virginia Formation and perhaps younger Cretaceous strata. They were deposited by a beaded esker system during retreat of the Rainy Lobe. Overlying the gravels is a fine-grained till deposited by the St. Louis Sublobe. On top of 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 m (1200') elevation. The uppermost portion of the sequence is a blanket of eolian sediment exported from the surrounding lake plain to the upland after final drainage of Upham.

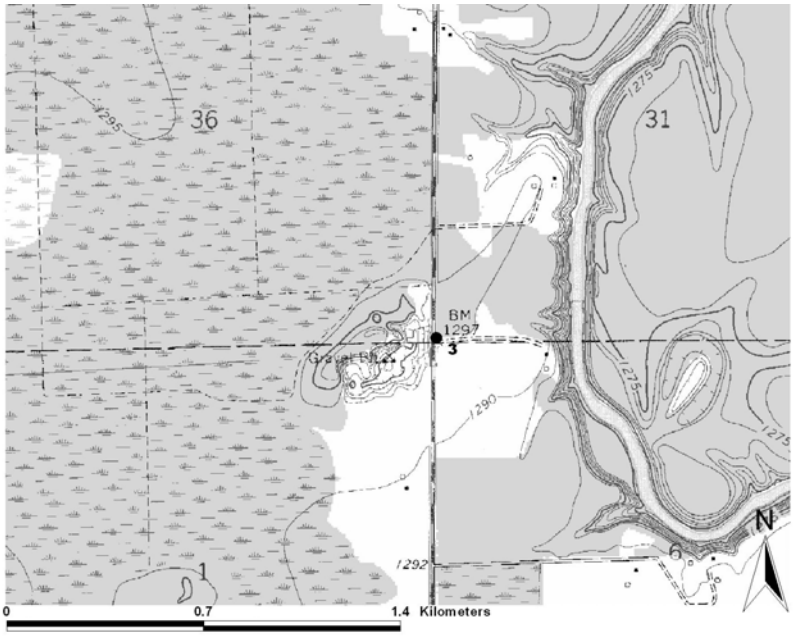


Figure 6-9. Stop 6-3. Wave-washed esker island in Glacial Lake Upham II.

STOP 6 – 4. <u>Rainy Lobe collapsed outwash, Glacial Lake Upham I sediments, and St. Louis Sublobe till</u> (NE¼ SE¼ Sec. 15, T57N, R19W, St. Louis County, *Kirk 7.5' USGS quadrangle; 520373E, 5251907N, NAD83*)

The Glacial Lakes Aitkin and Upham basins were occupied by glacial lakes on two separate occasions during the Late Wisconsin glaciation. The retreat of the Rainy and Superior Lobes from their maximum positions formed a series of ice-cored recessional moraines bordering the lake basins to the south and west leading to ponding of water and formation of Glacial Lakes Aitkin and Upham I (Wright 1972; Hobbs 1983; Lehr and Hobbs 1992). The initial retreat of the Rainy Lobe from the Mille Lacs and Outing moraines (Mooers 1988) led to the formation of Aitkin I (Fig. 6-10). Continued retreat of the Rainy Lobe from the Sandy Lake moraine led to the formation of Upham I. Although the exact time of formation of Aitkin and Upham I is not known, it occurred after the Rainy Lobe retreated from the St. Croix moraine about 15.5 kyr B.P. (Clayton and Moran 1982; Mooers and Lehr 1997), and before the St. Louis Sublobe advanced into the Aitkin and Upham basins from the northwest.

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. 6-11 and 6-12).

The relationships visible in this exposure indicate that the St. Louis Sublobe advanced while a substantial amount of Rainy Lobe ice was still present south of the Giant's Range.

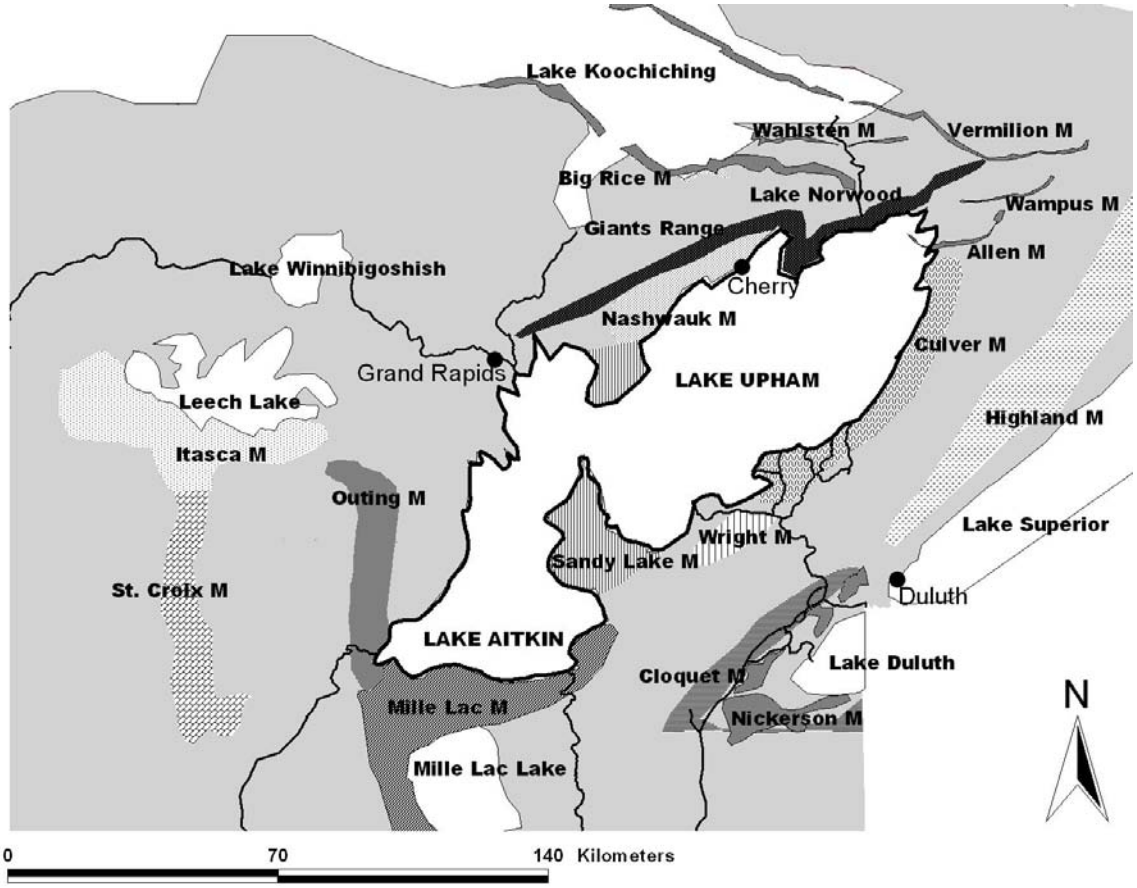


Figure 6-10. Prominent moraines in the Glacial Lakes Aitkin and Upham basin with some modern features included. Modified from Mooers (1988) and Lehr and Hobbs (1992).



Figure 6-11. Subaqueously deposited Rainy Lobe outwash exposed in a gravel pit near Cherry, MN.



Figure 6-12. St. Louis sublobe till containing Glacial Lake Upham I sediment.

STOP 6 – 5. <u>Eolian dunes</u>. (SW¼ SW ¼ Sec. 31, T56N, R20W, St. Louis County, *Riley 7.5' USGS quadrangle, 504626E, 5237098N, NAD83*)

Exposed in the sand pit is a small (<2 m) dune composed of characteristic 4-φ sand (62.5 microns or 230 mesh). Most of the sand exposed in the pit is massive and structureless. However, relict cross bedding can be observed in small patches near the base of the sequence. Primary sedimentary structures in the upper portion of the exposure have been obliterated by bio and crioturbation.

Clusters of eolian dunes are widely distributed throughout the Glacial Lakes Aitkin and Upham basins. They formed as Glacial Lakes Aitkin and Upham II incrementally drained, exposing areas of littoral sediment to wind erosion. Dune heights range from 1 to 5 meters and are composed of ~~a~~ characteristic fine to very fine grained sand (Marlow 2004). They occur as parabolic or longitudinal dunes, but are commonly distorted in morphology as a result of forming in a variable hydrologic environment (Fig. 6-13) (Bagnold 1941) There are a large number of longitudinal dunes and elongate clusters of dunes oriented in a NW-SE direction, suggesting they formed under prevailing NW winds (Fig. 14). Bagnold's (1941) description of dunes indicates that they "tend to occur in belts or chains, whose direction coincides with that of the resultant long-period

sand vector Q", with Q being sand flow because of the sum of the strong and gentle wind directions, and the width of the belt at right angles to Q."

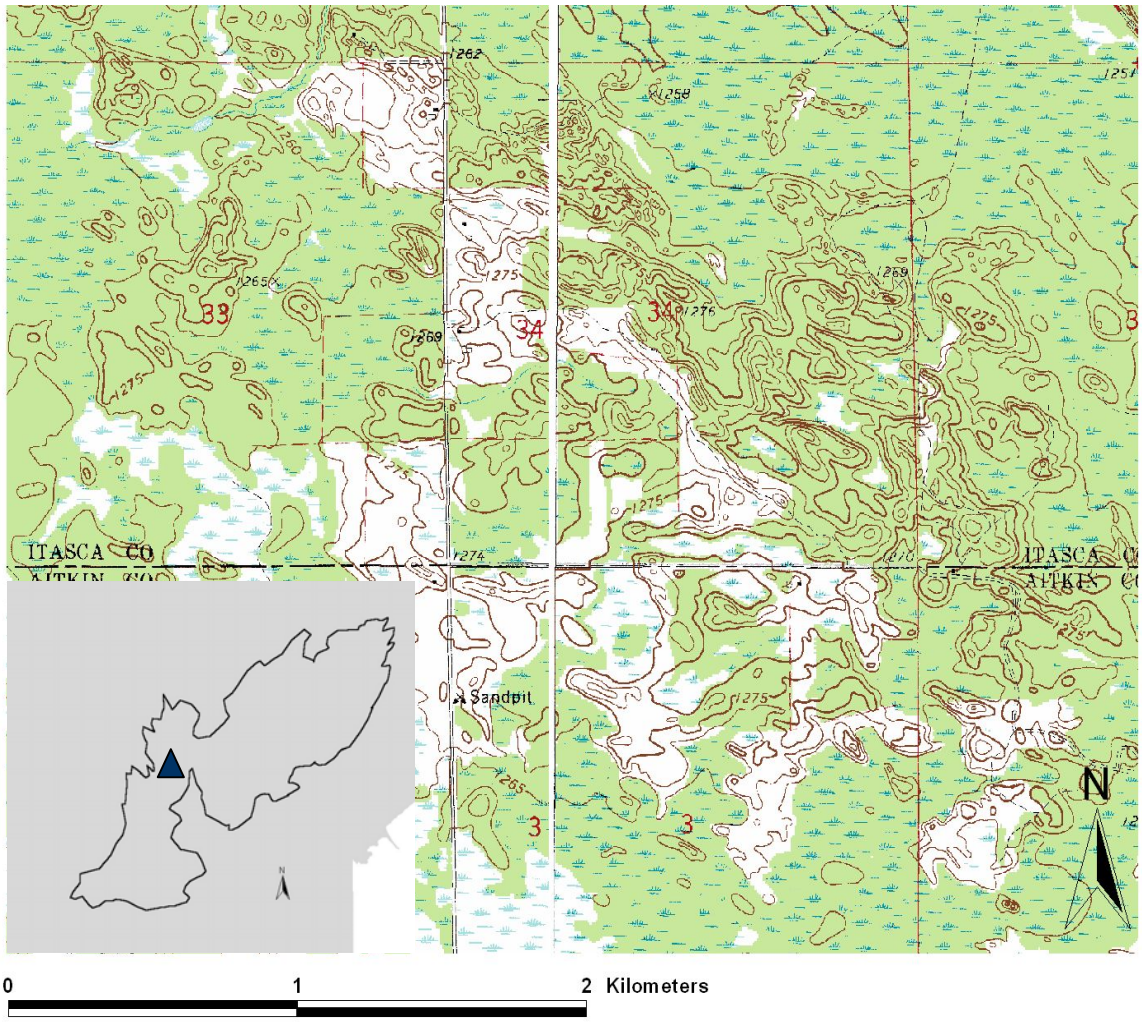


Figure 6-13. Distorted dunes in the northern end of Lake Aitkin basin showing a combination of crescentic and longitudinal dunes (Jacobson and Split Hand 7.5 minute USGS quadrangles).

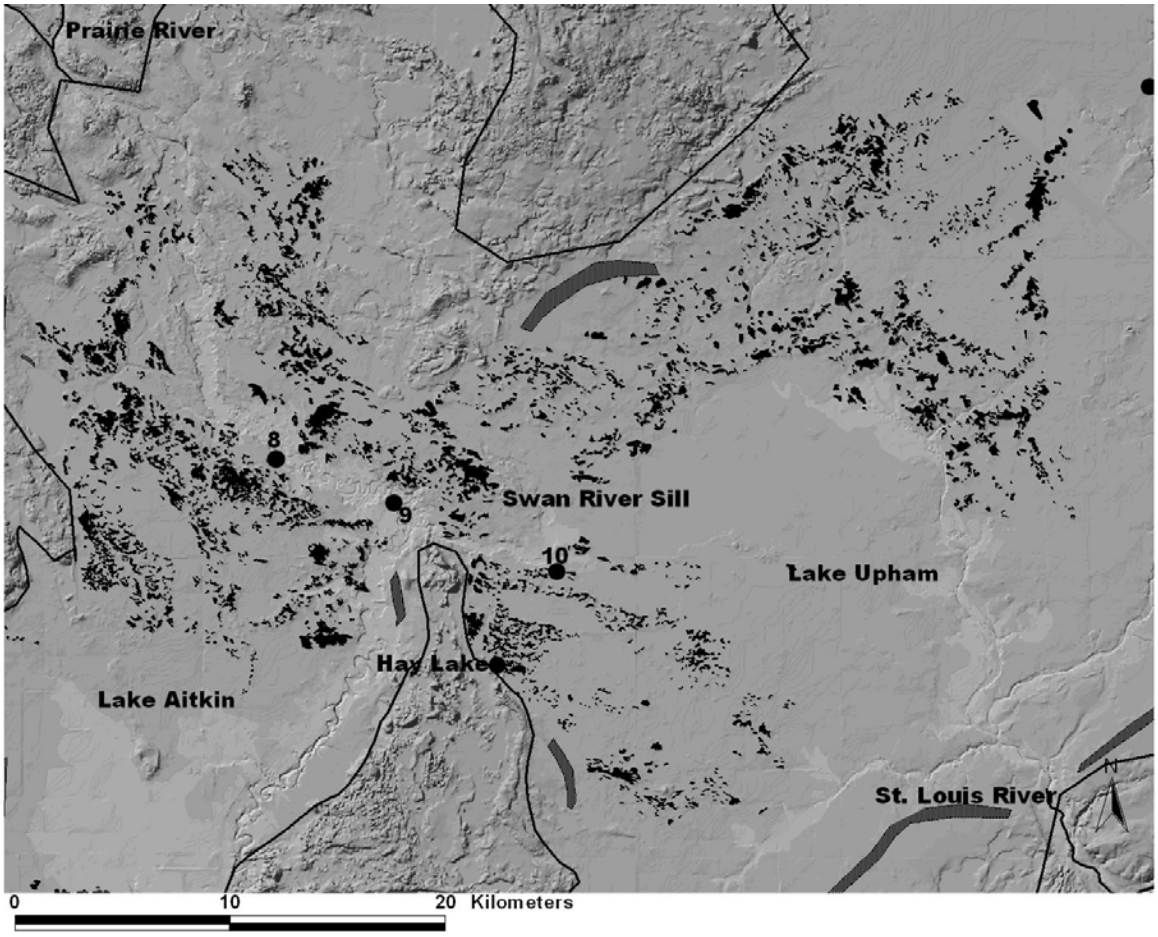


Figure 6-14. Trend in dune clusters indicating a NW-SE wind direction. Elongate polygons are beaches. Circles indicate stops 8, 9, and 10 sites.

STOP 6 – 6. <u>Glacial Lake Upham II Shoreline</u>. (SE¼ SE¼ Sec. 33, T56N, R21W, St. Louis County*, Silica 7.5' USGS quadrangle, 499595E, 5236726N, NAD83*)

After the time of upper beach formation, Lake Upham experienced a drop in water level documented at Townline beach (411 m; 1350') through the use of Ground Penetrating Radar (Fig. 6-15). The GPR results are interpreted as a down stepping of shoreline deposits (Fig. 6-16). The progradation of the bedform resulted in a constructional shoreline and indicates regression of the lake. The regression is an example of Forced Regression as discussed by Posamentier and Allen (1999). Forced regression takes place when there is a relative sea-level fall that progressively exposes the sea (or lake) floor, thereby causing the shoreline to migrate seaward (Posamentier and Allen 1999).

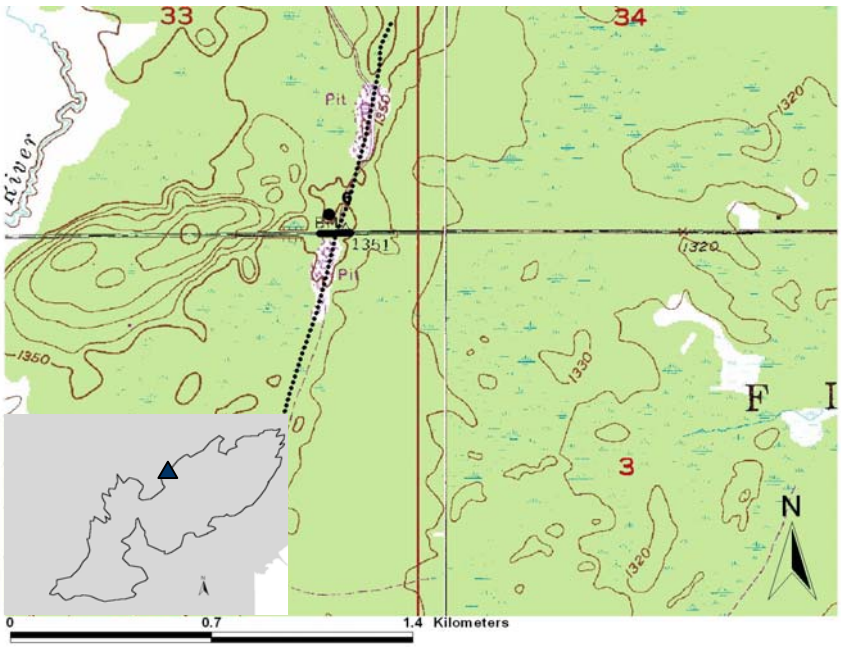


Figure 6-15. Shoreline of Glacial Lake Upham II along Townline Road south of Hibbing. Shoreline indicated by black dotted line. Heavy solid line indicates where GPR data was collected (Silica & Riley 7.5 ‘ USGS quadrangle).

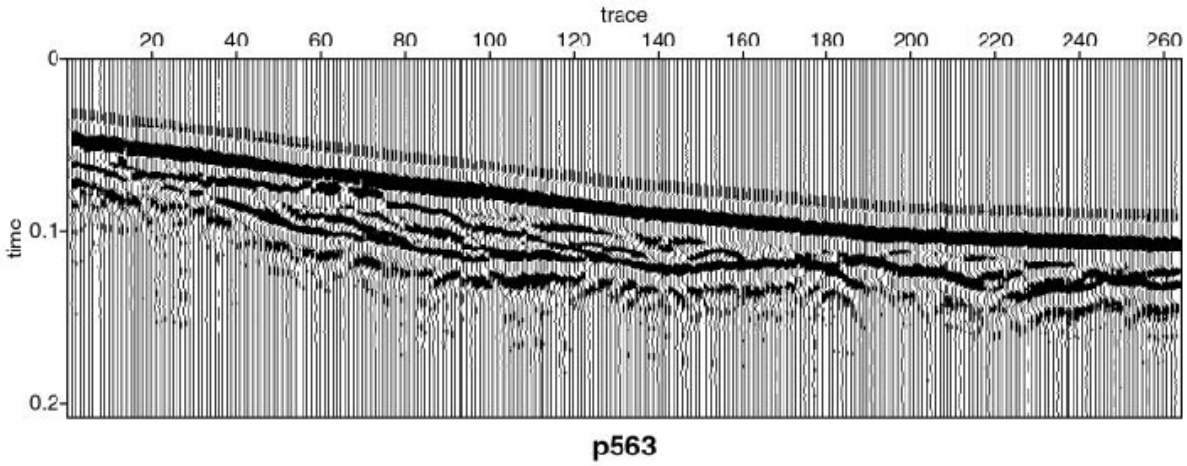


Figure 6-16. Ground Penetrating Radar profile of Glacial Lake Upham II shoreline. Vertical data represents ~9 meter; horizontal data ~121 meters. (Data processing courtesy of Nigel Wattrus at Large Lakes Observatory, University of Minnesota-Duluth)

STOP 6 – 7. <u>Goodland Esker</u>. (SE¼ NE¼ Sec. 9, T55N, R23W, Itasca County, *Calumet 7.5' USGS quadrangle, 479736E, 5234496N, NAD83*)

This stop is a gravel pit exposure in the Goodland esker, one of the most prominent glacial landforms in Itasca County. The location of the esker and details of its formation place important constraints on the timing of the St. Louis Sublobe advance and deposition of the Alborn drift (Fig. 6-17).

The morphology of the esker system is best described by the model of Shreve (1985). The proximal (with respect to the Giants Range) segment is an ~600 m wide, 10 km long channel incised into bedrock and older drift (tunnel valley), occupied by a multi-crested esker. The medial segment consists of a single broad-crested esker segment deposited in a 10 km long channel incised into the overlying ice. The distal component of the esker system is a supraglacial fan complex.

The base of the subglacial channel drops from 428 m at the crest of the Giant's Range to <400 m at a point 6 km downstream, then rising to 409 m over the next 4 km. Downstream of this point, the elevation of the top of the broad-crested esker rises 19 m over 10 km, from 434 to 455 m at the apex of the distal fan.

The transition from a multi-crested to broad-crested esker is apparently triggered by an increase in the adverse slope up which the subglacial channel flowed. Multi-crested eskers form in areas of moderate upgradient flow as the channel tends to migrate laterally, rather than upward, in the ice, while broad-crested eskers form in areas of steeper upgradient flow due to freezing of the tunnel walls, a process favoring low, wide tunnel geometries (Shreve 1985).

Lithologies in the multi-crested segment esker are dominantly Archean granitics and greenstones transported from north of the Giant's Range. The broad-crested segment has a markedly different lithologic assemblage than the multi-crested segment. In addition to Archean lithologies, an abundance of Paleoproterozoic Animikie Basin lithologies are present including quartzite, sulfidic mudstones, iron formation, and pisolitic lateritic iron formation. The abundance of locally-derived Animikean lithologies indicates material in the broad-crested segment was deposited as incision of the N-channel was taking place upstream. The broad-crested segment is therefore older than the multi-crested segment. The wide range of mean sizes of the various Animikean

lithologies attests to the strong control of primary rock characteristics on particle size. Note in particular the abundance of Pokegama quartzite as large (~1 m) boulders and its paucity in smaller size fractions.

Deposition of the distal fan occurred throughout the period the esker system was active. The initial phase was characterized by deposition of a supraglacial fan on stagnant ice. Fan deposits are identifiable up to 10 km from the fan apex, and cover in excess of 100 km^2. The maximum elevation on the fan complex is about 477 m, indicating at least 20 m of ice was present above the outlet of the englacial channel at the fan apex. During later stages of the esker system, underlying stagnant ice melted collapsing the earliest deposited fan sediment. The final phase of fan formation was characterized by incision of the collapsed fan head down to about 450 m.

The subglacial drainage system that deposited the esker was probably short-lived. It could not have formed before the St. Louis Sublobe occupied the area around Grand Rapids, blocking the natural southwesterly flow of meltwater from the watershed of the present-day Prairie River. Similarly, it could not have persisted after St. Louis Sublobe ice wasted away and the Prairie River began flowing into Glacial Lake Aitkin II. Consequently, it may have been active a few hundred years at most. Despite its short duration, the large size of the esker system – it is one of the largest in Minnesota – attests to an enormous discharge of meltwater responsible for its formation. Meltwater was gathered from an interlobate zone that existed to the north of the Giant's Range between the St. Louis Sublobe and the Rainy Lobe. This interlobate area may have drained well in excess of 1000 km^2 of the ice sheet.

The elevation difference between the fan apex (477 m) and the surrounding (subglacial) landscape (400 m) indicate that a continuous cover of Rainy Lobe ice at least 75 m thick was present south of the Giant's Range at the time of esker formation and advance of the St. Louis Sublobe. This argues strongly against the postulated ice-free zone between the northern margin of the St. Louis Sublobe and the southern margin of the Rainy Lobe. Previous workers have postulated a relatively late advance for the St. Louis Sublobe (ca. 11.7 kyr BP), correlating it with the Vermilion Phase (ca. 12.0 kyr BP), or an even later phase, of the Rainy Lobe (Wright 1972; Hobbs 1983). However, the presence of active Rainy Lobe ice just north of the Giant's Range at the time of the

advance of the St. Louis Sublobe and esker formation provides further evidence that the St. Louis Sublobe advanced at a significantly earlier date.

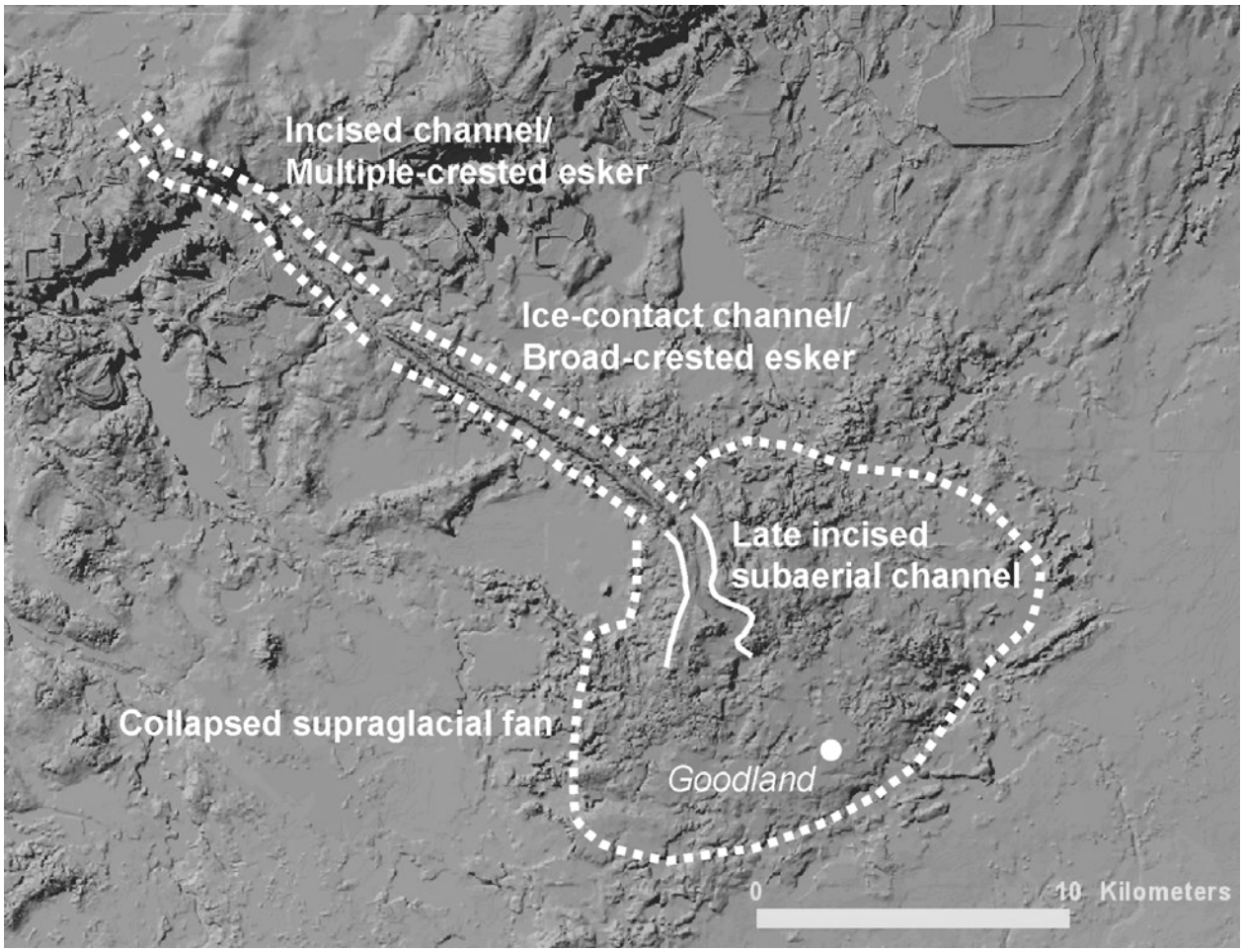


Figure 6-17. Main features of the Goodland esker system. (Image from 30 m DEM of MN)

STOP 6 – 8. <u>Prairie River underflow fan in Glacial Lake Aitkin II</u>. (NE¼ NE¼ Sec. 35, T53N, R24W, Itasca County, *Jacobson 7.5' USGS quadrangle, 473831E, 5209702N, NAD83*)

Glacial Lakes Aitkin and Upham II had two successive major meltwater inlets, the Embarrass Gap and the Prairie River. Inflow of the Prairie River into Aitkin II resulted in deposition of a large underflow fan extending 50 km from its apex (Hobbs 1983). The fan was later incised by the Mississippi River.

The cutbank at this stop exposes ~9 m of sandy underflow fan sediments overlying a lacustrine clay (Fig. 6-18). Visible within the underflow sediments are several Bouma sequences.

Outflow from Glacial Lake Koochiching into Glacial Lake Upham II through the Embarrass Gap ceased when a lower outlet opened north of Grand Rapids because of collapse of stagnant ice-cored terrain. Meltwater then flowed down the Prairie River into Glacial Lake Aitkin II, lowering the level of Koochiching from 427 to 411 m (1400' to 1350') (Hobbs 1983) The Prairie River entered Glacial Lake Aitkin II at an elevation of 396 m (1300') (Fig. 6-1). Although it is not known exactly when the Embarrass Gap was abandoned, it must have occurred prior to a 10.2 kyr B.P. transition from predominantly clastic to organic lake sedimentation in Sabin Lake located in the Embarrass Gap (Björk 1988; Lehr and Hobbs 1992). Fenton (1983) suggests that the Prairie River outlet was initiated between 12.3 and 10.8 kyr B.P. based on the chronology of the Lake Agassiz basin to the west. Clayton (1983) suggests 11.5 kyr B.P. for the inception of the Prairie River outlet.

The Prairie River inlet was abandoned once Glacial Lake Koochiching began to flow into Glacial Lake Climax. However, Glacial Lakes Aitkin and Upham II persisted after abandonment on the last meltwater inlet. Upham finally drained as the St. Louis River outlet was incised to its modern level. Hobbs (1983) indicates that water could have still been flowing into Aitkin after drainage of Upham by way of the Mississippi River, which has a prominent terrace at 389 m (1275').

Farnham and others (1964) obtained a date of 11,635 yr BP on a paleosol in the southwestern part of the Glacial Lake Aitkin II basin. This paleosol is overlain by a thin

marl and 3 feet of clay, indicating Aitkin was in existence well after this time. A radiocarbon date of 10,000 yr BP was obtained from a marl from the Ball Bluff quadrangle in the northeastern part of the Aitkin basin (Hobbs 1983) suggests the lake persisted at least until this time.

There are two alternative explanations explaining how the southwestern portion of the Glacial Lake Aitkin II basin was successively drained and re-inundated.

The close similarity between the presumed initiation of the Prairie River inlet at 11.5 kyr BP (Clayton 1983) and the re-inundation of the southwestern part of the Glacial Lake Aitkin II basin at 11,635 yr BP suggests a causal relationship. Glacial Lake Aitkin II may have assumed a lower, stable level by ca. 11.6 kyr BP. The relatively rapid lowering of Glacial Lake Koochiching would have sent a surge of meltwater into the lake, resulting in a rise in lake level even if only temporarily. The lacustrine sediment overlying the Farnham (1964) paleosol was thus derived from the Prairie River inlet.

The alternate explanation is that much of the Aitkin basin was drained through Glacial Lake Upham II by downcutting of the St. Louis River outlet. The Swan River Sill controlled the water level in the remaining lake. As isostatic rebound of the basin progressed, the sill on the northeastern side of the basin rose at a higher rate than the southwestern portion of the basin. Lake level transgression resulted in re-inundation of lake bottom exposed by drainage of Upham.



Figure 6-18. Stop 6-8. Exposure of the Prairie River underflow fan at a Mississippi River cutbank.

STOP 6 – 9. <u>Rhythmite in Glacial Lake Aitkin II</u>. (SW¼ NW¼ Sec. 4, T52N, R23W, Aitkin County, *Jacobson 7.5' USGS quadrangle, 479288E, 5207689N, NAD83*)

A rhythmite is overlain by cross-bedded sands, which is then overlain by underflow sediments. This is an interesting site that may document a dramatic inception of the Prairie River inlet into a basin that was previously depositing clays.

STOP 6 – 10. <u>Eolian dunes in Glacial Lake Upham II basin</u>. (NW¼ NE¼ Sec. 18, T 52 N R 22 W, Aitkin County, *Vanduse Lake 7.5' USGS quadrangle, 486857E, 5204509N, NAD83*)

Exposed at this site are eolian sand dunes overlying lacustrine clay of Glacial Lake Upham II (Fig 6-19). Dunes in the Glacial Lakes Aitkin and Upham basins are most densely concentrated near the inlet of the Prairie River to the Aitkin basin and east of the Swan River Sill in the Upham basin (Fig. 6-14). The 4-φ mean grain size of the dunes sampled in these areas in the basin match that of sediments from the underflow fan sediments seen at Stop 8. These sediments have been mapped in the Aitkin and Itasca Soil Surveys (Nyberg 1987, 1999) as the Zimmerman, Cowhorn, and Wawina soil series, all composed of fine to very fine grained sands of lacustrine origin.

From the inception of Glacial Lakes Aitkin and Upham II net flow of water was from Aitkin to Upham. The initiation of the Prairie River inlet greatly increased this discharge. The flow of water from Aitkin to Upham resulted in formation the Ball Bluff Spit (Marlow, 2004) extending eastward from the Swan River Sill (Fig. 6-19). The dense dune clusters east of the Swan River Sill in the Upham basin formed from sediment transported from the underflow fan.

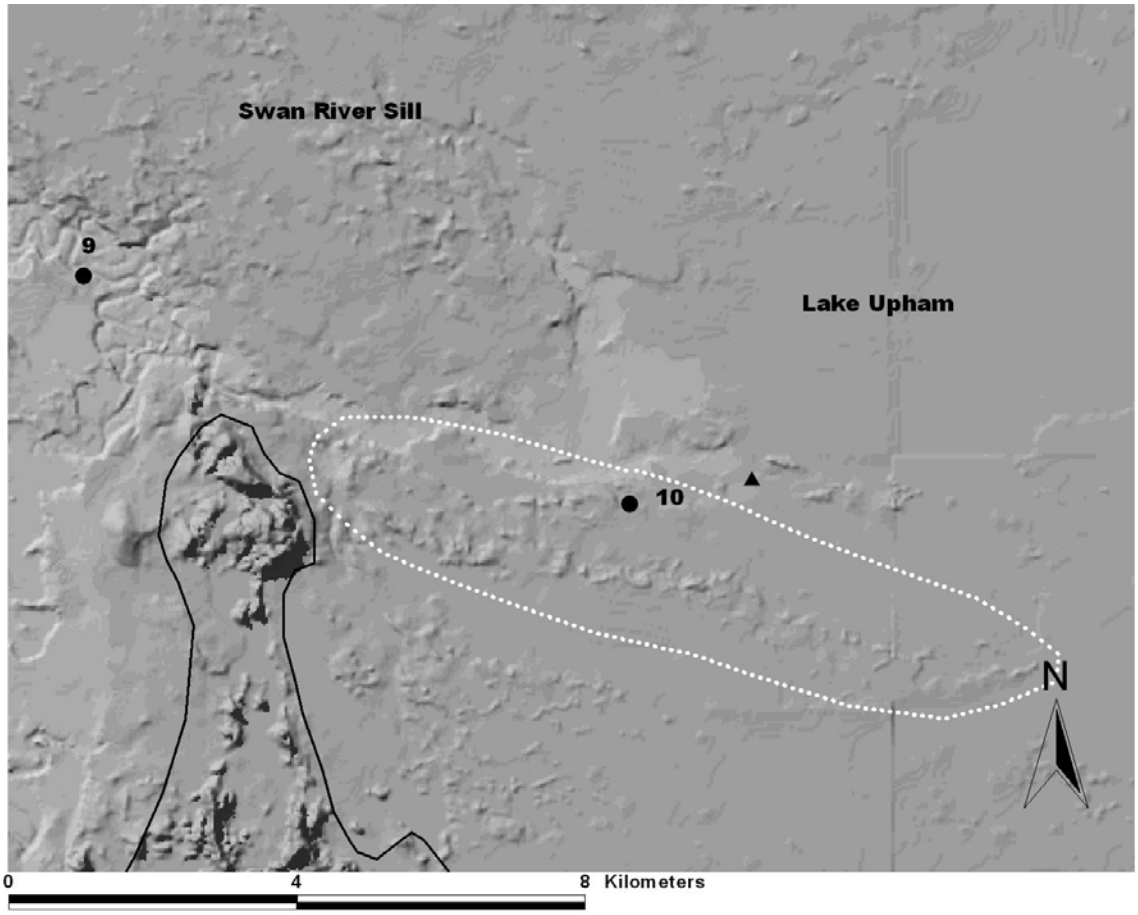


Figure 6-19. Location of the Ball Bluff spit located within the dashed line. Circle at stop 10 indicates where lacustrine clay is overlain by 1 m. of sand derived from the underflow fan. Triangle indicates site where lake clay is exposed at the surface.

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Road Log

MILES Begin in Duluth, MN.
0-3 Interstate 35 south to Grand Avenue/MN 23 exit.
3-22 14 miles to County 4. East (left) 2 miles to railroad crossing.
Stop 1 Glacial Lake Duluth underflow fan.

22-39 County 4 west ~12 miles to Interstate 35.
39-49 North on Interstate 35 ~12miles to MN 33.
49-69 North on MN 33 through Cloquet for ~18 miles to US 53.
69-73 North on US 53 for ~4 miles.
Cloquet River
Hellwig Creek
Stop 2 Hellwig Creek outlet and Glacial Lake Upham II basin overlook.

73-84 North on US 53 for ~12 miles to County 52 (Arkola Road) at Cotton, MN.
84-99 West on County 52 for ~16 miles to County 5.
99-101 North on County 5 for ~2 miles to gravel pit on left.

Stop 3 Wave-washed esker island (Toivola, MN).

101-115 North on County 5 from previous stop for ~13 miles to MN 37.
115-119 East on MN 37 for ~3.5 miles to County 25.
119-120 North on County 25 for ~1.5 miles to the gravel pit on the left.

Stop 4 Rainy Lobe collapsed outwash, Glacial Lake Upham I sediments, and St. Louis Sublobe till (Cherry, MN).

120-121 South on County 25 to MN 37.
121-135 West on MN 37 for ~9.5 miles to County 57 (1st Avenue).

135-140 South on County 57 for ~8 miles. Stop just before County 442 (Townline Road South).

Stop 5 Eolian dunes.

140-140 South on County 57 to County 442 (Townline Road South).
140-144 West on County 442 for ~3 miles. Stop at gravel pit access road on north side.

Stop 6 Glacial Lake Upham II Shoreline.

144-145 West on County 442 (Townline Road South) to MN 73.
145-145 North on MN 73 for ~0.5 miles to County 16.
145-156 West on County 16 for ~11 miles to County 20.
156-164 West on County 20 for 0.25 miles to MN 65.
164-168 North on MN 65 for ~8.5 miles to County 12.
168-169 West on County 12 for ~3.5 miles to County 70.
169-170 South on County 70 for ~1 mile to County 434.
170-170 South on County 434 for ~1 mile to gravel pit on right

Stop 7 Goodland Esker.

170-175 Continue south on County 434 for ~5 miles to County 10.
175-176 West on County 10 for ~1.5 miles to County 71.
176-181 South on County 71 for ~4 miles to US 2.
181-181 Cross US 2 and continue for ~ 0.25 miles to County 441 (Bluebird Lane)
181-184 West on County 441 for ~2.5 miles to County 3 (River Road).
184-197 South on County 3 for ~2.5 miles to cutbank along Mississippi River on left.

Stop 8 Prairie River Underflow fan in Glacial Lake Aitkin II.

197-200 Continue southbound on County 3 (River Road) for ~2 miles to Mississippi River Boat Landing Road on the left

Stop 9 Rhythmite in Glacial Lake Aitkin II.

200-201 Return to County 3.
201-202 South on County 3 to MN 200.
202-203 East on MN 200 for 1.5 miles to MN 65.
203-203 North on MN 65 for 0.5 miles to MN 200.
203-208 East on MN 200 for ~4 miles to 154th Avenue.
208-208 North on 154th Avenue.

Stop 10 Eolian dunes in Glacial Lake Upham II basin.

To return to Duluth:
208-208 Return to MN 200.
208-213 East on MN 200 to US 2.
213-262 East on US 2 to Interstate 35.
North on Interstate 35 to start point.

FIELD TRIP 7

ECONOMIC GEOLOGY OF ARCHEAN GOLD OCCURRENCES IN THE VERMILION DISTRICT, NORTHEAST OF SOUDAN, MINNESOTA

By

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INTRODUCTION

A widespread area of gold mineralization occurs in numerous prospects east of Lake Vermilion, within the Vermilion greenstone belt of northeastern Minnesota. The mineralization occurs in rocks of the Neoarchean (~2.7 Ga) Bass Lake sequence (Peterson and Jirsa, 1999a) of the Wawa subprovince of the Canadian Shield. This zone of abundant gold mineralization is bounded to the south by the Mud Creek shear zone and to the north by the Vermilion fault (Fig. 7-1). The main access to these prospects is along the Mud Creek road (St. Louis County 38).

A brief period of mineral exploration for lode-gold deposits in this immediate area of the Vermilion district occurred in the mid 1980s to early 1990s. These programs typically consisted of grid-based geologic mapping, bedrock sampling, ground geophysics, and the completion of soil geochemical surveys. Conversations with many of the people involved in gold exploration programs in the immediate field trip area (centered on Section 6, Township 62 North, Range 14 West), and compilation of all exploration data from the district as a whole (data from the terminated lease files of the Minnesota DNR), has led to the conclusion that interpretation of linear structural elements exposed in outcrops were not used in designing exploratory drilling plans in the field trip area. Therefore, many of the prospects discovered as a result of these exploration programs remain untested by drilling.

The authors recently completed a project (Peterson and Patelke, 2004) consisting of detailed geologic and structural mapping within many of the known gold prospects in the field trip area, and will highlight some of the insights generated as a result of this work during the course of this one day field trip. This recent work consisted of detailed outcrop mapping (at scales ranging from 1:1,000 to 1:3,000) that was focused on structural (i.e. shear zones, lineations, intersecting foliations and small-scale folds), geological (i.e. contact relationships, competence contrasts), geochemical (i.e. gold assays, trace element characteristics), and alteration (Fe-bearing carbonate, sericite, pyrite, silicificaton) features within and around areas of gold mineralized exposures. Based on detailed mapping of features described above, the project's goal was to try to determine the location and orientation at depth of gold-rich mineralized zones exposed in outcrops. The mapped geological features were incorporated into the data compiled and described by Peterson (2001), and are available online as a 1:12,000 scale geological map, as well as ArcView GIS data files at http://www.nrri.umn.edu/egg/.

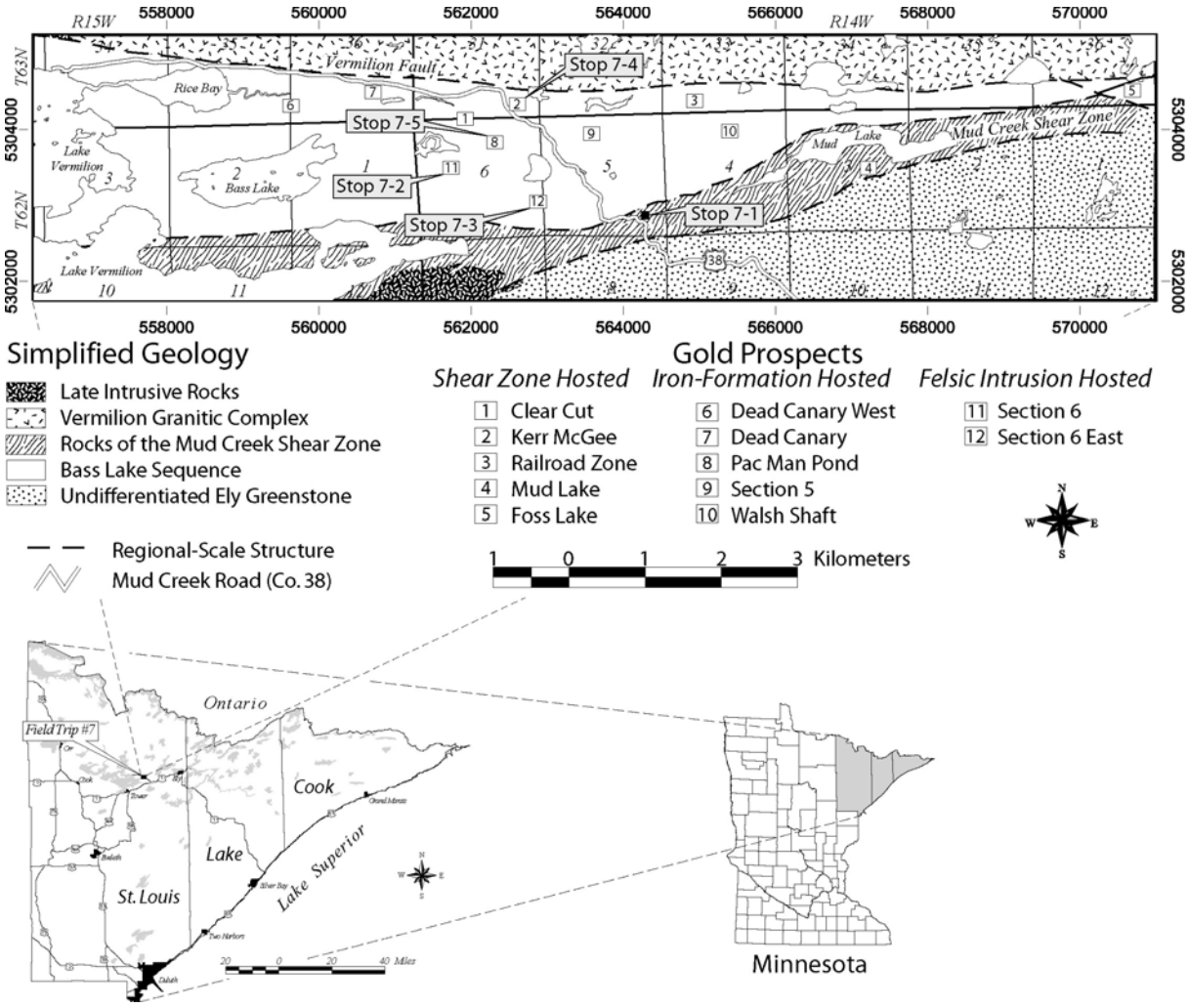


Figure 7-1. Location, simplified geology, lode gold prospects, and field trip #7 stops.

LODE GOLD ORE DEPOSIT MODEL

The brief description of Archean lode-gold deposits that follows is presented as both a basic reference and also to highlight the important features of the model that will be seen during the field trip. Archean lode-gold deposits are one category of ore deposit classified as mesothermal lode-gold deposits (Hodgson, 1993). This deposit type has also been called orogenic gold (Groves et al., 2000), greenstone gold (Robert et al., 1991), Archean lode-gold, mesothermal gold-quartz veins, shear-hosted gold, low-sulfide gold-quartz veins (Berger, 1986b), lode-gold, Mother Lode veins (Bohlke and Kistler, 1986), and iron formation-hosted gold deposits (Berger, 1986a; McMillan, 1996; Rye and Rye, 1974; Fripp, 1976; Kerswill, 1993; Thorpe and Franklin, 1984; and Vielreicher et al., 1994).

Whatever the name, they are a widespread group of epigenetic ore deposits that have formed in similar settings throughout geologic time. In general, the deposits form during

compressional or transpressional deformation at convergent plate margins in accretionary or collisional orogens (Fig.7-2). They form over a large crustal-depth range (2 to 20 km) from deep-seated, low-salinity H_2O-$CO_2 \pm CH_4 \pm N_2$ ore fluids, with Au transported as reduced sulfur complexes. The ore fluids are believed to be generated during lower crustal metamorphism from dehydration reactions. Regional structures provide the main control on distribution of lode gold deposits and mining camps. In many terranes, first-order faults or shear zones appear to have controlled regional fluid flow, with greatest ore-fluid fluxes in, and adjacent to, subsidiary faults, shear zones and/or large folds. Highly competent and/or chemically reactive rocks are the most common hosts to the larger deposits. Gold deposition occurs late during the evolutionary history of the host terranes, normally within D_3 or D_4 in a D_1-D_4 deformation sequence. Absolute ages of mineralization support their late-kinematic timing, and, in general, suggest that deposits formed diachronously towards the end of the evolutionary history of hosting orogens.

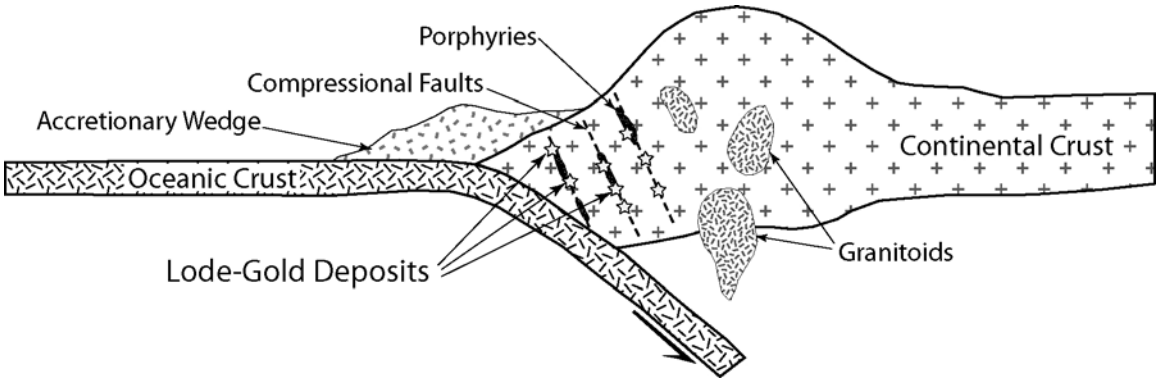


Figure 7-2. Generalized tectonic model for the formation of mesothermal gold deposits, after Groves et al., 2000.

The late timing of lode-gold deposits is critical to geology-based exploration methods, and hence mineral potential evaluations for these deposits. The late timing is critical because of the present structural geometry of: (1) the deposits, (2) the mining camps, and (3) the enclosing geologic terranes are essentially all similar to the structural geometry during gold mineralization. Therefore the interpretation of bedrock geological maps and cross-sections can be used to discern the physical conditions that existed at the time of ore deposition. Exploration for mesothermal lode-gold deposits should incorporate various aspects of the ore deposit model into criteria that can vector into the most favorable areas for hosting such mineralization. The most fundamental characteristic of this class of deposit is the spatial association of the deposits to regional structures in metamorphic terranes. Zones of widespread carbonate alteration (adjacent to regional structures) should be identified and used to focus subsequent exploration. Within carbonate alteration zones, gold is typically only in areas containing quartz veins, silicification, and/or sericite alteration (with or without sulfides). Two general structural controls on the orientation of lode gold ore shoots include deflections and curvatures of shear zones, and where high strain zones intersect favorable geological elements (Poulsen and Robert, 1989). A generalized sequence of events associated with the formation of typical Archean mesothermal gold deposits is outlined below. The possible observation of many of these important features during the course of the field trip is highlighted by the bracketed, bold text.

I. Deposition of Archean Supracrustal Strata (Volcanic and Sedimentary Rocks)
II. Compression and Early (D_1) Folding **[stop 7-1]**
III. Renewed Compression and Transpression (D_2)
  1) Development of Major Crustal Shear Zones **[stop 7-1]**
  2) Strain Partitioning and Development of Subsidiary Shear Zones **[stop 7-3, 7-4]**
  3) Intrusion of Porphyritic Intrusions, Continued D_2 Deformation **[stop 7-2, 7-3 and 7-5]**
      -Local Development of Competency Contrasts **[stop 7-3, 7-5]**
  4) Lower Crustal Dehydration and Fluxing of CO_2-rich Fluids up Structures
      - Fe-Carbonate Alteration Zones **[stop 7-3, 7-4, 7-5]**
  5) Fluxing of Gold-Bearing Hydrothermal Fluids up Subsidiary Shear Zones, Latest D_2
      - Gold Mineralization Associated with:
        A) Quartz-Pyrite Veins **[stop 7-2, 7-3, 7-4, and 7-5]**
        B) Silicification with Pyrite **[stop 7-4]**
        C) Pyritization of Fe-Rich Host Rocks **[stop 7-5]**
        D) Quartz-Sericite-Pyrite Schists/Phyllites **[stop 7-4]**
        E) Contact Zones Between Rock Units **[stop 7-5]**
IV. Erosion

REGIONAL GEOLOGIC SETTING OF THE VERMILION DISTRICT

The field trip area is located in the Neo-Archean (~2.7 Ga) Vermilion Greenstone Belt of the classic Vermilion district of northeastern Minnesota. Supracrustal rocks in the Vermilion district consist of volcanic-dominated stratigraphic sequences of the Wawa subprovince of the Superior Province of the Canadian Shield. Rocks of the Wawa subprovince in northern Minnesota are divided on the basis of stratigraphic and structural setting into: (1) the southern Soudan belt and (2) the northern Newton belt (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 designated informally as the Leech Lake structural discontinuity (Jirsa et al., 1992). In the Vermilion district, 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).

The Soudan belt 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, east- to northeast-trending, and mostly northward-younging volcanic and volcaniclastic 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 relationship between stratigraphic units within each belt is 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. A simplified regional geological map of the Neo-Archean terranes of northeastern Minnesota and adjacent Ontario is presented in Figure 7-3.

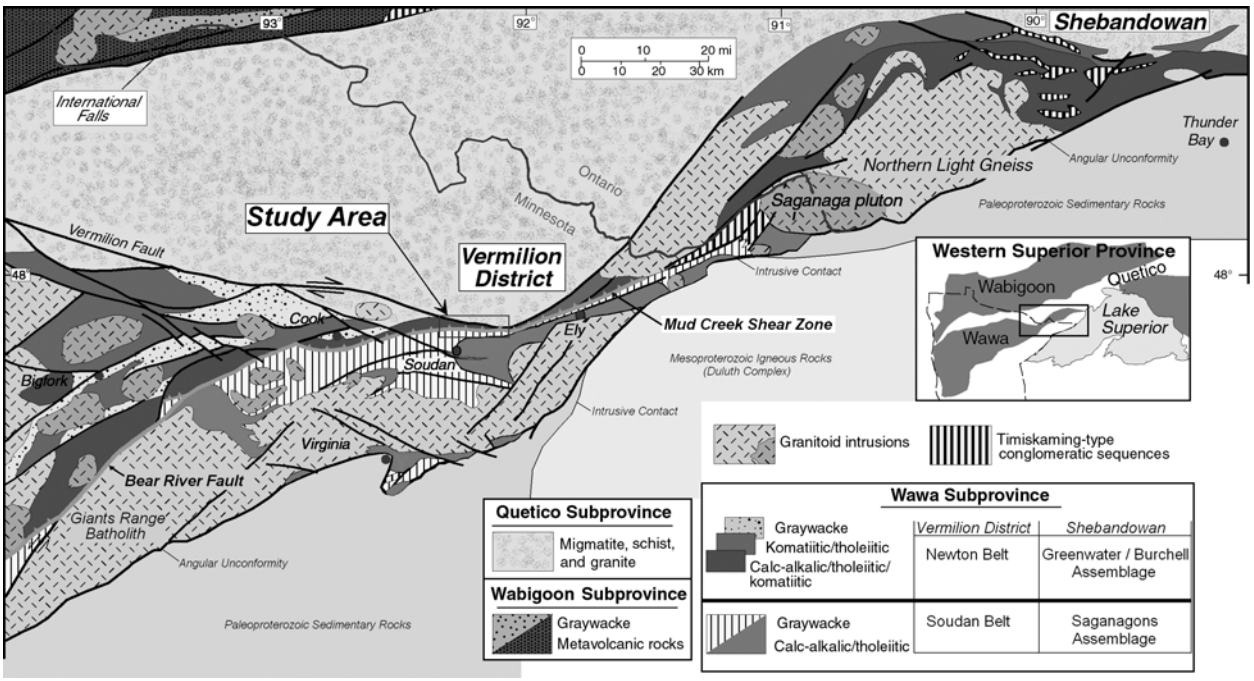


Figure 7-3. Simplified correlation map of Neo-Archean assemblages across the U.S. - Canada border, modified from Peterson et al. (2001). Inset shows major subprovinces of the southwestern Superior Province.

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 (D_1) produced broad, locally recumbent folds within the Soudan belt and major fault zones throughout the region. In the Newton belt, D_1 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 sequences in local fault-bounded basins occurred late in D_1 deformation (Jirsa, 2000). A large, map-scale structure related to D_1 deformation in the Soudan Mine area 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 (S_1) locally.

A second deformation event (D_2) associated with synchronous regional metamorphism resulted in foliation development and structures having largely dextral asymmetry. D_2 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 D_2 deformation affected all of the supracrustal rocks in the area and is reasonably constrained by geochronology, the regional foliation (S_2) can be used in the field to

relate other structural, intrusive, and deformation events. The relationship between S_2 fabric and shear structures indicates that most shearing occurred relatively late in the D_2 event. Major shearing that produced the Mud Creek and related shear zones is attributed to the late stages of D_2 dextral transpression.

Regional D_2 strain patterns in the Vermilion district vary from north to south in the belt (Hudleston et al., 1988; Schultz-Ela and Hudleston, 1991). These patterns include flattening strains that occur to the north, near the present Vermilion Fault and constrictional strains to the south. The field trip area lies within the zone of flattening strain of Hudleston et al. (1988) and Schultz-Ela and Hudleston (1991). Schultz-Ela and Hudleston (1991) mathematically modeled the observed strain patterns as deformation paths that produced flattening strains (west plunging ë 1 axes) by dextral shear of the pre-existing constrictional strains (east plunging ë 1 axes). The linear E-W strain patterns and minor rotations about horizontal axes during the deformation, as described by Hudleston et al. (1988) and Schultz-Ela and Hudleston (1991), preclude early theories (Hooper and Ojakangas, 1971; and Hudleston, 1976) on the structural setting of the greenstone belt by infolding and shear off the flanks of rising granitic diapirs. An estimate of 50% north-south shortening across the belt is proposed by Schultz-Ela and Hudleston (1991), and their model favors an origin of the Vermilion district rocks at a convergent margin, most likely as a N-dipping subduction complex with shallow slab dip. The origin of the southern constrictional strains remains enigmatic.

Structures related to the third deformation event (D_3) include abundant NE- and NW-trending faults that dissect the stratigraphic assemblages. Named structures related to D_3 include the NE-trending Waasa and Camp Rivard faults SSE of Ely, and the WNW-trending, crustal-scale Vermilion and related faults that form the Wawa-Quetico Subprovince boundary immediately north of the field trip area.

LOCAL GEOLOGIC SETTING; THE BASS LAKE SEQUENCE

The informally named Bass Lake sequence (Peterson and Jirsa, 1999a) of the Newton Belt occurs in an east-west trending, fault-bounded panel that widens to the west. The sequence is bounded by segments of the Mud Creek and Shagawa shear zones, and the Bear River, Haley, Vermilion, Burntside Lake, and Wolf Lake faults (Fig. 7-4). The continuation of stratigraphic and geophysical trends from areas of excellent exposure (east of Lake Vermilion) into poorly exposed areas (to the west) have led to the inclusion of rocks of the informally named Cook sequence (Southwick, 1993; Southwick et al., 1998) in the Bass Lake sequence. The western portion of the sequence consists of linear belts of graywacke and basalt cut by late felsic to intermediate composition intrusions.

East of Lake Vermilion, the geology of the Bass Lake sequence is dominated by six basic rock types, which include: (1) Tholeiitic pillowed basalt flows (see Fig. 7-4) interpreted to have formed in a deep-water setting based on volcanic textures; (2) Gabbro sills interpreted as synvolcanic in age due to their stratigraphic continuity and similar deformation as the enclosing pillowed basalts; (3) Felsic porphyries (feldspar porphyry and quartz-feldspar porphyry) interpreted to have intruded during late stages of D_2 deformation based on field relationships and geochronology (quartz-feldspar porphyry from the Pac Man Pond prospect returned a $^{207}Pb/^{206}Pb$ age of 2683.0 +/- 1.4 Ma (Peterson et al., 2001)); (4) Algoma-type iron-formation; (5) Thinly-

bedded argillite and siltstone; and (6) Sheared rocks, which are dominated by chlorite-rich schist, phyllite, and phyllonite. In addition, localized areas of fragmental felsic volcanic rocks occur stratigraphically below distinct iron-formation horizons. Regional and detailed geologic maps of the sequence are presented in Fig. 7-5.

In the last twenty years, numerous gold prospects have been discovered in the eastern portion of the sequence (Fig. 7-1). These prospects generally fall into one of three categories; (1) auriferous quartz-carbonate-pyrite veins and sulfidized zones in iron-formation; (2) auriferous quartz-sericite-ankerite-pyrite schists; and (3) felsic intrusive-hosted auriferous quartz veins and stockworks. All of the prospects are found within areas of moderate to strong iron-carbonate alteration, with the best mineralization commonly found within sericitic alteration zones. Numerous equigranular and porphyritic felsic intrusions occur within the areas of alteration and gold mineralization, and are a good guide for locating mineralized structures. The gold mineralization is generally related to deformation in subsidiary structures associated with movement along the D_2 Mud Creek shear zone. Geological and geochemical descriptions of many of the gold prospects have previously been given by Peterson and Jirsa (1999b).

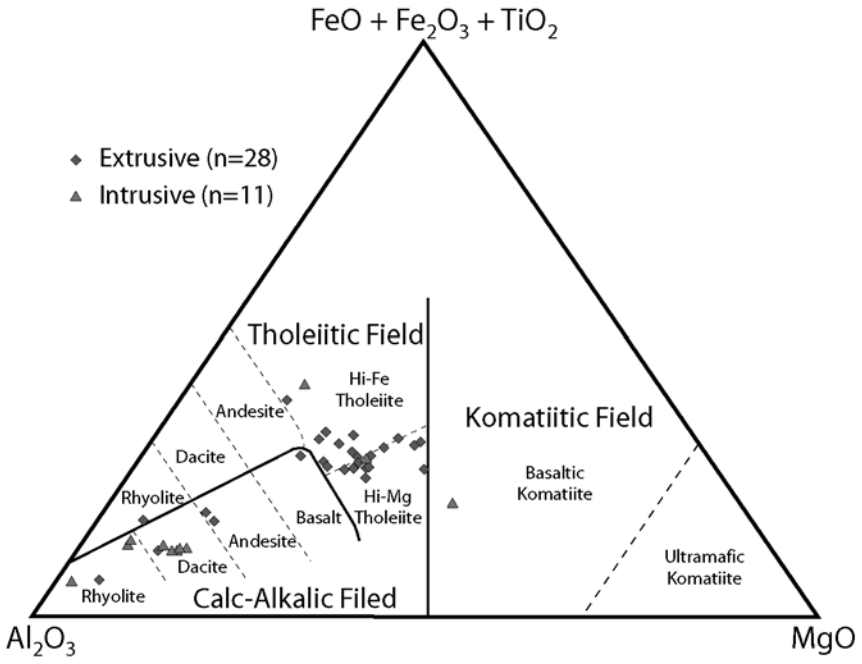


Figure 7-4. Jensen cation plot (Jensen, 1976) for volcanic and hypabyssal rocks of the Bass Lake sequence.

Undoubtedly the most striking structural feature in the immediate study area is the juxtaposition of the Soudan and Newton belts along the Mud Creek shear zone. In addition, numerous east-northeast trending highly strained zones occur to the north of the Mud Creek shear zone and these subsidiary sheared zones host a majority of the gold prospects in this area of the Bass Lake sequence. A stereonet projection of planar and linear structural features within the field trip #7 area of the Bass Lake sequence is shown in Fig. 7-6.

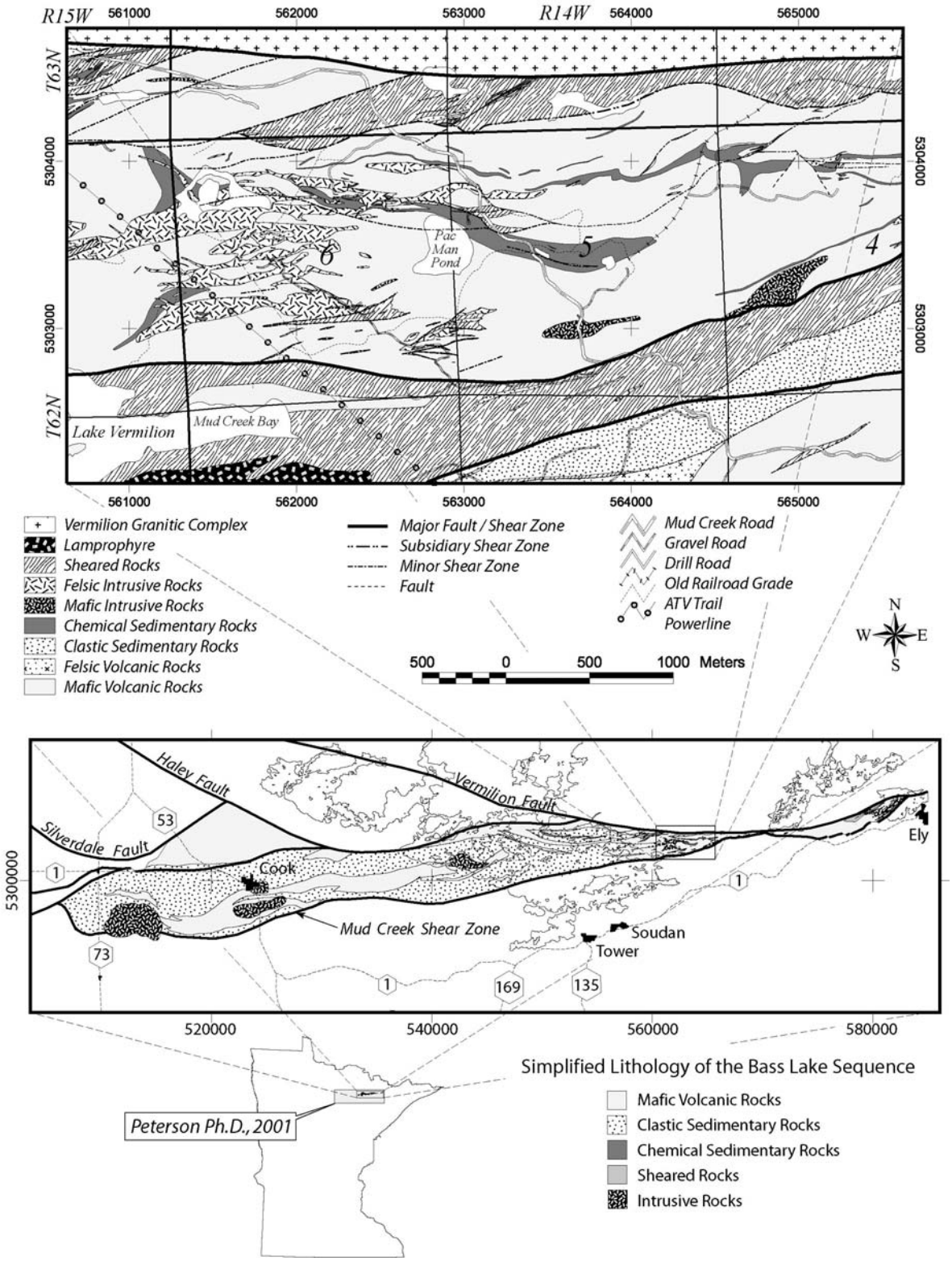


Figure 7-5. Regional to detailed geological maps of the Bass Lake sequence. Geology simplified from Peterson (2001) and Peterson and Patelke (2004).

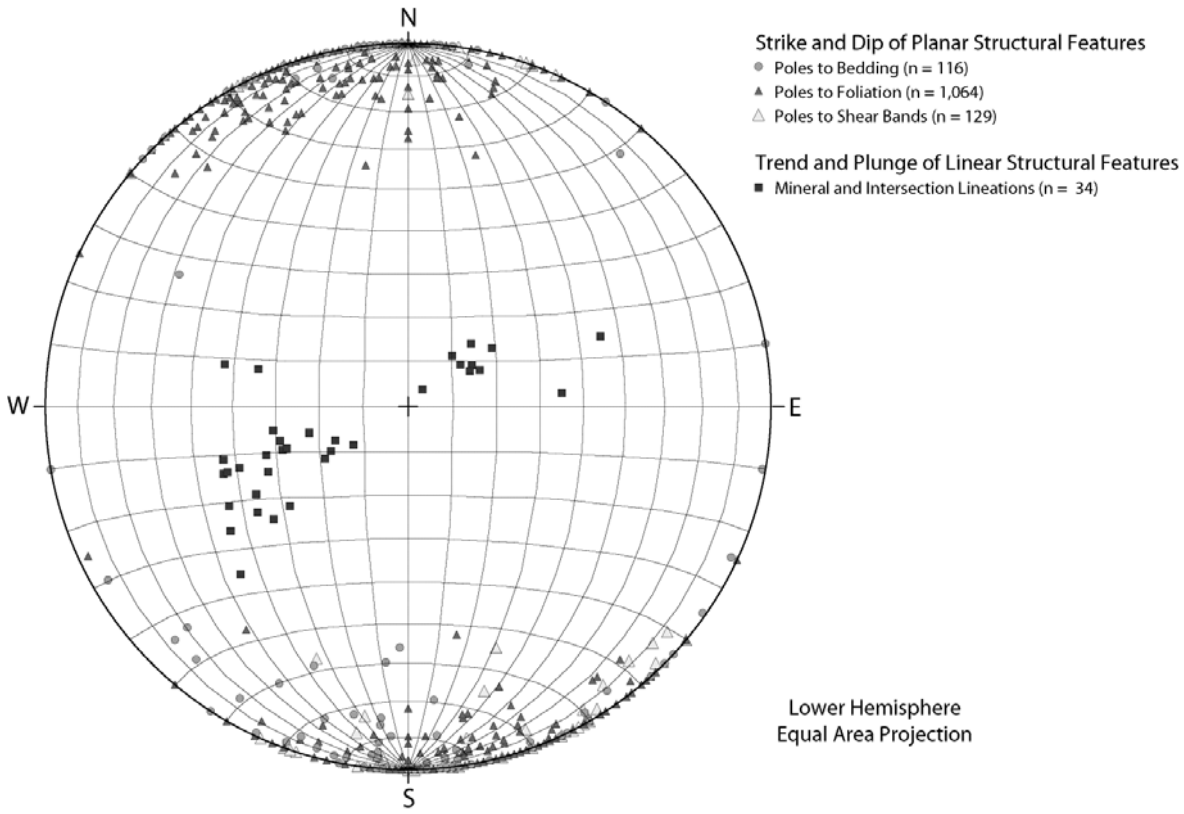


Figure 7-6. Stereonet projections of planar and linear structural features from the field trip area.

GLACIAL HISTORY

Late Pleistocene glacial deposits (Late-Wisconsin glaciation) in the Vermilion district are associated with the stepwise retreat of the Rainy Lobe of the Laurentide Ice Sheet, approximately 14,000 to 12,000 years ago. The repeated glaciations of the Pleistocene epoch modified the pre-existing topography of northeastern Minnesota, i.e. the surface was scoured by glacial ice, exposing fresh bedrock, and new surficial materials were deposited following the retreat of the glaciers. During the retreat of the glacier, the margin of the ice-sheet blocked the natural drainage to the north, and pro-glacial lakes formed in front of this barrier (glacial lakes Norwood, Koochiching, and Agassiz). Subglacial streams left sinuous ridges of sorted sand and gravel (eskers), and delta/fan complexes formed where these streams exited the ice margin and entered the pro-glacial lakes. The field trip area lies within the scoured bedrock terrain of northeastern Minnesota, immediately north of the Vermilion Moraine. Regional and detailed digital elevation maps of the area are presented in Figure 7-7, and display the pronounced recent bedrock scouring of the area, especially along the Vermilion fault and Mud Creek shear zone. Late Pleistocene surficial deposits and landforms that unconformably overlie the Neoarchean bedrock in the field trip area include a thin veneer of basal till, local outwash deposits, and sinuous eskers.

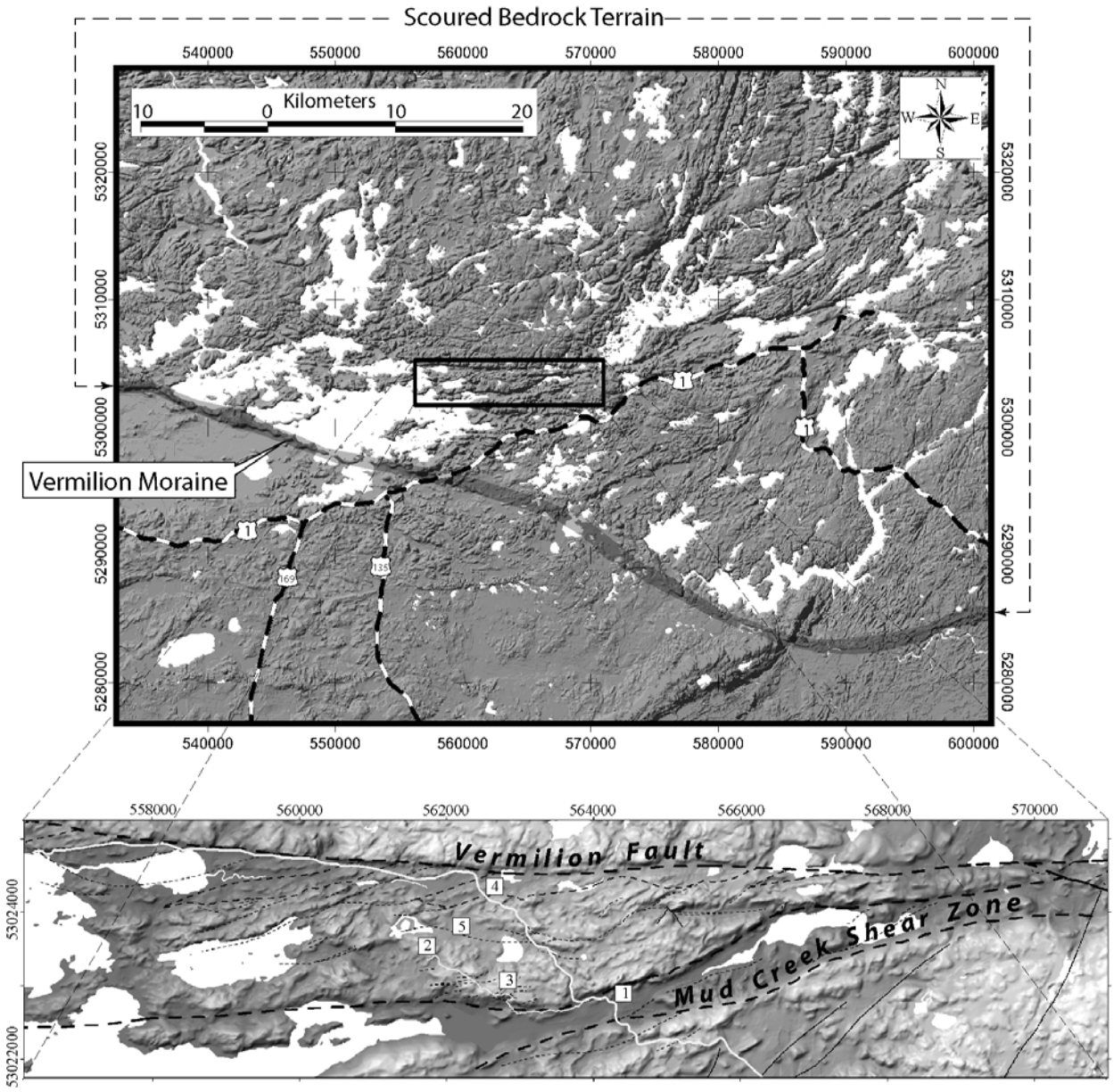


Figure 7.7. Regional (top) and local (bottom) digital elevation maps depicting superimposed glacial landforms. Grid values in the detailed map are UTM coordinates, in meters. The locations of field trip stops are depicted in the lower image.

FIELD TRIP STOPS

Stops for this field trip will include short traverses that visit a number of outcrops for each locality. The traverses are designed to highlight specific features of the lode-gold ore deposit model for each area, though time constraints preclude detailed analysis of each area.

STOP 7-1: **MUD CREEK SHEAR ZONE**

Location: (SE, SE, Sec. 5, T.62N., R.14W., NAD83 UTM 564200E, 5302800N)

Description: The regional scale Mud Creek shear zone occupies the east-northeast trending valley of Mud Creek, which is clearly visible at this location. This shear zone separates rocks of the Newton Belt (here the Bass Lake sequence) to the north and rocks of the Soudan Belt (Gafvert Lake sequence and the Upper Ely Greenstone Formation) to the south. The Mud Creek shear zone is analogous with major faults (Destor-Porcupine fault) and "breaks" (Cadillac-Larder Lake break) of major lode-gold mining districts in Canada. Historic gold assays taken from rocks of the shear zone itself are essentially devoid of gold, as is the case for most major structures within Archean lode-gold mining camps. The short field trip traverse is located within the northern margin of the internal highly strained zone of the shear, and will visit outcrops of: (1) ankerite-sericite-quartz-green mica-pyrite schist with quartz and tourmaline knots, and (2) highly folded and compositionally banded phyllites with quartz veins (Fig. 7-8). A simplified geology and field trip traverse map of stop 7-1 is presented in Figure 7-9.



Figure 7-8. Outcrop photographs of ankerite-sericite-quartz-green mica-pyrite schist (A and B), and highly folded and compositionally banded phyllites (C and D).

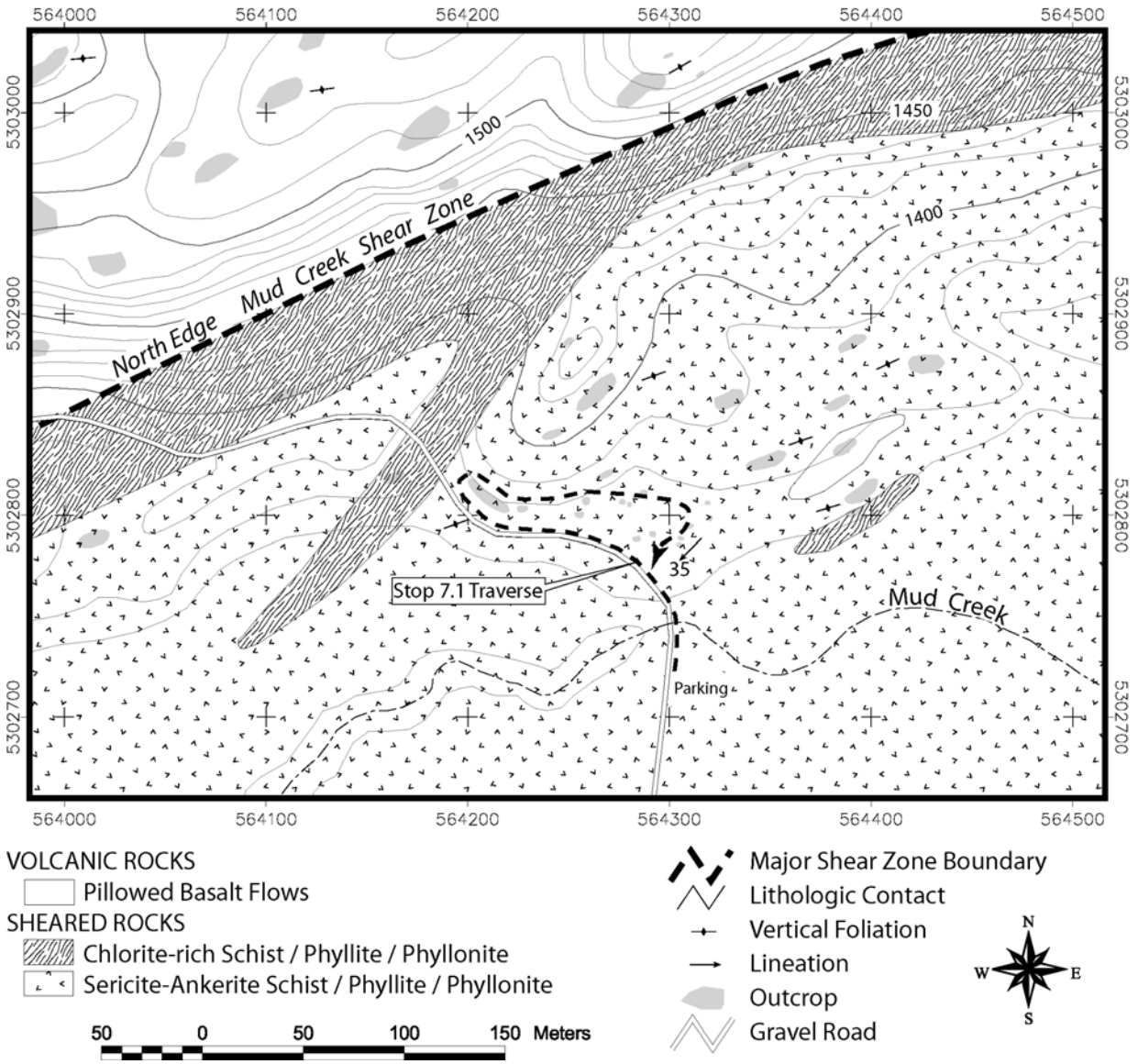


Figure 7-9. Simplified geology and field trip traverse map of stop 7-1, geology from Peterson and Patelke, 2004.

STOP 7-2: SECTION 6 GOLD PROSPECT

Location: (SE, SE, Sec. 6, T.62N., R.14W., NAD83 UTM 561895E, 5303410N)

Description: Gold mineralization within the Section 6 prospect is dominantly associated with quartz-pyrite veins and stockworks in feldspar ± quartz porphyry. The stockwork-style mineralization is generally not apparent until one takes a detailed look at the outcrops, then the thin quartz veining is seen seemingly everywhere. The highest-grade gold assays taken to date occur in areas of strong sulfide oxidation (sulfide burn) in the porphyry. Mineralization also occurs in sulfidized and epidote-altered basaltic rocks adjacent to the felsic porphyries, and may

locally contain significant quantities of chalcopyrite. In addition to porphyry and basaltic rocks, thin dikes and sills of diorite, peridotite, and inclusion-rich lamprophyre will be observed during the field trip stop. Outcrop photographs and a geologic map of the Section 6 prospect area are presented in Figures 7-10 and 7-11, and a table of anomalous gold assays is given in Table 7-1.

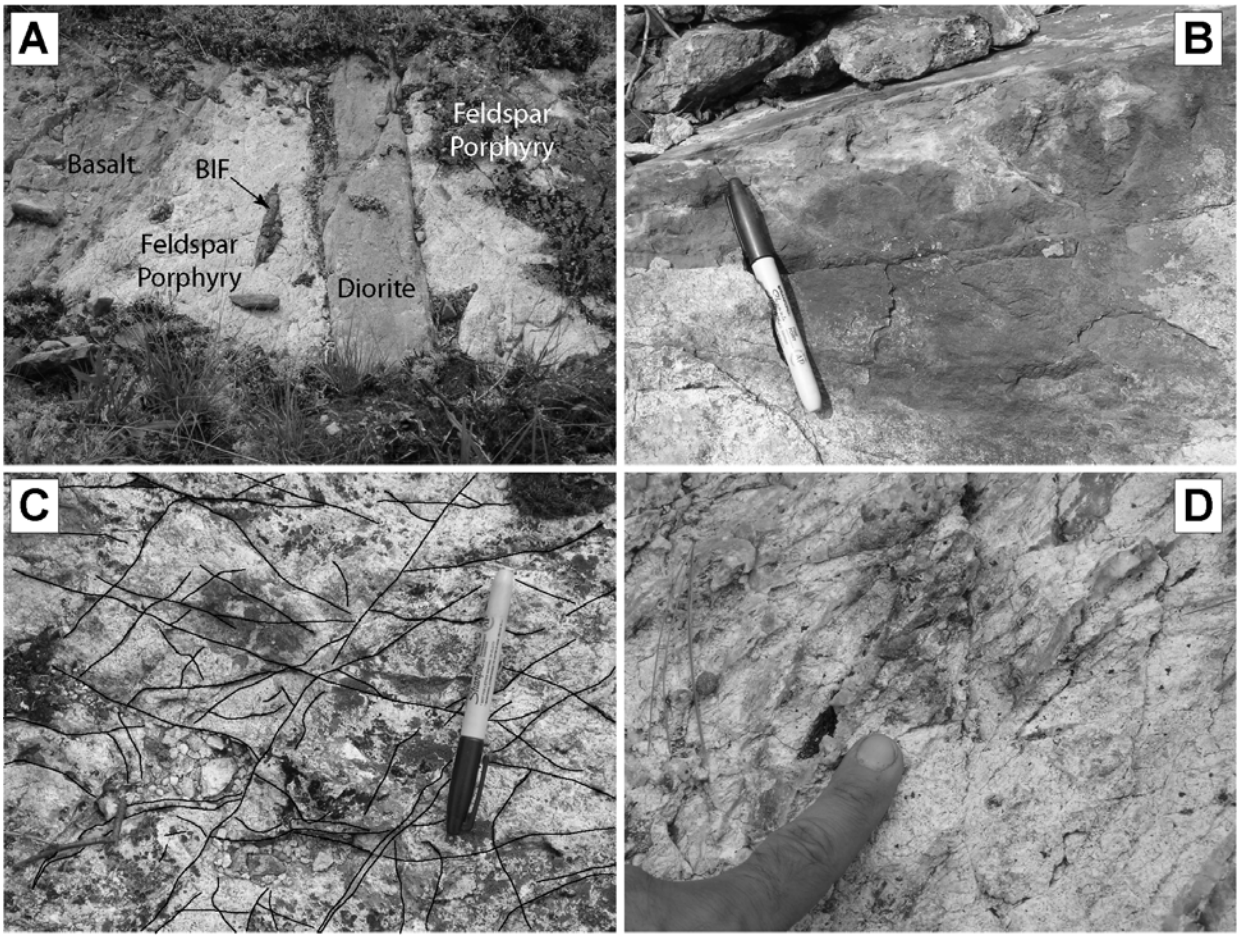


Figure 7-10. Outcrop photographs from the Section 6 prospect. (A) 1 Ft.-wide dioritic dike near a basalt-porphyry (with BIF inclusion) contact; (B) strong sulfide burn within feldspar porphyry; (C) stockwork quartz-veins (black lines) and local sulfide burn within feldspar porphyry; and (D) sulfide burn in quartz-veined feldspar porphyry.

Table 7-1. Anomalous gold assays from the Section 6 prospect.

Description	Au (ppb)
Siliceous, sericitic and strongly pyritic basalt	1,460
Carbonatized quartz-feldspar porphyry with pyrite	1,360
Carbonatized feldspar porphyry with pyrite and quartz veins	1,135
Pyritic selvage about a 3' quartz vein in basalt	940
Pyritic selvage about a 3' quartz vein in basalt	890
Epidote-silica-hematite altered feldspar porphyry with pyrite-rich veins	860

Description	Au (ppb)
Pyritic quartz-feldspar porphyry	855
Chloritic feldspar porphyry with ankerite, 1-3% pyrite, and quartz-ankerite-pyrite veins	840
Chlorite-epidote-silica altered basalt with pyrite and chalcopyrite	790
Carbonatized quartz-feldspar porphyry with pyrite	780
Chlorite-epidote-silica altered basalt with pyrite and chalcopyrite	445

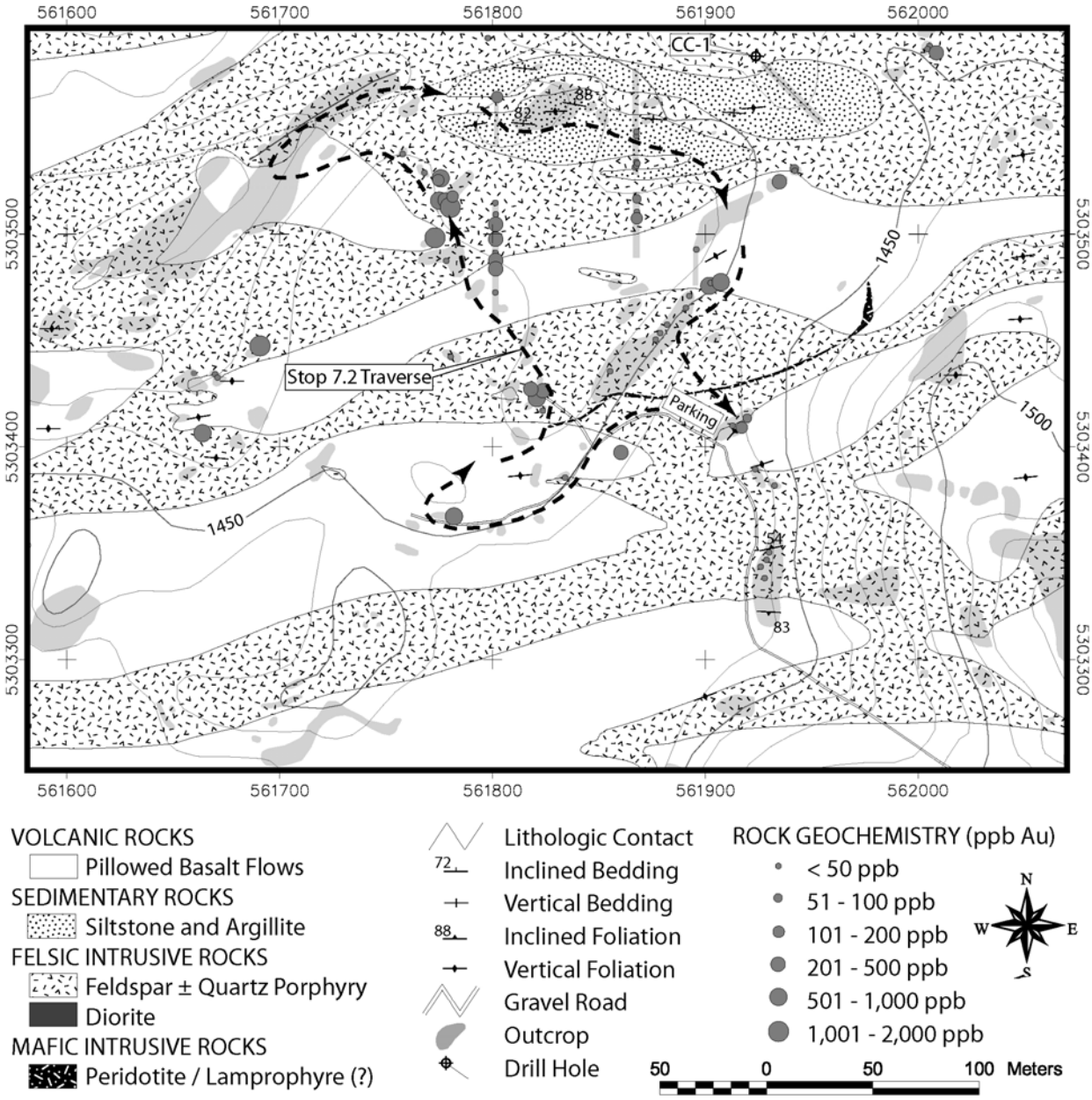


Figure 7-11. Simplified geology and field trip traverse map of stop 7-2, geology from Peterson and Patelke, 2004.

STOP 7-3: **SECTION 6 EAST GOLD PROSPECT**

Location: (SE, SE, Sec. 6, T.62N., R.14W., NAD83 UTM 562840E, 5302905N)

Description: Three features of the generalized Archean lode-gold ore deposit model are beautifully exposed along the traverse of this stop. These features include: (1) intense iron-carbonate alteration, (2) competency contrast and associated auriferous quartz veining, and (3) shear deformation. Anomalous gold was first discovered by soil sampling in this locality in 1988 by Chevron Resources. Gold mineralization is generally confined to quartz-pyrite veins in iron-carbonate altered feldspar-porphyry intrusive rocks. The felsic porphyries are generally located adjacent to linear zones of intense strain, and are believed to have intruded along these structural breaks during D_2 deformation. Outcrop photographs and a geologic map of the Section 6 East prospect area are presented in Figures 7-12 and 7-13, and anomalous gold assays in Table 7-2.

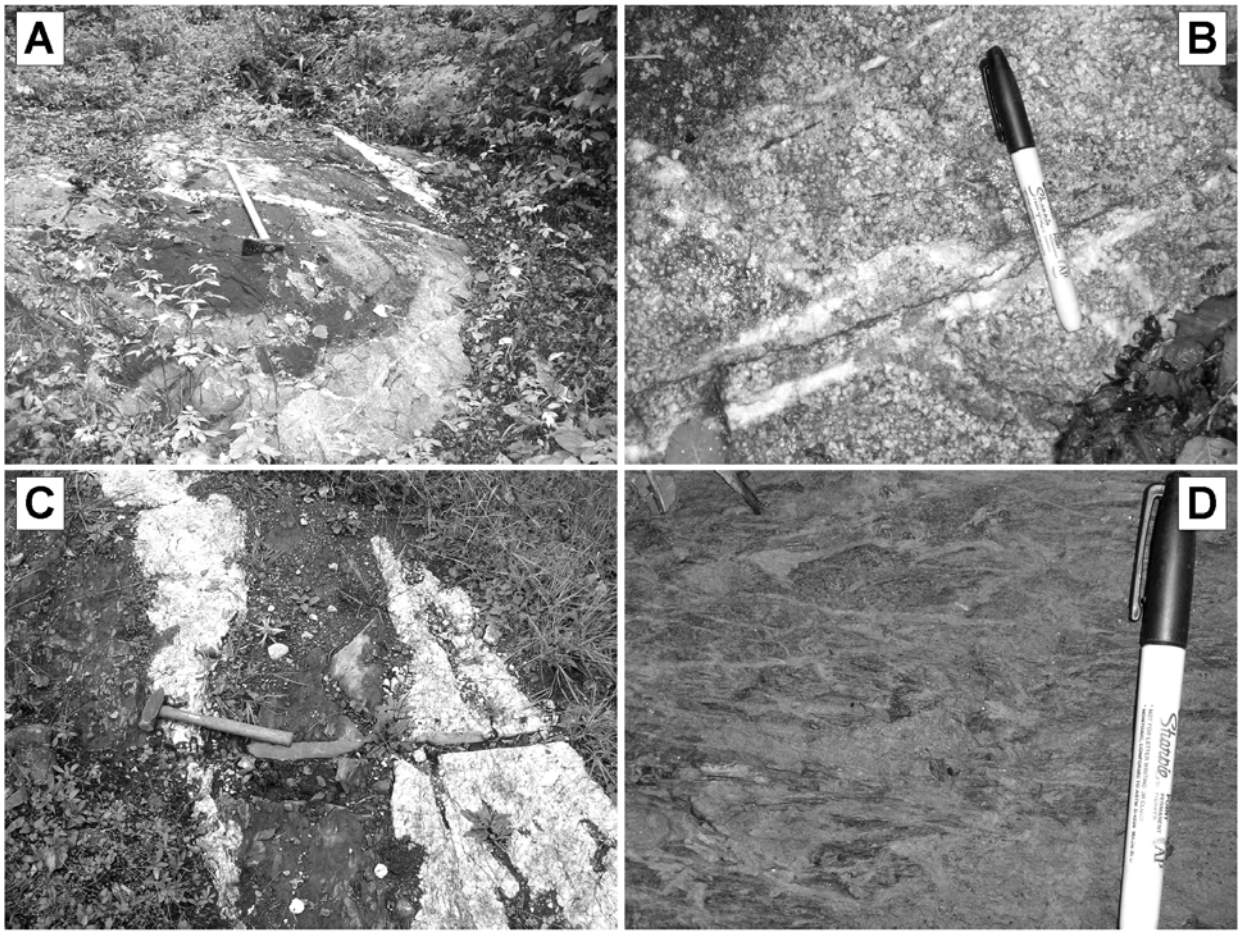


Figure 7-12. Outcrop photographs from the Section 6 East prospect. (A) iron-carbonate altered feldspar porphyry with quartz veins; (B) close-up of quartz-pyrite veins in carbonatized feldspar porphyry; (C) large quartz ± pyrite ± galena veins in massive iron carbonate; and (D) intense iron carbonate alteration

Table 7-2. Anomalous gold assays from the Section 6 East prospect.

Description	Au (ppb)	Description	Au (ppb)
6" quartz vein in sheared feldspar porphyry	6,210	Soil Sample	1,520
Feldspar porphyry with ankerite, pyrite, and quartz veins	2,400	Soil Sample	622
Feldspar porphyry with ankerite, pyrite, and quartz veins	1,760	Soil Sample	615
Feldspar porphyry with local fracturing and shearing	1,380	Soil Sample	294
Feldspar porphyry with a rusty quartz-pyrite vein	1,130	Soil Sample	170
Ankerite-sericite altered basalt with pyrite	1,030	Soil Sample	101

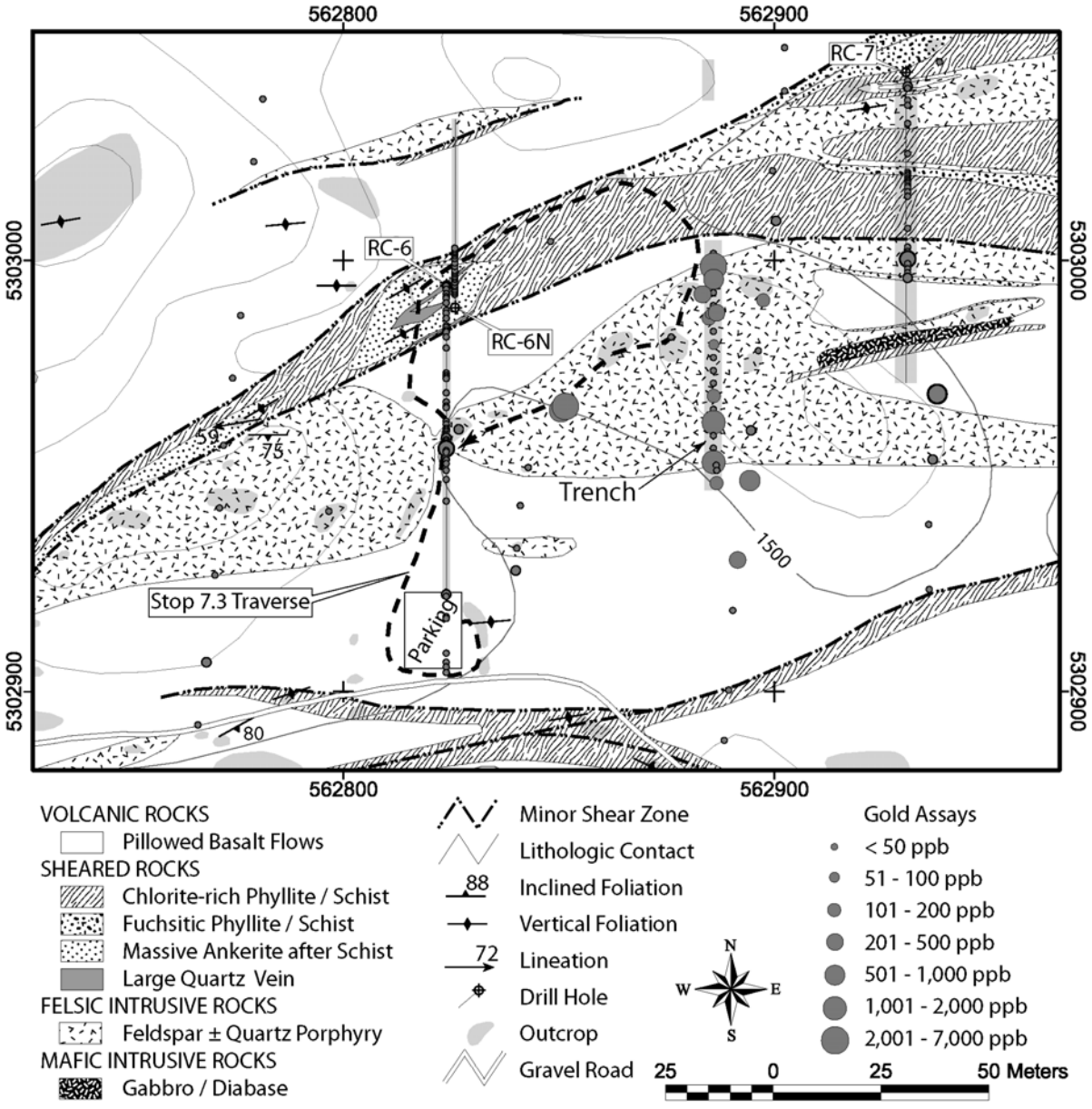


Figure 7-13. Simplified geology and field trip traverse map of stop 7-3, geology from Peterson and Patelke, 2004.

STOP 7-4: KERR MCGEE GOLD PROSPECT

Location: (SE, SE, Sec. 31, T.63N., R.14W., NAD83 UTM 562480E, 5304310N)

Description: The Kerr McGee gold prospect is hosted within an extensive zone of highly strained rocks, interpreted to be a subsidiary structure associated with the Mud Creek shear zone. Moderate to high-grade gold mineralization at the Kerr McGee prospect occurs within multiple thin (0.2 – 2.0 meter) zones of quartz-sericite-ankerite-pyrite ± green mica ± tourmaline schist hosted by an extensive zone of essentially gold-barren chlorite-rich schist. Thin and probably boudined iron-formation horizons occur locally in the chlorite-rich schist, and locally are strongly mineralized in this area. Mineralized zones locally contain extensive foliation and shear parallel quartz, ankerite, and/or quartz-ankerite veins, and may widen in zones of silicification. The style of gold mineralization exposed in the Kerr McGee prospect is similar to both the Clear Cut (~½ mile west) and Railroad Zone (1½ miles east) prospects. In fact, the sericitic zone that hosts the mineralization may have continuity to both of these other prospects. The field trip traverse will take us to numerous outcrops in the prospect area, and discussions will highlight the style of mineralization, possible continuity to other prospects, and shear deformation. Anomalous gold assays from the Kerr McGee prospect are presented in Table 7-3. Outcrop photographs and a geologic map of the prospect area are presented in Figures 7-14 and 7-15 respectively.

Table 7-3. Anomalous gold assays from the Kerr McGee prospect.

Description	Au (ppb)	DDH	Interval (ft)
1' wide zone of quartz-sericite-ankerite-pyrite tourmaline schist	8,010		
Quartz-pyrite-sericite knot in ankeritic sericite schist	5,030		
1' wide zone of quartz-sericite-ankerite-pyrite tourmaline schist	4,991		
1' wide zone of sericite-quartz-pyrite-ankerite phyllite	4,660		
1' wide zone of sericite-quartz-pyrite-ankerite phyllite	4,410		
1' wide zone of sericite-quartz-pyrite-ankerite phyllite	3,430		
0.5' wide zone of sericite-quartz-pyrite-ankerite phyllite	2,770		
1.1' wide zone of sericite-quartz-pyrite-ankerite phyllite	2,740		
Sericite-chlorite schist	2,440		
0.33' wide zone of sericite-quartz-ankerite-pyrite phyllite	1,675		
0.5' wide zone of sericite-quartz-ankerite-pyrite phyllite	1,140		
Road cut of siliceous quartz-sericite-pyrite schist with quartz veins	980		
1.2' wide zone of sericite-quartz-pyrite-ankerite phyllite	695		
1.1' wide zone of sericite-quartz-pyrite-ankerite phyllite	685		
Chlorite-sericite-ankerite schist with 10-20% pyrite	1,980	RC-3	1.0
Chlorite-quartz-ankerite schist with pyritic (5-10%) siliceous zones	1,470	RC-3	3.0
Chlorite-quartz-ankerite schist with pyritic (5-10%) siliceous zones	1,040	RC-3	4.0
Chlorite-quartz-ankerite schist with 1-5% pyrite	620	RC-3	7.0
Quartz-ankerite-pyrite (2-3%) vein	613	RC-3	1.0
Chlorite-sericite-ankerite schist with 10-20% pyrite	448	RC-3	1.0
Brecciated ankeritic-pyritic (5-10%) chert	369	RC-3	2.0
Sericite-chlorite-quartz-ankerite-pyrite (5%) schist	334	RC-3	2.0

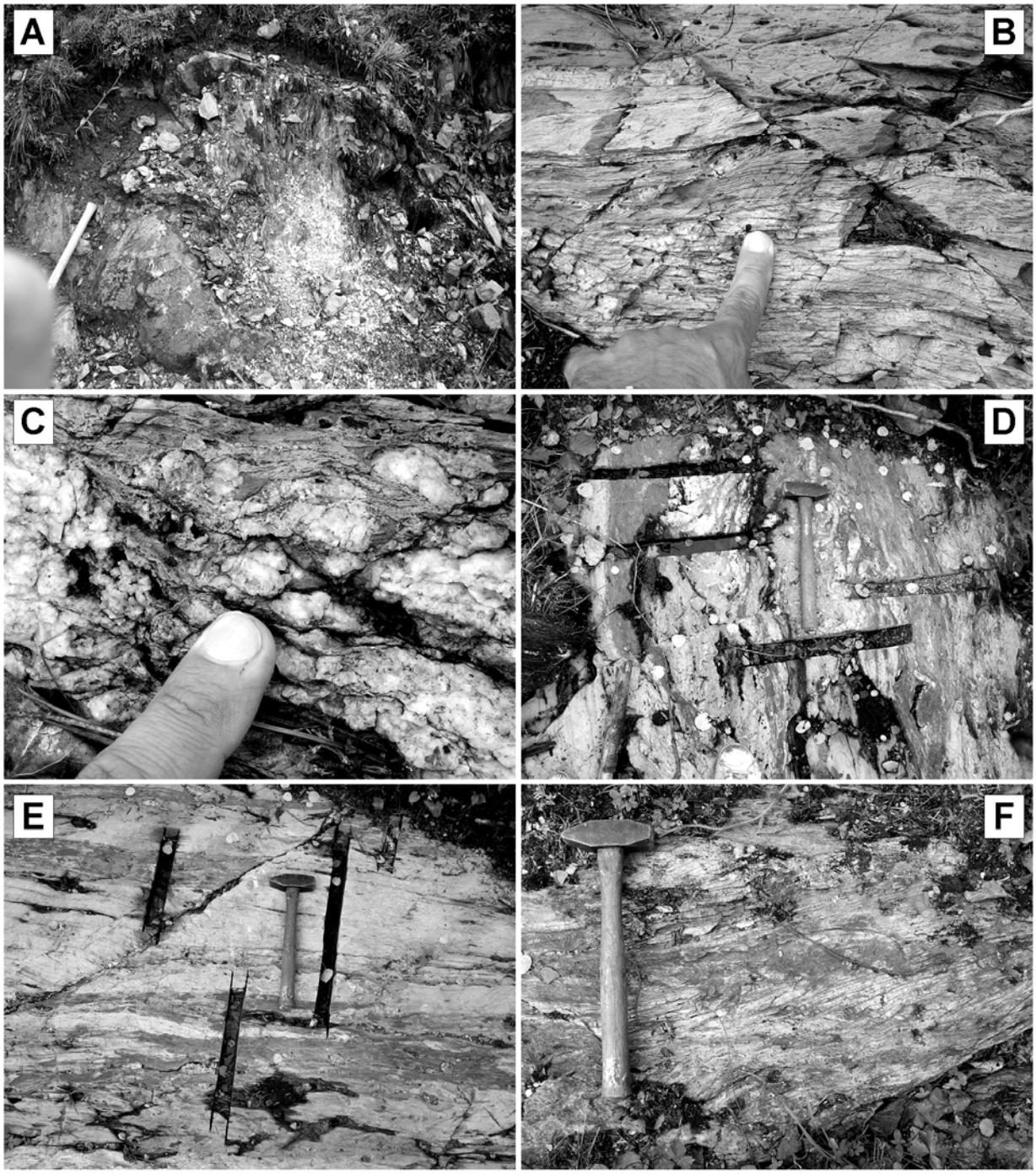


Figure 7-14. Outcrop photographs from the Kerr McGee prospect. (A) Road side outcrop of quartz-sericite-pyrite-ankerite-green mica-tourmaline schist; (B) minor northwest-trending fault offsetting highly foliated chloritic and sericitic schists; (C) close-up view of complex deformation fabric associated with quartz veins and knots; (D) channel samples taken across thin mineralized zones; (E) more channel samples taken across thin mineralized zones in sericite-rich schist, barren chlorite-rich schist at the bottom of the photograph; and (F) dextral shear fabrics in a mineralized zone at the east end of the outcrop area.

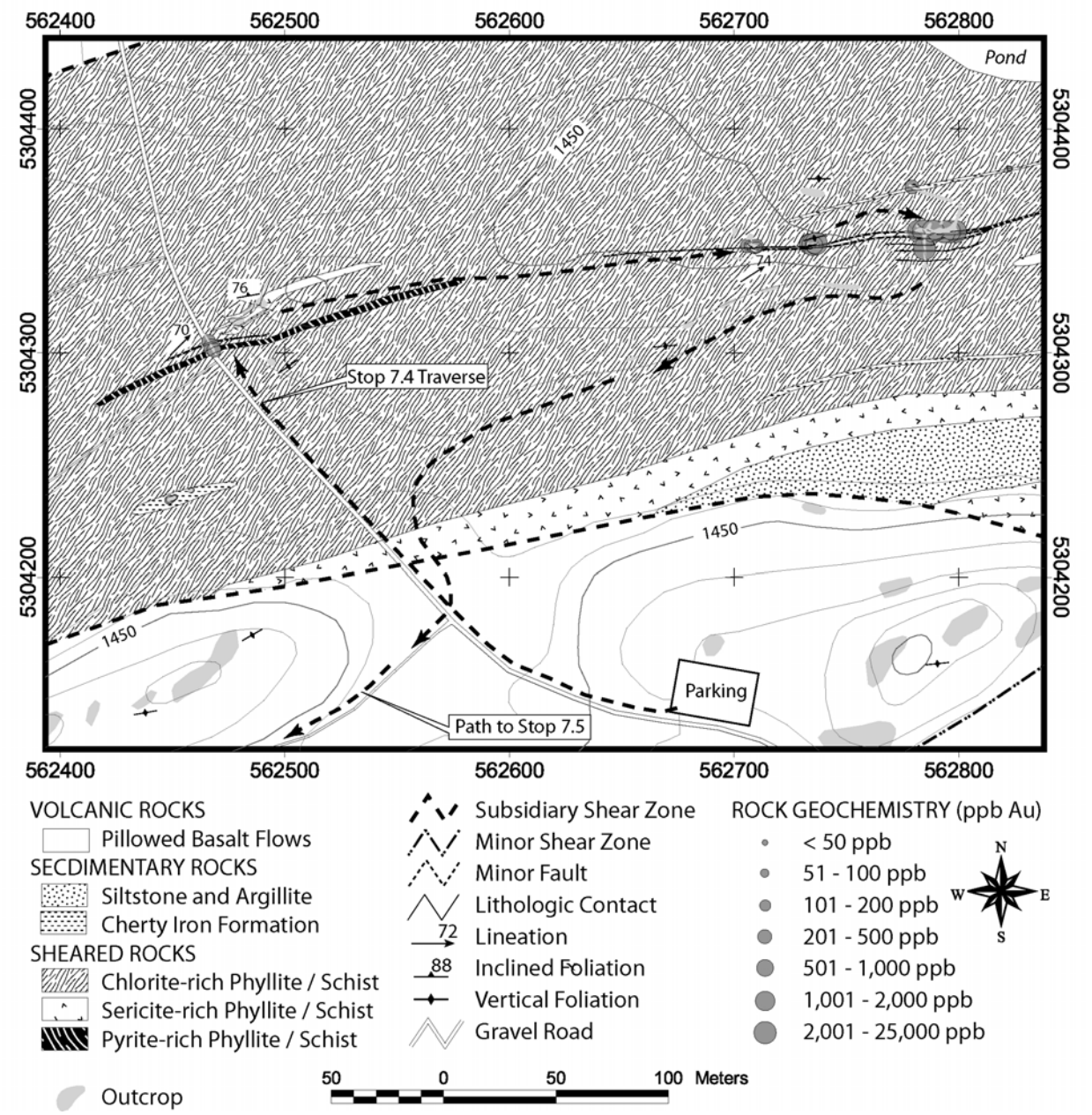


Figure 7-15. Simplified geology and field trip traverse map of stop 7-4, geology from Peterson and Patelke, 2004.

Three-dimensional visualization (Fig. 7-16) of the detailed lithological and structural mapping by Peterson and Patelke, (2004) within the Kerr McGee prospect area reveals important information that can be used to design drilling plans that significantly increase the chance of intersecting gold mineralization exposed in outcrop at the surface. For example, drill hole RC-3, which is located 100 meters east of the map presented in Figure 7-15, was drilled due north (at a dip of 45º) and targeted to intersect the mineralization exposed in outcrop at the Kerr McGee

showing. Chevron Resources drilled this hole in 1987, at the western boundary of their lease property (the Kerr McGee prospect was then held by Kerr McGee). Detailed structural mapping in these outcrops reveals that the rocks within the mineralized zone have moderate to strong elongation and intersection (foliation and shear planes) lineations trending 60º and dipping northeast at 72º. The best interpretation of the down-dip orientation of the mineralized zone is this lineation trend and plunge, and drill hole RC-3 never intersected the mineralized zone. A three-dimensional view of these relationships is given in Figure 7-16.

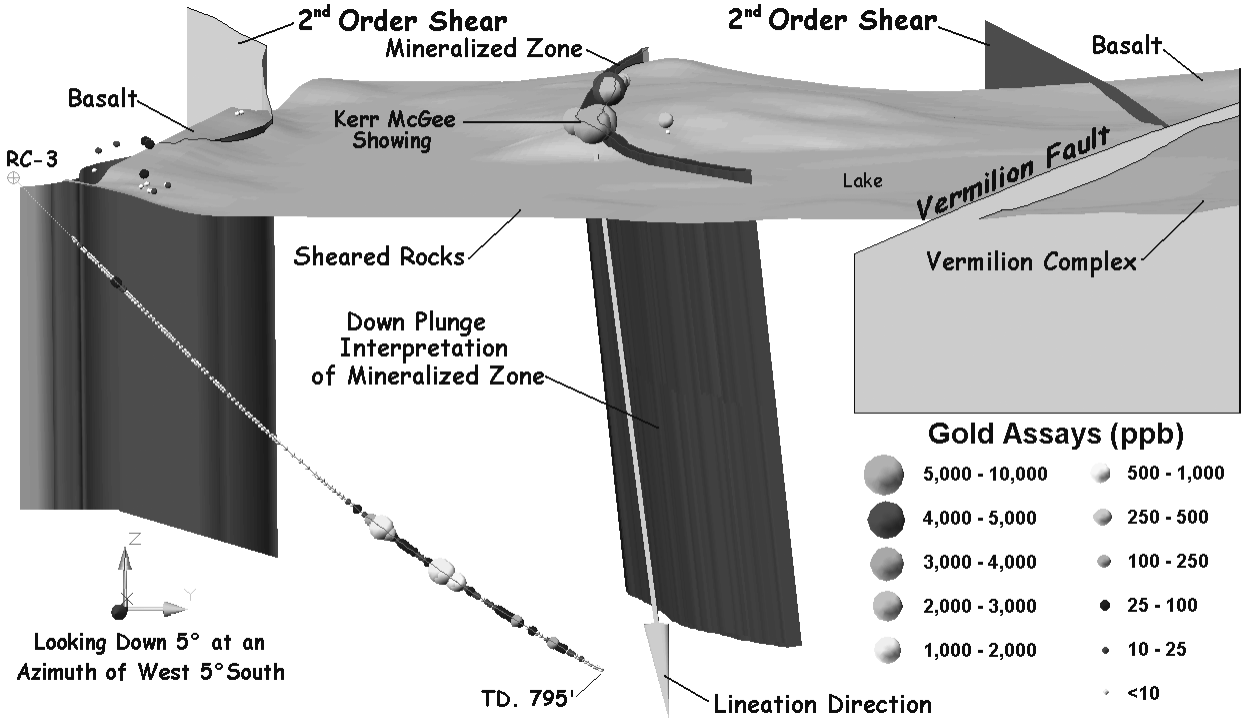


Figure 7-16. Three-dimensional view of the relationship between structural boundaries, the mineralized zone exposed on the surface at the Kerr McGee prospect, and drill hole RC-3. Upward extension to the surface of the two anomalous zones (> 1,000 ppb gold) intersected in hole RC-3 would place these zones in the black spruce and cedar swamp located south-southeast of the prospect.

As briefly described in the preceding paragraphs, the sericitic-schist hosted style of mineralization at the Kerr McGee prospect is similar to the mineralization at both the Clear Cut and Railroad Zone prospects, and may form a continuous zone of anomalous to ore-grade gold mineralization over a strike length of > 2.5 miles. By analogy to many Canadian and Australian Archean lode gold deposits, the odds of discovering high-grade zones within this trend would be significantly increased if possible future exploration ventures utilized structural elements exposed in outcrop. A simplified geologic map encompassing the highly strained zone that hosts the Clear Cut, Kerr McGee, and Railroad Zone gold prospects is presented in Figure 7-17. In addition, the trend and plunge of measured lineations are given in both the plan map of Figure 7-17 and the underlying east-west cross-section. Although the field trip will not visit the Clear Cut

and Railroad Zone prospects, anomalous gold assays from these associated prospects are presented in Tables 7-4 and 7-5 respectively.

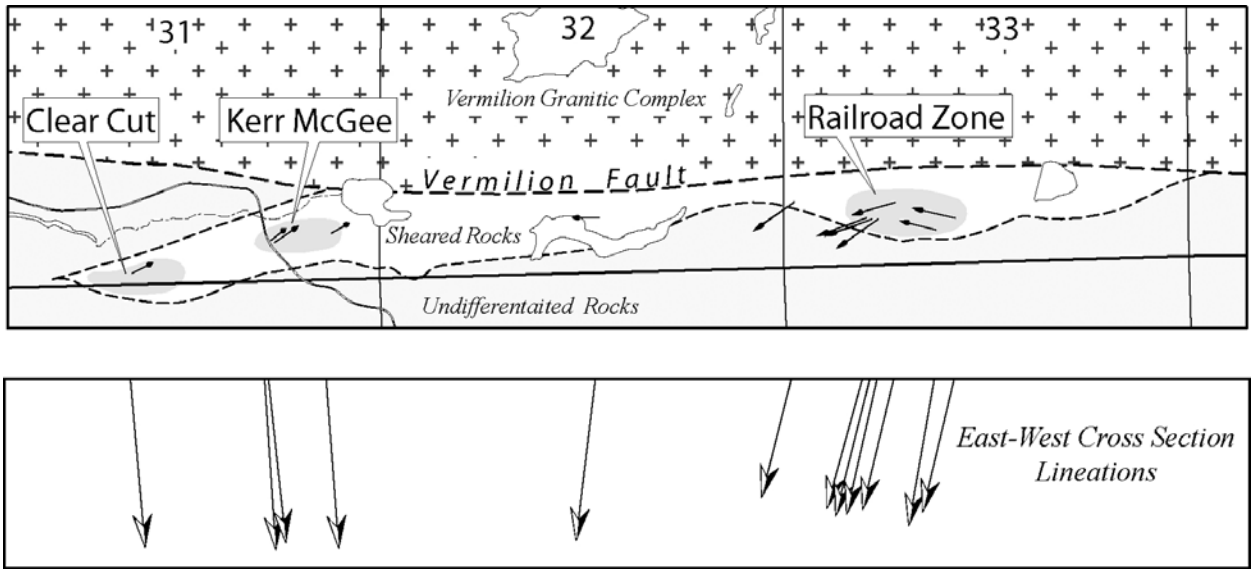


Figure 7-17. Highly simplified geologic and lineation map of the highly strained shear zone that hosts the Clear Cut, Kerr McGee, and Railroad Zone gold prospects. The total lengths of the lineation arrows depicted in the east-west cross-section (lower portion of the figure) are all equal, their apparent difference in length is related to their individual trend and plunge in relation to the view angle of the section.

Table 7-4. Anomalous gold assays from the Clear Cut prospect.

Description	Au (ppb)	DDH	Interval (ft)
2' wide zone of quartz-sericite-pyrite schist	22,000		
Quartz-sericite-pyrite schist	7,100		
Quartz-sericite-pyrite schist	2,940		
Quartz-sericite-pyrite schist	1,970		
Quartz-ankerite vein with a sericite-pyrite selvage in ankerite-chlorite schist	1,970		
1/2" wide quartz-pyrite-arsenopyrite vein	598		
Oxide-facies iron-formation with 10% pyrite and 5% arsenopyrite	590		
Quartz-sericite-pyrite schist	568		
Oxide-facies iron-formation with 1-2% cross-cutting pyrite	440		
Oxide-facies iron-formation with 10% pyrite and 5% arsenopyrite	402		
Chlorite-pyrite phyllite with quartz veining and 5% pyrite	1,510	RC-5	2.0
Chlorite phyllite with trace to 5% pyrite	715	RC-5	2.0
Chlorite-pyrite phyllite with quartz veining and 1% pyrite	550	RC-5	5.0
Chlorite-pyrite phyllite with quartz veining and 5% pyrite	528	RC-5	2.0
Chlorite-pyrite phyllite with quartz veining and 5% pyrite	471	RC-5	2.0

Table 7-5. Anomalous gold assays from the Railroad Zone prospect.

Description	Au (ppb)	DDH	Interval (ft)
Grab sample from large boulder encased in the roots of an overturned tree	37,324		
Sericite-chlorite-ankerite-pyrite phyllite with quartz-ankerite-pyrite veins	17,100		
4' Sericite-chlorite-ankerite-pyrite phyllite with quartz-ankerite-pyrite veins	13,623		
Chlorite-ankerite-pyrite schist	3,110		
Siliceous sericite schist with ankerite and pyrite	2,670		
Chlorites schist with pyrite and sericite	2,410		
Chlorite schist with pyrite and sericite	970		
Chlorite schist with pyrite and sericite	960		
Siliceous chlorite schist with pyrite	960		
Massive pyrite lens in sericite-ankerite schist	776		
Chlorite-ankerite-pyrite phyllite with quartz-ankerite veins	705		
Chlorite schist with pyrite and sericite	620		
5' zone of very rusty and weathered, sericite-ankerite-quartz-pyrite phyllite	605		
Chlorite-ankerite phyllite with numerous quartz-ankerite-pyrite veins	777	V-2	1.5
Sericite-quartz-ankerite-pyrite phyllite with 5% pyrite	713	V-4	1.8
Quartz-sericite-pyrite phyllite with 5% pyrite	591	V-1	6.0

STOP 7-5: PACMAN POND GOLD PROSPECT

Location: (SE, SE, Sec. 6, T.62N., R.14W., NAD83 UTM 562225E, 5303780N)

Description: The Pac Man Pond gold prospect was explored by Noranda in the late 1980s and early 1990s. They completed a quite extensive program that consisted of geological mapping, bedrock sampling, till prospecting, trenching, and drilling. Three rock types, that include pillowed basalt, iron-formation, and intrusions of feldspar ± quartz porphyry, dominate the prospect geology, with rare occurrences of argillaceous and graphitic sediments as well as carbonate facies iron-formation intersected in drill holes. Gold mineralization is hosted dominantly by brecciated, quartz-pyrite veined, and sulfidized iron-formation, with locally anomalous gold in sulfidized basalt and porphyry. The outcrops observed on the field trip will be concentrated on one of the areas that Noranda completed extensive trenching and gold assaying. Outcrop photographs from the Pac Man Pond gold prospect are given in Figure 7-18, and a geologic map of the Pac Man Pond gold prospect area is presented in Figure 7-19. Anomalous gold assays from bedrock exposures and Noranda drill hole intersections are given in Table 7-6.

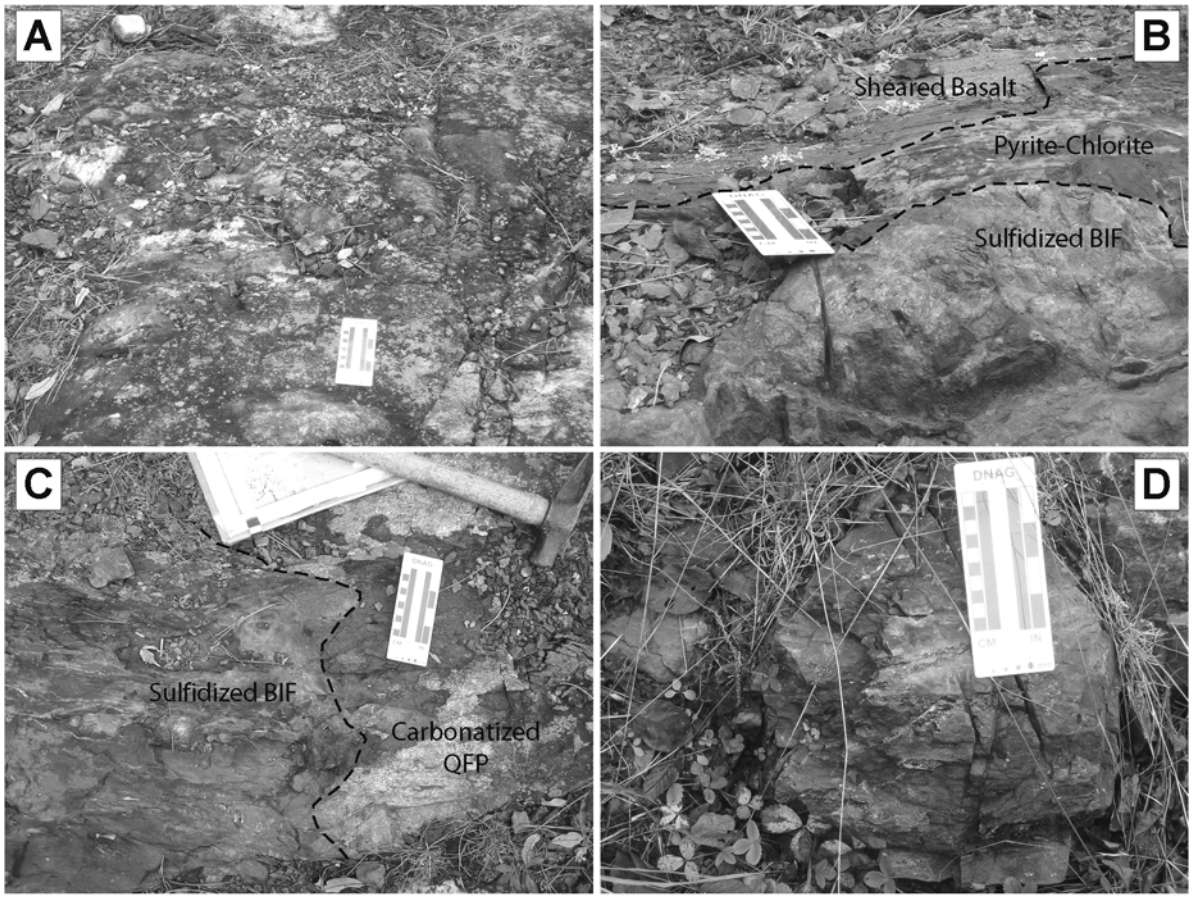


Figure 7-18. Outcrop photographs from the Pac Man Pond prospect. (A) Strongly iron-carbonate altered quartz-feldspar porphyry; (B) sheared and laminated pyrite and chlorite at the contact between foliated to sheared basalt and sulfidized iron-formation; (C) irregular contact between carbonatized quartz-feldspar porphyry and sulfidized iron-formation; and (D) strong sulfide burn on highly sulfidized iron-formation.

Sawed samples of gold-bearing, pyritic iron-formation reveal one of the classic chemical models for the precipitation of gold out of hydrothermal solutions. In this model, when silica-saturated auriferous hydrothermal fluids (with gold carried by HS complexes) encounter a rock rich in FeO (such as Fe-tholeiitic basalt or iron-formation), quartz, pyrite and gold are precipitated out of solution. The classic chemical equation is presented below.

$$2HAu(HS)_{2\ aq} + \tfrac{1}{2}O_{2\ aq} + 2FeO_{\ s} \leftrightarrow 2FeS_{2\ s} + 2Au_{\ s} + 3H_2O_{\ aq}$$

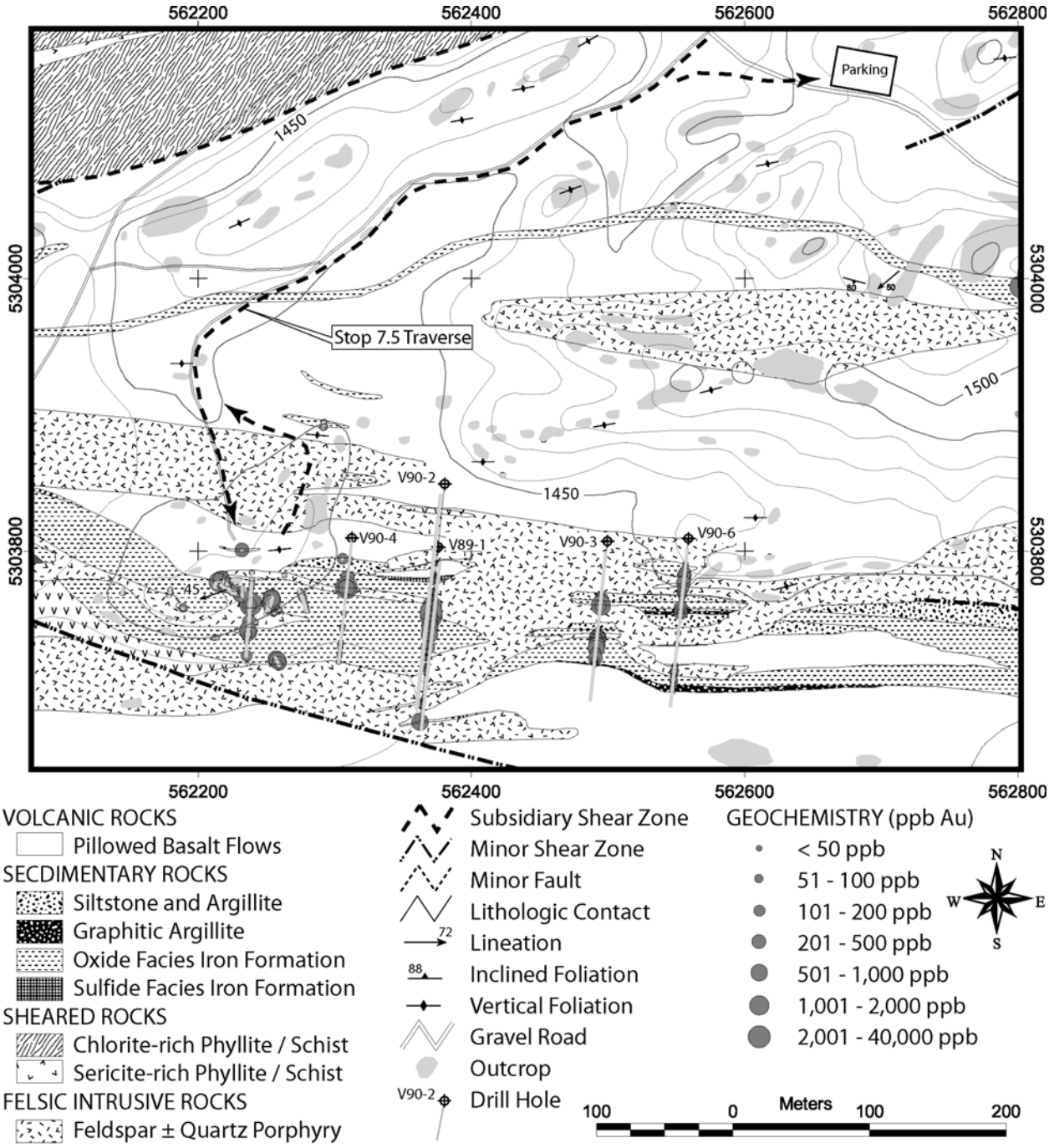


Figure 7-19. Simplified geology and field trip traverse map of stop 7-5, geology from Peterson and Patelke, 2004.

Table 7-6. Anomalous gold assays from the Pac Man Pond prospect.

Description	Au (ppb)	DDH	Interval (ft)
Large angular boulder of pyrite-rich (5-50%) iron-formation	2,701		
Sulfidized iron-formation	2,190		
Siliceous, pyritic iron-formation with limonite	1,940		
Siliceous, pyritic iron-formation	1,810		
Sulfidized iron-formation	1,750		
Sulfidized iron-formation with quartz-pyrite veins	1,635		
Limonite stained iron-formation with 5-15% pyrite	1,020		
Sulfidized (10-70% pyrite) basalt	985		
Siliceous, pyritic iron-formation	890		
Iron-formation with trace pyrite	875		
Sulfidized iron-formation	845		
Laminated iron-formation with trace pyrite	845		
Pyrite-rich iron-formation near contact with quartz-feldspar porphyry	815		
Sulfidized iron-formation with 5-20% pyrite	811		
Quartz-feldspar porphyry with a rusty quartz vein	785		
Laminated iron-formation with trace pyrite	760		
6' wide zone of sulfidized iron-formation with 3-10% pyrite	758		
Mineralized iron-formation with 5-15% pyrite	705		
Brecciated iron-formation with 5-20% pyrite in matrix	28,460	V89-1	2.0
Brecciated iron-formation with 5-20% pyrite in matrix	5,400	V89-1	1.7
Iron-formation with 1-2% pyrite in veinlets	4,212	V-90-2	3.5
Brecciated iron-formation with quartz-pyrite (10%) veins	2,255	V-90-4	2.0
Thin-bedded and folded iron-formation with 3% pyrite	1,747	V-90-7	1.8
Brecciated iron-formation with chlorite-quartz-pyrite (5-10%) matrix	1,364	V-90-2	1.8
Brecciated iron-formation with 5-10% pyrite	1,026	V-90-3	3.2
Weakly brecciated iron-formation with 3-10% pyrite in matrix	960	V89-1	2.3
Fine-grained mafic dike	856	V-90-4	1.0
Quartz-feldspar porphyry with trace pyrite	856	V-90-4	1.0
Oxidized iron-formation with ankerite and 8% pyrite	782	V-90-3	2.0

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FIELD TRIP 8

Geology and Mineralization of the Western Contact of the Duluth Complex, Partridge River and South Kawishiwi Intrusions, Northeastern Minnesota

Leaders:

Mark Severson

Natural Resources Research Institute, University of Minnesota Duluth

and

Jim Miller

Minnesota Geological Survey, University of Minnesota



Railroad-cut exposures in the Wetlegs area of the Partridge River Intrusion (Stop 8-3)

FIELD TRIP 8

Geology and mineralization of the western contact of the Duluth Complex, Partridge River and South Kawishiwi intrusions, northeastern Minnesota

by

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Natural Resources Research Institute, University of Minnesota Duluth
and

Jim Miller
Minnesota Geological Survey, University of Minnesota

INTRODUCTION

The Duluth Complex and associated Keweenawan intrusions in northeastern Minnesota constitute one of the largest mafic intrusive complexes in the world, second only to the Bushveld Complex of South Africa. These rocks cover an arcuate area of over 5,000 km^2 (Fig. 8-1) and give rise to two strong gravity anomalies (+50 & +70 mgal) that imply intrusive roots to more than 13 kilometers depth (Allen and others, 1997). The intrusive rocks of northeastern Minnesota were emplaced into a comagmatic volcanic edifice during formation of the Midcontinent Rift between 1108 and 1086 Ma.

The Keweenawan geology of northeastern Minnesota has been recently reinterpreted in a 1:200,000 scale geologic map (MGS Miscellaneous Map, M-119; Miller and others, 2001) and in a comprehensive report that focused on the geology and mineral potential of the Duluth Complex (MGS Report of Investigation, RI-58; Miller and others, 2002). The general description of the Duluth Complex given below is largely excerpted from that report.

The Duluth Complex is physically defined as a more or less continuous mass of mafic to felsic plutonic rocks that extends in an arcuate fashion from Duluth to nearly Grand Portage (Fig. 8-1). It is bounded by a footwall of predominantly Paleoproterozoic and Archean rocks, a hanging wall of largely mafic volcanic rocks and hypabyssal intrusions, and internally, it contains scattered bodies of strongly recrystallized mafic volcanic and sedimentary hornfels. Defining the Duluth Complex more genetically, it is composed of multiple discrete intrusions of mafic to felsic tholeiitic magmas that were episodically emplaced into the base of a comagmatic volcanic edifice in two general stages - an early stage at about 1108 Ma and an main stage at 1099 Ma. Within the Duluth Complex, four general rock series are distinguished on the basis of age, dominant lithology, internal structure, and structural position. These are:

Felsic Series – massive granophyric granite and smaller amounts of intermediate rock that occurs as a semicontinous mass of intrusions strung along the eastern and central roof zone of the complex and was emplaced during early stage magmatism (~1108 Ma).

Early Gabbro Series – layered sequences of dominantly gabbroic cumulates that occur in two major intrusions along the northeastern contact of the Duluth Complex and were also emplaced during early stage magmatism (~1108 Ma)

Anorthositic Series – a structurally complex suite of foliated, but rarely layered, plagioclase-rich gabbroic cumulates that was emplaced throughout the complex during main stage magmatism (~ 1099 Ma).

Layered Series – a suite of stratiform troctolitic to ferrogabboic cumulates that comprise at least 11 variably-differentiated mafic layered intrusions and occurs mostly along the base of the

Duluth Complex. These intrusions were emplaced during main stage magmatism, but generally after the anorthositic series.

The Partridge River (PRI) and South Kawishiwi (SKI) intrusions, the focus of this field trip, occur along the northwestern margin of the Duluth Complex (Fig. 8-1) and are two of the earliest-formed intrusions of the Layered Series. These two intrusions are most renown for hosting the largest tonnage of Cu-Ni sulfide ore in the world (Naldrett, 1997). The bedrock geology of the northwestern margin of the Duluth Complex, as portrayed on the regional geologic map of northeastern Minnesota (M-119; Miller and others, 2001), is shown in Figure 8-2. Detailed mapping in the Babbitt NE, Babbitt SE, and Babbitt SW quadrangles, which was conducted since the publication of M-119 (Miller, Severson, and Foose, 2002; and unpublished field mapping, Miller, 2002-03), has revealed a somewhat different picture of the geology that has yet to be incorporated into the regional map. Mapping in this area will continue this summer with the objective of producing 1:24,000-scale maps of all four of the Babbitt 7.5' quadrangles in the summer of 2005.

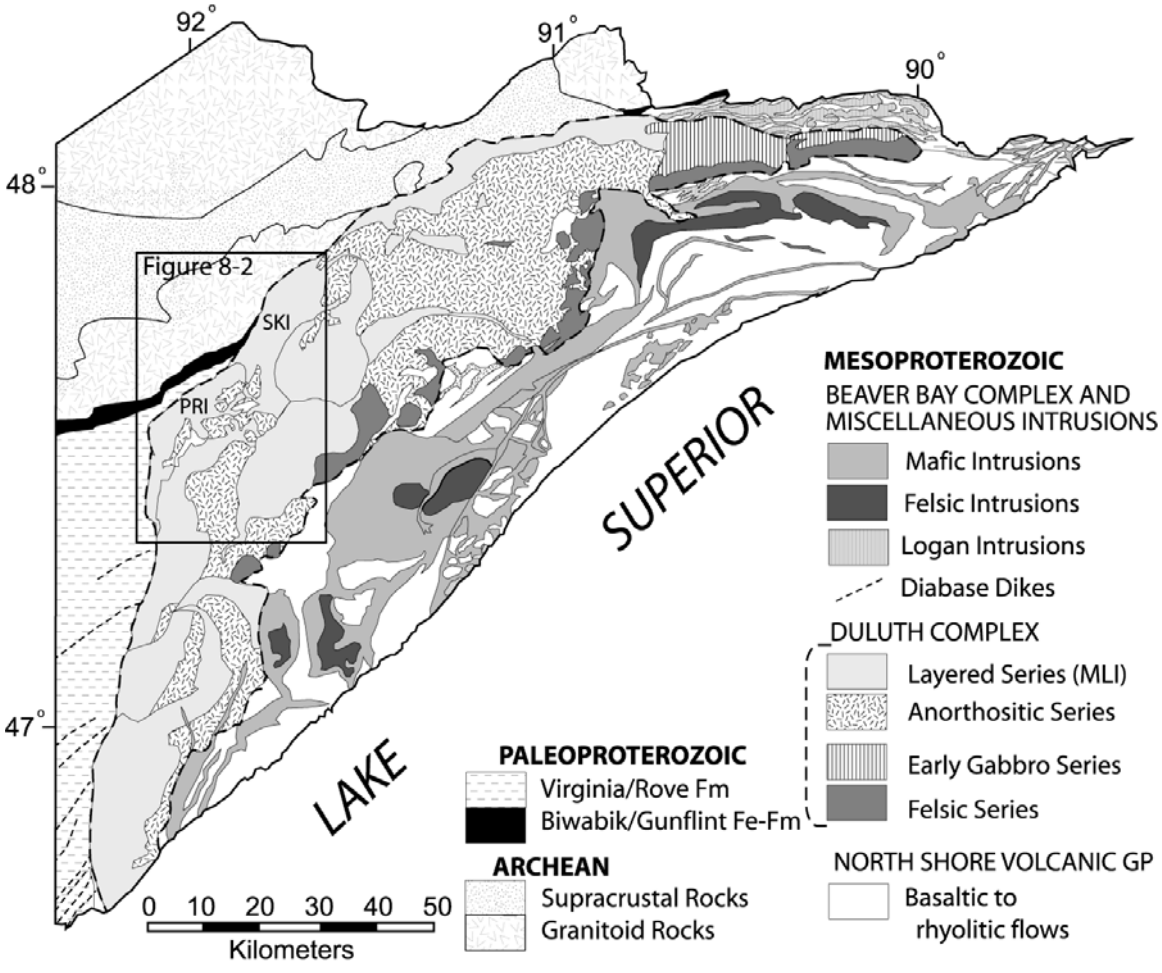


Figure 8-1. Generalized geology of northeastern Minnesota showing the field trip area. Partridge River intrusion (PRI) and South Kawishiwi intrusion (SKI) are labeled.

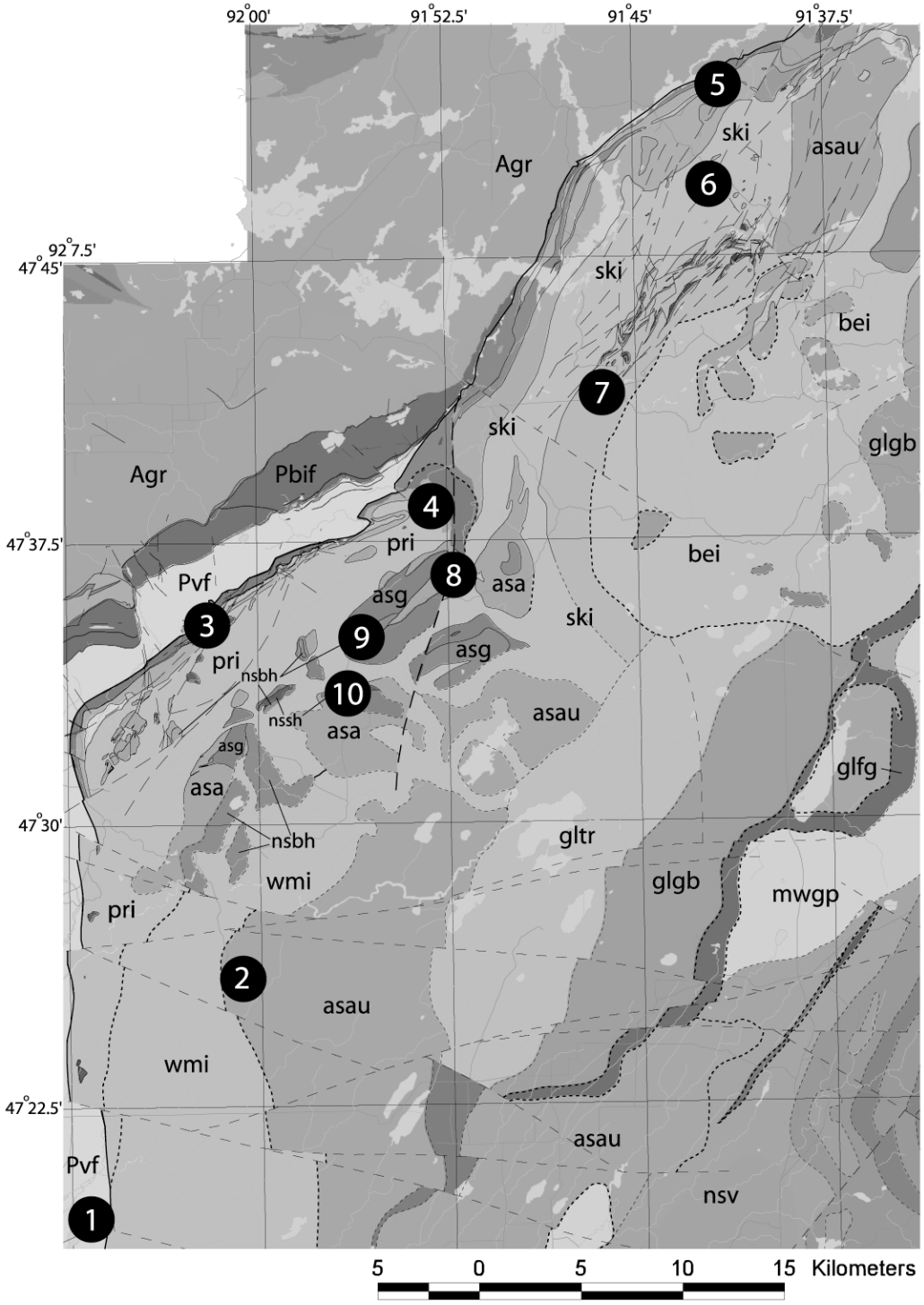


Figure 8-2. Geology of the northwestern margin of the Duluth Complex showing field stops for Field Trip 8 (see Miller and others, 2001 for more details). Geologic units: Agr - Giants Range granite, Pbif-Biwabik Iron Fm, Pvf - Virginia Fm, pri - Partridge R. intr., ski-South Kawishiwi intr., bei-Bald Eagle intr.; gltr-Greenwood Lake intr-troctolitic zone; glgb-Greenwood Lake intr, gabbroic zone; glfg-Greenwood Lake intr, ferrogabbroic zone; mwgp-Mount Weber granophyre, asau-anorthositic series undifferentiated; asa-anorthositic series, anorthositic rocks; asg-anorthositic series, gabbroic rocks; nsv-NSVG undifferentiated; nsbh-NSVG basaltic hornfels; nssh-NSVG sedimentary hornfels.

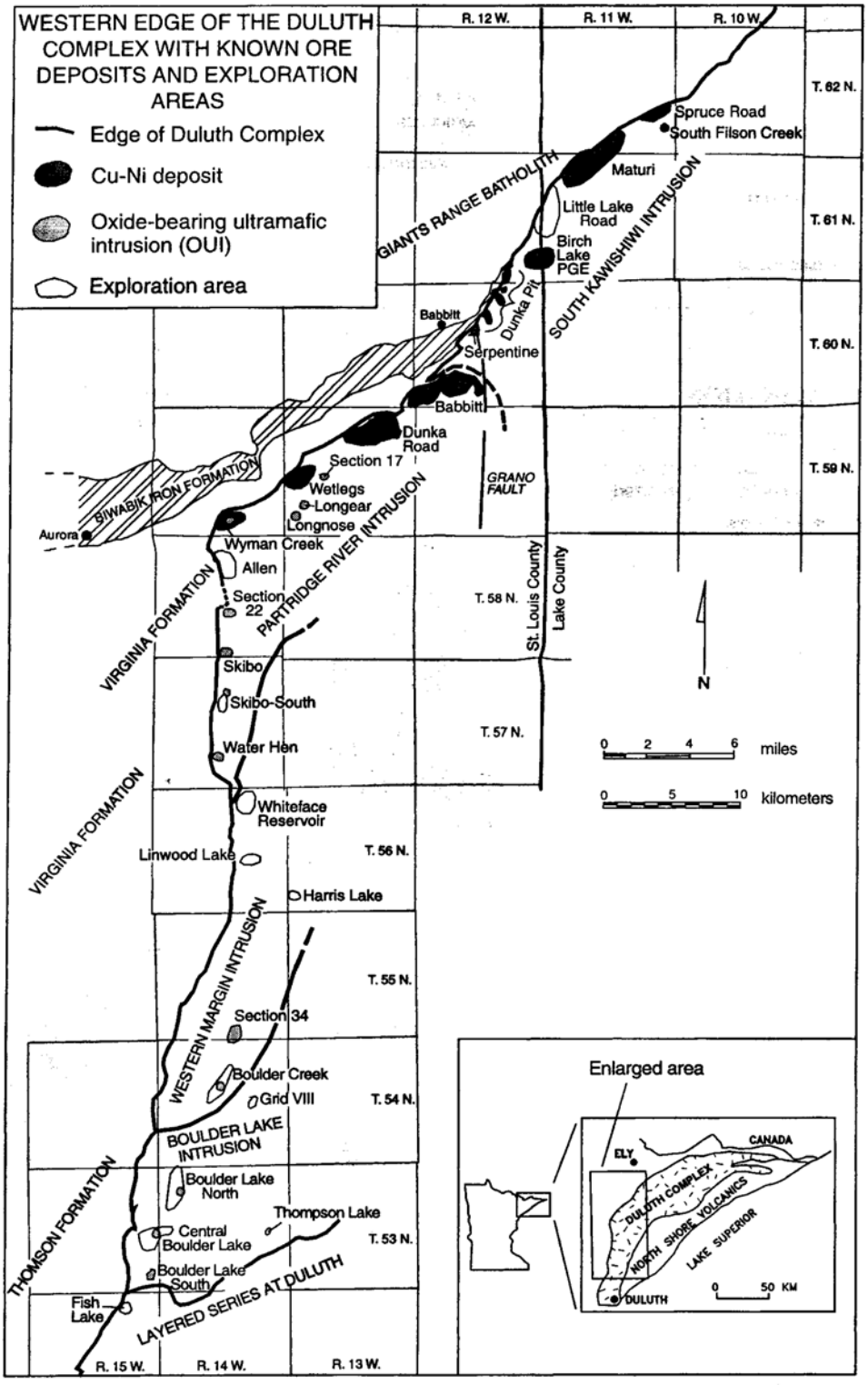


Figure 8-3. Location of copper-nickel sulfide deposits, Fe-Ti±V deposits, and other exploration areas along the western base of the Duluth Complex (from Miller and others, 2002; Fig. 2.3)

Partridge River Intrusion

The Partridge River intrusion (PRI, Bonnichsen, 1974) is perhaps one of the most well studied intrusions of the Duluth Complex because it hosts at least four sub-economic Cu-Ni deposits and at least seven potential Fe-Ti deposits. The PRI consists mainly of troctolitic cumulates that are exposed in an arc-shaped area that extends from the Waterhen Fe-Ti deposit area (T.57N., R.14W., Sec. 27) to the Babbitt Cu-Ni deposit (T.60N., R.12W., Secs 31-33; Fig. 8-3). Miller and Ripley (1996) estimate that the PRI is 2.5 km thick. The footwall of the PRI includes the Paleoproterozoic Virginia Formation (slate and graywacke), and to a lesser extent, the Biwabik Iron Formation. The top of the PRI is in complex contact with anorthositic rocks (unit asa, Fig. 8-2), gabbroic rocks (unit asg), mafic volcanic hornfels (unit nshb), and in one location, an unusual fine-grained and cross-bedded sedimentary hornfels (Colvin Creek hornfels of Bonnichsen, 1972; Patelke, 1996; unit nssh). This assemblage of anorthositic, gabbroic, and hornfelsic rocks are also present as large inclusions within the interior of the PRI (Severson and Miller, 1999) and are thought to represent earlier roof zone septa that were overplated and isolated by later injections of PRI magma.

The lower 900 meters of the PRI is known in great detail from the abundance of exploration drill core. This marginal zone, consisting of varied troctolitic and gabbroic rock types, is subdivided into seven stratigraphic units (Severson, 1991; 1994; Severson and Hauck, 1990; 1997; Geerts, 1991) that can be correlated over a strike length of 24 km (Fig. 8-4). All of these igneous units generally exhibit shallow dips (10-20°) to the southeast. The stratigraphy shown in Figure 8-4 is based on the relogging of almost 700 drill holes, or 672,000 feet of drill core. The units of the Partridge River marginal zone are briefly described below starting at the base of the PRI. Units I and III will be viewed at Stop 8-3. For locations of ore deposit areas mentioned below, see Figure 8-3.

Unit I - This lower unit consists of a heterogeneous mixture of ophitic troctolitic to gabbroic rocks that contain abundant inclusions of hornfelsic sedimentary footwall rocks and minor thin, discontinuous layers of melatroctolite and peridotite. It is the dominant sulfide-bearing member of the PRI. Noritic rocks are common at the basal contact and peripheral to sedimentary hornfels inclusions, probably due to contamination of the magma. An ultramafic interval, consisting of an oxide-bearing peridotite overlying an oxide-bearing pyroxenite, is present at the base of Unit I in areas where the PRI is in direct contact with the Biwabik Iron Formation (extreme southeastern portion of the Babbitt deposit).

Unit II - This unit exhibits considerable variation from one Cu-Ni deposit to the next. At the Dunka Road and Babbitt deposits, Unit II consists of homogenous troctolitic rocks, with minor sulfide mineralization, and a fairly persistent basal ultramafic layer that separates Unit II from Unit I. The upper contact with Unit III is gradational at the Dunka Road deposit. At the Wetlegs deposit, Unit II is either 1) a single ultramafic layer immediately beneath Unit III, and/or 2) the Wetlegs Layered Interval (Fig. 8-3) which consists of repeated, thin cyclic units that internally grade upward from ultramafic rock to troctolite. Still farther to the west at the Wyman Creek deposit, Unit II consists of a single ultramafic horizon that separates sulfide-bearing and heterogeneous troctolitic rocks of Unit I from homogenous troctolitic rocks of Unit IV.

Unit III – This unit consists of poikilitic leucotroctolite that commonly grades into poikilitic/ophitic augite troctolite. Because of its fine-grained texture and distinctive olivine oikocrysts that impart a mottled appearance, Unit III is a useful marker horizon, although it is absent from the Wyman Creek deposit and portions of the Babbitt deposit. Hornfelsed basalt inclusions, or roof rocks, are commonly associated with Unit III at the Dunka Road deposit. This relationship, and the highly gradational contact of Unit III

with Unit II, suggest that Unit III may have formed as a roof cumulate during crystallization of Unit II.

Unit IV - Homogenous ophitic augite troctolite that contains a local basal unit of ultramafic rock. Unit IV typically exhibits a highly gradational upper contact with Unit V. A cyclic sequence of alternating troctolitic and ultramafic layers, termed the Bathtub Layered Interval, is present in the Bathtub ore zone of the Babbitt deposit.

Unit V – Unit V consists of homogenous, coarse-grained leucotroctolite. At the Wyman Creek deposit, drill hole relationships suggest that Unit V cuts downward into the lower units (Fig. 8-3).

Units VI and VII - Each of these units consist of homogenous leucotroctolite that locally grades into ophitic augite troctolite; both also contain a fairly persistent ultramafic base. Several other similar units (Unit VIII and up) are present above these two units but they are not defined by drilling.

OUI - Several late-stage pegmatitic plugs and vertical lenses of oxide-bearing ultramafic intrusions (OUI) intrude the troctolitic rocks of the PRI. The acronym OUI was first used by Severson and Hauck (1990) to designate cross-cutting pegmatitic bodies of peridotite, melatroctolite, melagabbro, and clinopyroxenite that contain a high percentage of coarse-grained oxides (15-100%). Known OUI bodies in the PRI, with Fe-Ti potential, occur at Section 17, Longnose, Longear, Wyman Creek, Section 22, Skibo, Skibo-South, and Waterhen. In all instances, the OUI are spatially arranged along linear trends suggesting that structural control was important to their genesis. In addition, the Longnose-Longear-Section 17 group of OUIs are positioned over a window in the basal contact where the Biwabik Iron-Formation is the footwall rock. This relationship suggests a genetic link between assimilation of iron-formation at the basal contact and formation of OUI along a coincident fault zone.

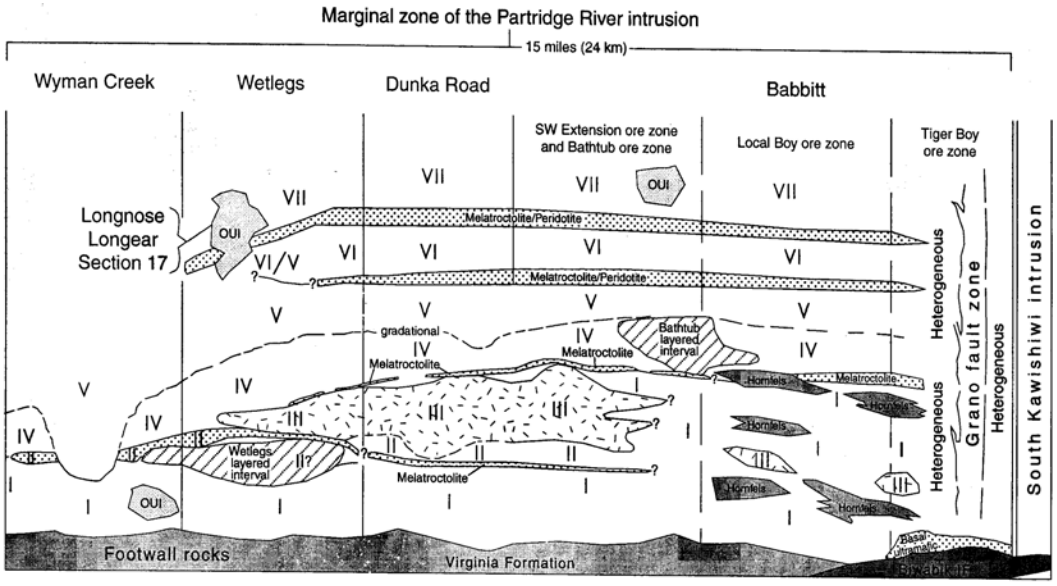


Figure 8-4. Generalized stratigraphy of the marginal zone of the Partridge River intrusion (modified from Severson, 1994). Stratigraphic relationships for the area between the Wyman Creek deposit and the Water Hen deposit are poorly understood and are not portrayed (taken from RI-58 figure 6.10).

Attributes of the more heterogeneous units (I and II) near the base of the PRI are interpreted to indicate rapid magma replenishment in a progressively developing magma chamber, and magmatic contamination from assimilated footwall rocks. The more homogeneous upper units (IV-VII), each floored by a persistent ultramafic layer, were probably emplaced later in a well-developed magma chamber. The ultramafic layers in the upper units, and abundant ultramafic layers of the Wetlegs and Bathtub layered intervals, probably represent the inception of episodic magma injection that crystallized more primitive ultramafic layers before mixing with the resident magma.

The contact zone between the PRI and South Kawishiwi Intrusion (SKI) is poorly exposed and poorly drilled (Fig. 8-2). All of the units of the PRI marginal zone, including even the upper units, become unrecognizable in the contact zone with the SKI and correlation of igneous units in one drill hole to a nearby drill hole is tenuous at best. Severson (1994) reasoned that the heterogeneous contact zone of the PRI originally formed at the contact with pre-Keweenawan footwall rocks, and that later emplacement of the SKI effectively removed the footwall portion and positioned SKI intrusive rocks up against PRI intrusive rocks. Complicating the picture of the contact zone is the coincident Grano fault (Fig. 8-16), which appears to have been repeatedly reactivated during, and after, emplacement of the PRI and SKI. The Grano fault may have also served as a feeder channel to the massive sulfides of the Local Boy ore zone at the Babbitt deposit (Fig. 8-15).

More petrologic studies have been conducted on the PRI than any other intrusion in the Duluth Complex. These studies pertain to drill holes in the Babbitt deposit (see Miller and others, 2002 for references). Most of the studies pertain to a stratigraphic section intersected in one or a few drill holes that represent only a small fraction of the PRI magmatic system. Divergent progenetic interpretations have resulted from these studies and have been summarized by Miller and Ripley (1996).

South Kawishiwi Intrusion

The South Kawishiwi intrusion (SKI) of Green and others (1966) hosts at least five sub-economic Cu-Ni deposits and a potential PGE-Cu-Ni deposit (Fig. 8-2). The SKI is dominantly composed of troctolitic cumulates that are exposed in an 8- x 32-kilometer arcuate band. Footwall rocks include the Virginia Formation, in the Serpentine and Dunka Pit deposits, the Biwabik Iron Formation in the Dunka Pit and Birch Lake deposits, and the Archean Giants Range Batholith in the Dunka Pit deposit north to the Spruce Road deposit. The presence of Biwabik Iron Formation as inclusions as far north as the Spruce Road deposit indicates that the majority of Paleoproterozoic units were assimilated and removed from the footwall during emplacement of the SKI leaving the Giants Range Batholith as the dominant footwall rock type. Also present as inclusions in the Dunka Pit and Serpentine deposits, are mafic volcanic hornfels (probable North Shore Volcanic Group) and quartz sandstone hornfels (probably either the Puckwunge or Nopeming sandstones. Anorthositic rocks (unit asau Fig. 8-2) abut the SKI on the northeast and enclose a possible SKI feeder dike that extends farther northeast. To the east, the SKI is inferred to be in semi-conformable contact with the Bald Eagle Intrusion (unit bei; Fig. 8-2). However, based on their relative cross cutting relationships to the Greenwood Lake intrusion, it is clear that the Bald Eagle is younger than the SKI (Chandler, 1990; Miller and others, 2002).

On the regional Duluth Complex map (M-119, Miller and others, 2001), the SKI is subdivided into five major map units. These are, from the base upward,

1. a basal contact zone that is a heterogeneous mix of sulfide-bearing troctolitic, gabbroic, and noritic rocks with abundant hornfels inclusions (unit skcz, M-119);
2. a thick unit of subophitic to ophitic augite troctolite (unit skat, M-119) that contains an internal ophitic olivine gabbro unit (unit skog, M-119) ;

3. discontinuous and localized layers of poikilitic leucotroctolite (unit skpt, M-119);
4. a thick homogeneous sequence of ophitic troctolite (unit sktr, M-119) ; and
5. an uppermost thick sequence of homogeneous troctolite (unit skta, M-119) that contains numerous anorthositic layers (designated as unit skan where anorthositic layers are mapped in detail).

Detailed mapping of the Babbitt NE 7.5' quadrangle in 2001 and 2002 and published as an MGS open-file map (Miller, Severson and Foose, 2002), has subdivided the southern part of the SKI into eight stratiform map units. Most of these units show considerable variation in thickness and some pinch out altogether. Moreover, whereas Foose and Cooper (1978) have interpreted most of the anorthositic rock occurrences in the upper part of the SKI to be conformable plagioclase cumulate layers interleaved within the largely troctolitic cumulates, we have interpreted most anorthositic rock occurrences to be inclusions.

Severson (1994) and Zanko and others (1994) further subdivide the marginal zone of the SKI (including units skcz, skat, skog, and sktr, Fig. 8-2) into 17 different lithostratigraphic units (Fig. 8-5) that are present in over 180 drill holes over a strike length of 31 kilometers. Sulfide mineralization is confined to the BH, BAN, UW, and U3 units, and to a lesser extent the U1 and U2 units. Major marker horizons that are correlated in drill hole include three horizons with abundant cyclic ultramafic layers (U1, U2, and U3 units) and a pegmatite-bearing unit (PEG Unit) that was initially recognized by Foose (1984). A large anorthositic inclusion (> 1 km thick) is intersected in six deep drill holes in the Highway One corridor area (AN-G Unit in Fig. 8-5).

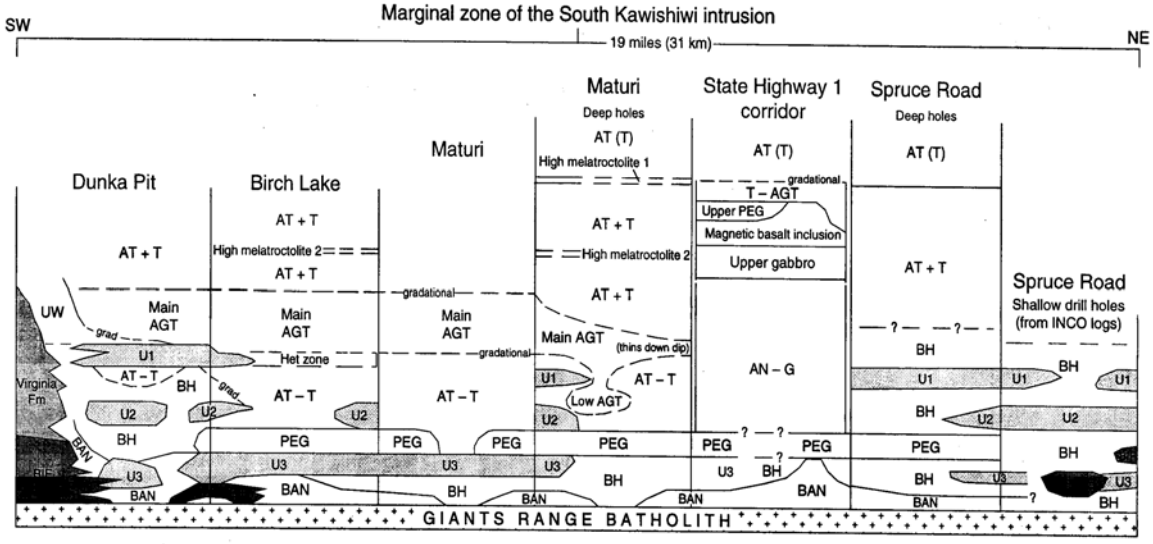


Figure 8-5. Generalized stratigraphy of the marginal zone of the South Kawishiwi intrusion (modified from Severson, 1994; taken from RI-58 figure 6.12).

The lowest units of the marginal zone (Fig. 8-5) are the most varied with respect to textures, rock types, and sulfide content. They are very unevenly distributed along the strike length of the SKI in a "compartmentalized" fashion, suggesting a complicated intrusive history. The lowest units were emplaced early into several restricted magma chambers via repeated and close-spaced magmatic pulses. The U1, U2, and U3 units (Fig. 8-5) represent periods of rapid and continuous magma injection that crystallized more primitive ultramafic layers before mixing with the resident magma. The U3 Unit is unique among the lower units in that it contains several massive

oxide pods (titanomagnetite-rich) along its entire length. A spatial correspondence between the U3 Unit and footwall iron-formation suggests that most of the massive oxide pods are iron-rich "restite" produced by the magmatic digestion of iron-formation. The U3 Unit also contains the majority of the high PGE values that have been sampled to date within the SKI. The upper units of the SKI are more laterally continuous throughout the intrusion, though recent mapping in the Babbitt NE quadrangle shows significant variation in thickness and, in some cases, pinching out of these upper troctolitic units. Still, the generally greater lateral continuity and monotonous troctolitic composition of these upper units suggest that they crystallized in a more quiescent and open magmatic system characterized by widely-spaced, large volume magmatic pulses. The occurrence of abundant anorthositic inclusions, especially in the uppermost unit (Fig 8-2), implies that much of the growth of the SKI chamber was accomplished by emplacement of new magma between previous troctolitic injections and an anorthositic series hanging wall.

Disseminated Cu-Ni Sulfide Mineralization in the PRI and SKI

Large resources of low-grade copper-nickel sulfide ore that locally contain anomalous PGE concentrations are well documented by drilling in the basal zones of the Partridge River and South Kawishiwi intrusions. At least nine subeconomic deposits (Fig. 8-3) have been delineated in the basal 100 to 300 meters of both intrusions. The mineralization consists predominantly of disseminated sulfides that collectively constitute over 4.4 billion tons of material averaging 0.66% Cu and 0.20% Ni (Listerud and Meineke, 1977). Overall, the copper to nickel ratio averages 3.3:1; however, there are wide variations in the Cu:Ni ratio from one Cu-Ni deposit to another deposit and also internally within each of the deposits. PGE concentrations average about 10 ppm Pt+Pd (recalculated to 100% sulfide), but may range as high as 50 ppm Pt+Pd (recalculated to 100% sulfide) in associated stratabound zones, such as at the Dunka Road and Birch Lake deposits.

The disseminated sulfide deposits are hosted by taxitic troctolitic to gabbroic rocks that contain abundant inclusions of footwall rock types. Within the Partridge River intrusion, the basal unit - Unit I of Severson and Hauck (1990; Fig. 8-4) hosts the vast majority of the disseminated sulfides. Similarly, mineralization within the South Kawishiwi intrusion is confined to the bottom-most units (Fig. 8-5) that include the following units: BH (basal heterogeneous); BAN (basal augite troctolite and norite); three ultramafic units (U1, U2, and U3); and UW (updip wedge). The disseminated sulfide minerals (dominantly pyrrhotite, chalcopyrite, cubanite, and pentlandite) occur as interstitial grains that make up between trace amounts and 10% of the rock by volume (visual estimation). Pyrrhotite is generally the dominant sulfide, especially closer to the basal contact.

Although this mineralization type is categorized as being present within the basal portions of the intrusions, it is important to stress that the rock units do not always contain sulfides throughout their entire vertical section. Mineralized zones are independent of rock type and are typically extremely erratic in their spatial extent and ore grades. Zones that are barren of sulfides commonly "interfinger" with mineralized zones in a random pattern. This erratic pattern of mineralization, in part, mirrors the lithologic heterogeneity of the basal units. The only exception to this random mineralization pattern is the Maturi deposit, and its downdip extension (Maturi Extension; Peterson, 2001), where the uppermost portions of the BH Unit generally exhibit copper values in excess of 1.0% that gradually decrease with depth toward the basal contact.

Contrasting with this heterogeneity of rock types and mineralization, some internal PGE-bearing sulfide zones within the lower units of both intrusions exhibit a stratabound relationship to the igneous stratigraphic section. Examples include Dunka Road and Birch Lake (see discussion below).

Basal Cu-Ni Massive Sulfide Mineralization

In a few localized areas along the basal zones of the SKI and PRI, semi-massive to massive sulfide mineralization is present at the basal contact. In most cases, the massive to semi-massive sulfide is proximal to either sulfide-rich footwall rocks or structures such as faults and pre-Complex folds. Massive sulfide zones that are spatially related to sulfide-rich footwall rocks are intersected in scattered drill holes in the Babbitt, Serpentine, and Dunka Pit deposits (Severson and others (1994), Zanko and others (1994), and Severson (1994), respectively). All of these massive sulfides are pyrrhotite-rich (with generally <2% Cu) and are present at, or slightly above, the basal contact. In all cases, a pyrrhotite-rich member of the footwall Virginia Formation (BDD PO unit) is located at the basal contact and is situated up-dip of the massive sulfide occurrences. This relationship suggests that the BDD PO unit acted as a local sulfur source, which generated a Cu-poor, sulfide-rich melt that was gravitationally concentrated downdip, along the basal contact.

The massive sulfide occurrence in the Local Boy ore zone of the Babbitt deposit (Fig. 8-3) is clearly structurally controlled. At this locality, the massive sulfide zones are Cu-rich (generally 5-25% Cu) and are situated along the axis of an anticline defined by the footwall rock units. The highest PGE values (11 ppm Pd and 8 ppm Pt) yet found within the Duluth Complex are associated with these structurally-controlled Cu-rich massive sulfides. The massive sulfides are almost exclusively hosted by the Virginia Formation, present as both inclusions above the basal contact and in the footwall rocks below the basal contact. These relationships, plus sulfide textures that are indicative of structural preparation, suggest that the massive sulfides were "injected" into the footwall rocks. Ripley (1986) and Severson and Barnes (1991) propose that an immiscible sulfide melt, formed in an auxiliary magma chamber at depth, was injected into structurally prepared zones in the footwall rocks along the anticline to form the Local Boy ores. Late movement of Cl-rich fluids, along the axis of the anticline, further redistributed and concentrated the PGEs (Severson and Barnes, 1991). Recent studies (Severson and Zanko, in prep.) indicate that there is an overall increase in the Cu-PGE content of the massive sulfide in an east-to-west direction (Fig. 8-6); this is perhaps the result of fractional crystallization of immiscible sulfide melt as it migrated into the footwall rocks. In this scenario, the north-south trending Grano fault is inferred to be a potential feeder zone.

Structurally-controlled veins and irregular pods of massive sulfide are locally present within granitic footwall rocks immediately beneath the SKI. These occurrences are intersected in scattered holes that outline two northeast-trending belts (Fig. 8-7). The linearity of the belts suggests they are fault controlled. One of these belts crudely aligns with the Birch Lake fault zone which trends through the Birch Lake PGE prospect. The veins are moderately Cu-enriched due to fractional crystallization of the sulfide melt as it moved down through the footwall rocks (Bonnichsen and others, 1980; Severson, 1994).

The occurrence of local massive sulfide veins near and below the basal contact of the Duluth Complex is an indication that larger, potentially economic footwall massive sulfide deposits may yet be found. In the Sudbury Complex, pooling of a monosulfide solid solution (mss) melt at the basal contact appears to be an important prerequisite to the injection of fractionated sulfide melts (Naldrett, 1997).

Stratabound PGE Mineralization

Most of the PGE-enriched zones in both the Partridge River and South Kawishiwi intrusions are stratabound in nature. These stratabound PGE horizons are intimately associated with the Cu-Ni sulfide mineralization and with one or more ultramafic layers that indicate magma recharge events. Thus, their PGE-enrichment appears to be related to magma mixing. However, there are also indications that the PGE content in some of these horizons was locally modified by later Cl-rich hydrothermal solutions.

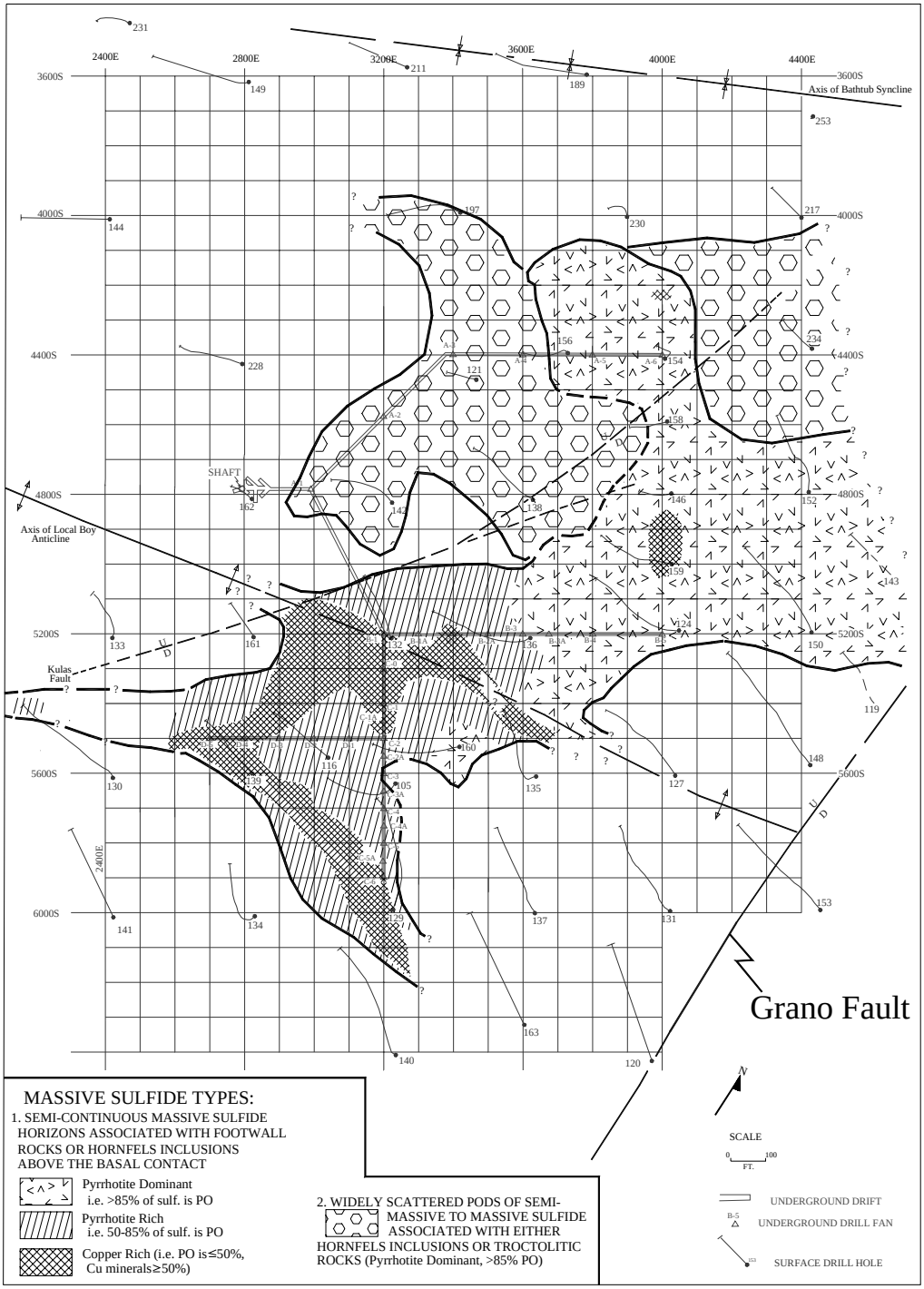


Figure 8-6. Potential distribution of semi-massive to massive sulfide types, relative to the Grano fault and Local Boy anticlinal axis, at the Local Boy ore zone of the Babbitt deposit (from Severson and Zanko, in prep.; taken from RI-58, Figure 8.4).

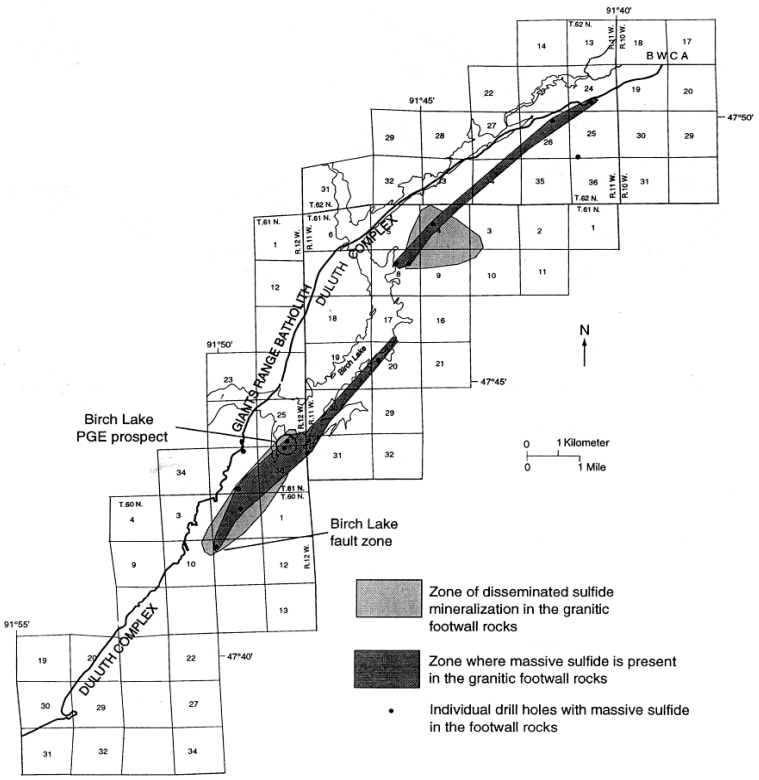


Figure 8-7. Linear distribution of massive sulfide, disseminated sulfide, and copper-rich veins in the Giants Range granitic footwall rocks beneath the South Kawishiwi intrusion (after Severson, 1994; taken from RI-58, Figure 8.5).

Stratabound PGE-enriched horizons, with low to moderate sulfide concentrations (0.05-1.0 wt.% S), are commonly associated with ultramafic layers in the Dunka Road, Babbitt, Wetlegs, and Birch Lake deposits (Fig. 8-3). Elevated Cu and PGE concentrations at the Dunka Road deposit occur at the extreme top of Unit I immediately beneath a laterally persistent ultramafic layer. This stratabound horizon (red horizon of Geerts, 1991, 1994) averages about 10 meters thick and contains an average of 1.0 ppm Pd+Pt. Recent work by Theriault and others (1997, 2000) suggests that the sulfur was largely derived from the mafic magma and that this stratabound horizon was formed as a result of magma mixing. Two similar stratabound PGE-enriched horizons, related to laterally discontinuous ultramafic layers, occur toward the middle of Unit I at Dunka Road (orange and yellow horizons of Geerts, 1991; 1994).

To the west of Dunka Road, the stratabound PGE horizon at the top of Unit I is also present at the Wetlegs and Wyman Creek deposits. There however, the overall Pd content in this horizon exhibits a definite decrease in an east-to-west direction suggesting that as the magma was intruded it became progressively impoverished with respect to PGE (Severson and Hauck, 1997; Theriault and others, 1997). Additional PGE-enriched stratabound horizons are situated much further above the basal contact in the Partridge River intrusion. These horizons are also often associated with ultramafic layers and are present at Dunka Road, Wetlegs and the Fish Lake area (Sassani, 1992; Severson, 1995).

Another example of a PGE stratabound horizon is at the Birch Lake PGE prospect within the SKI. There, PGE contents as high as 9 ppm Pd+Pt, and Cr_2O_3 contents locally as high as 10 wt.%, are associated with a wide variety of rock types within the U3 Unit. The U3 Unit consists of alternating troctolitic and ultramafic layers in which variably sulfide-mineralized zones and discontinuous pods of Cr-bearing massive oxide both occur (Severson, 1994). The massive oxide pods are interpreted, based on empirical relationships, to have been produced by assimilation and partial melting of the Biwabik Iron Formation. This oxide-rich partial melt may have initially acted as a trap that concentrated Cr and Ti, and through further assimilation and contamination of

the magma, may have led to precipitation of PGE. However, because the ultramafic layers of the U3 Unit are interpreted to record new influxes of more primitive magma (Severson, 1994), magma mixing may have played a more significant role in PGE mineralization.

Furthermore, the presence of Cl-rich drops on the surface of drill core from the Birch Lake area suggests that a hydrothermal model of concentrating the PGE could also be invoked. A model involving ascending Cl-rich hydrothermal fluids, depicted in Figure 8-8 and similar to a model proposed by Boudreau and McCallum (1992), may have remobilized and further concentrated the PGE along the northeast-trending Birch Lake fault (Severson, 1994). A possible explanation, depicted in Figure 8-8, is that the Birch Lake area represents an area where there was a local increase in the amount of upward-moving Cl-rich solutions that were concentrated, or funneled, along the Birch Lake fault.

Another explanation is that the Birch Lake Fault may have initially served as a subsidiary feeder zone to the South Kawishiwi intrusion (Hauck and others, in prep.). According to their model, uncontaminated PGE-enriched magmas were vented in the Birch Lake area. Upon mixing with the resident magma, which was contaminated due to interaction with the Biwabik Iron Formation, PGEs were deposited in zones of turbulent mixing. As the magma migrated away from the vent area (up-dip?) it became progressively impoverished with respect to PGEs.

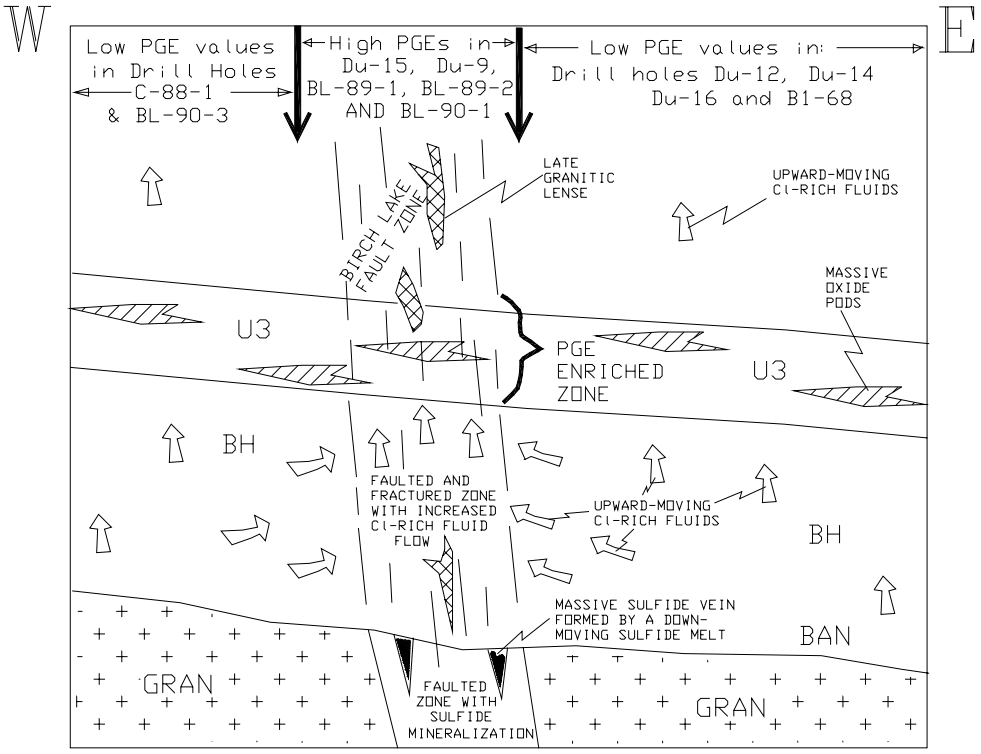


Figure 8-8. Schematic diagram showing the possible role that the Birch Lake fault may have played in funneling upward-moving Cl-rich solutions and the resultant reconcentration of significant magmatic PGE within the U3 Unit at the Birch Lake PGE area (modified from Severson, 1994; taken from RI-58, Figure 8.8).

Oxide Ultramafic Intrusions (OUI)

Exploratory drilling for Cu-Ni mineralization encountered several OUIs, that later were evaluated for their Fe-Ti±V potential. Many of the OUIs are expressed as aeromagnetic highs, commonly with an associated electromagnetic conductor, and thus they were initially drilled in search of conductive sulfide mineralization. At least thirteen OUIs have been intersected in drill holes along the basal contact of the Duluth Complex (Fig. 8-3).

The OUIs are plugs or pipe-like bodies that commonly have irregular apophyses. They intrude troctolitic rocks of the Partridge River, Western Margin, and Boulder Lake intrusions. The Waterhen OUI appears to be rootless as defined by detailed drilling; the three-dimensional configurations of the other OUIs are unknown due to insufficient drilling. In general, the OUIs are spatially arranged along linear trends suggesting that structural control was important to their genesis. Almost all of the OUIs are cross-cutting. Rock types include coarse-grained to pegmatitic clinopyroxenite, dunite, peridotite, melatroctolite, and minor melagabbro; all rock types are oxide-bearing. Some OUIs exhibit a crude zonation from an olivine-rich core (dunite, peridotite, melatroctolite) to an outer clinopyroxenite margin, whereas others consist of only one dominant rock type.

Oxide content is variable in the OUIs and ranges from 15% in disseminated zones to 100% in localized massive oxide zones. OUIs that contain thick intervals of massive oxide include Longnose (up to 30 meters thick; Linscheid, 1991) and Section 34 (up to 40 m thick; Severson, 1995). Titanomagnetite is dominant in some of the OUIs whereas ilmenite is dominant in other OUIs. Sulfide minerals (predominantly pyrrhotite) are ubiquitous in all the OUIs and range from trace amounts to 5% in disseminated zones to >70% in localized net-textured and massive sulfide zones, as at Waterhen, Boulder Lake South, and Fish Lake (Fig. 8-3).

STOP DESCRIPTIONS

STOP 8-1: Hornfels Virginia Formation, Duluth Complex Footwall at Linwood Lake

Location: Approximately 300 feet southwest of the intersection of Linwood Lake Road and Reagan Road. Harris Lake 7.5' quadrangle, T.56N., R.14W., Sec 22, NE of SE of SW; 567865E, 52407780N (NAD83)

Description: Outcrops of the Virginia Formation at Linwood Lake show a progressive increase in the amount of deformation, metamorphism, and degree of partial melting towards the basal contact of the Duluth Complex (which based on limited drilling in this area exhibits a near-vertical cliff-like geometry). In this area, outcrops furthest from the contact consist of interbedded argillite and fine-grained distal graywacke (Bouma sequence B with minor C) that exhibit shallow dips of 15° to the southeast. However, slightly closer to the Complex, bedding dips increase to 40-60° to the northeast and small wisps (<5 mm) of partial melt are present along bedding planes. Still closer to the Complex (Stop 8-1) the grade of metamorphism and associated deformation progressively increase and at least two metamorphic varieties are superimposed on the original sedimentary package. Both of these metamorphic varieties are visible at this stop and consist of the following:

Disrupted unit (Fig. 8-9)– In close proximity to the Complex the well-bedded sediments of the Virginia Formation are typically transformed into a highly deformed rock or metatexite (Sawyer, 1999). Textures that characterize this rock are bedding planes that are extremely chaotic and random in orientation due to small-scale folding, faulting, and brecciation. Superimposed on this chaotic pattern are abundant zones of partial melt that are also chaotic and folded. Albeit differences in grain size, the partial melts are composed of the same mineralogy as the surrounding sedimentary rocks (feldspar, quartz, orthopyroxene, cordierite, and biotite).



Figure 8-9. Photograph of disrupted unit of metamorphosed and folded Virginia Formation with abundant partial melt lenses and wisps in close proximity to the Duluth Complex at Linwood Lake, MN.

Recrystallized unit (Fig. 8-10)– This unit is a higher-grade metamorphic equivalent of the disrupted unit, and is properly classed as a diatexite (Sawyer, 1999). By contrast, rocks of the recrystallized unit were heated, generating 20-40% pervasive partial melts, which literally enabled the rocks to flow in response to stresses that were applied during emplacement of the Duluth Complex. All bedding planes are obliterated and what remains is a medium-grained recrystallized rock that contains plagioclase, cordierite, orthopyroxene, and decussate biotite. Within this recrystallized matrix are blocks/boudins of more structurally competent siltstone and calc-silicate hornfels (originally limey layers).



Figure 8-10: Photograph of recrystallized unit of metamorphosed Virginia Formation in close proximity to the Duluth Complex at Linwood Lake, MN.

STOP 8-2: <u>Anorthositic Series, Duluth Complex, Skibo Vista</u>

Location: Roadcut on St. Louis Co. Highway 110, approx. 3.7 mi. north of Co. Hwy 16.
Bird Lake 7.5' quadrangle; T.57N., R.13W., Sec 17, SE of NW; 574140E 5252806N (NAD83)

Description: Exposed in roadcuts on either side of Co. 110 is an exposure of poikilitic troctolitic anorthosite that is a common lithology of the anorthositic series of the Duluth Complex. The rock varies in texture and modal mineralogy on a meter scale over the outcrops ranging from medium- to medium coarse in grain size, from subpoikilitic to poikilitic in olivine habit, and from 80 to 90% in plagioclase mode. Plagioclase foliation is locally well developed and varies in orientation throughout the exposures. Olivine oikocrysts range from 1 to 5 cm in diameter (larger ones on north side of exposures) and are locally concentrated in planes that parallel foliation and thereby impart a subtle layering.

This exposure lies at the western extent of a large expanse of plagioclase-rich gabbroic rocks that comprise the Anorthositic Series of the Duluth Complex (Fig. 8-1). West of here, troctolitic cumulates of the Western Margin intrusion are inferred to occur (Figs. 8-1, 8-2). The lithologic characteristics and structural complexities of the Anorthositic Series, which are hinted at here,

have lead many workers (Grout, 1918; Taylor, 1964; Miller and Weiblen, 1990) to conclude that these rocks formed from multiple injections of plagioclase-enriched mafic magmas (plagioclase crystal mushes) early in the main stage of Duluth Complex magmatism. The effectively identical ages of Anorthositic and Layered Series rocks (~1099Ma, Paces and Miller, 1993) implies that intrusion of normal (crystal-free) magmas, forming layered series intrusions like the Western Margin Intrusion, followed soon after the emplacement and crystallization of the Anorthositic Series.

STOP 8-3a: Erie hornfels – mafic volcanic unit overlying the Virginia Formation, Erie/LTV RR Grade

Location: Access via Dunka Road (private mining company road) approximately 5 miles west the LTV guard shack. Allen 7.5' quadrangle, T.59N., R.13W., Sec 18, NW of SE of SW; 571815E 5271060N (NAD83)

Description: Inclusions of hornfelsed basalt bodies are documented in many places in the Duluth Complex – the earliest detailed studies were by Kilburg (1972), for the Ely's Peak basalts beneath the Complex near Duluth, MN, and Bonnichsen (1972), for several inclusions located within the Complex. The Erie hornfels was originally discussed by Bonnichsen (1972), and was more recently described in detail by Tyson (1976). Detailed geologic mapping of the area (Severson and Miller, 1999) indicates that the Erie hornfels directly overlies the Virginia Formation (based on the geology in surrounding drill holes) and is most likely situated near the base of the North Shore Volcanic Group in this area.

The Erie hornfels is exposed in a 215-meter long railroad cut. Tyson (1976) described it as a series of basalt flows, dipping 55° to the southeast, that are characterized by fine-grained granoblastic rocks containing variable amounts of plagioclase, olivine, and augite with lesser amounts of hornblende and biotite. Relict amygdules (plagioclase- or augite-filled ovoids) and relict plagioclase phenocrysts are present in many places in the railroad cut but the amygdules appear to be more concentrated in the upper portions of the flows. Tyson (1976) also reports that biotite-rich layers are locally present in the exposures and that these layers were probably formed from either interflow sedimentary rocks or altered flow tops.

STOP 8-3b: Marginal Units of the Partridge River Intrusion, Erie/LTV RR Grade

Location: Erie/LTV RR Grade approximately 1/3 mile east of previous stop. Allen 7.5' quadrangle, T.59N., R.13W., Sec 18, SE of SE, and Sec 19, SW of SW; traverse from 572430E, 5271135N to 572960E, 5271205N (NAD83)

Description: The Wetlegs Cu-Ni deposit was initially drilled by Bear Creek Mining in 1958-1960 (referred to as the A4 grid) on the basis of electromagnetic conductors which were found to be related to graphitic portions of the footwall Virginia Formation. Some indications of Cu-Ni mineralization were found in the initial drilling and after acquiring state mineral leases in 1966 Bear Creek conducted some follow-up drilling. However, the Cu-Ni grades were low overall, and Bear Creek abandoned further mineral exploration endeavors in this area of the Complex. Exxon Minerals Corp. leased the same lands and conducted drilling programs in 1975 and 1979. They estimated that the Wetlegs deposit contained 38 million tons of material grading 0.29% Cu and 0.10% Ni using a cutoff grade of 0.15% Cu. Recent sampling for platinum group elements (PGE) indicates that there are at least three potential stratabound zones at Wetlegs (Severson and Hauck, 2003).

A west to east traverse along the LTV railroad tracks will be conducted at this field stop. The traverse, approximately 0.5 miles long, will start at the basal contact and progress upwards through the igneous stratigraphy of the Partridge River intrusion. Outcrops of Units I and III will be viewed. A generalized geologic map of the Wetlegs area is shown in Figure 8-11. Detailed descriptions of the railroad cuts are listed below.

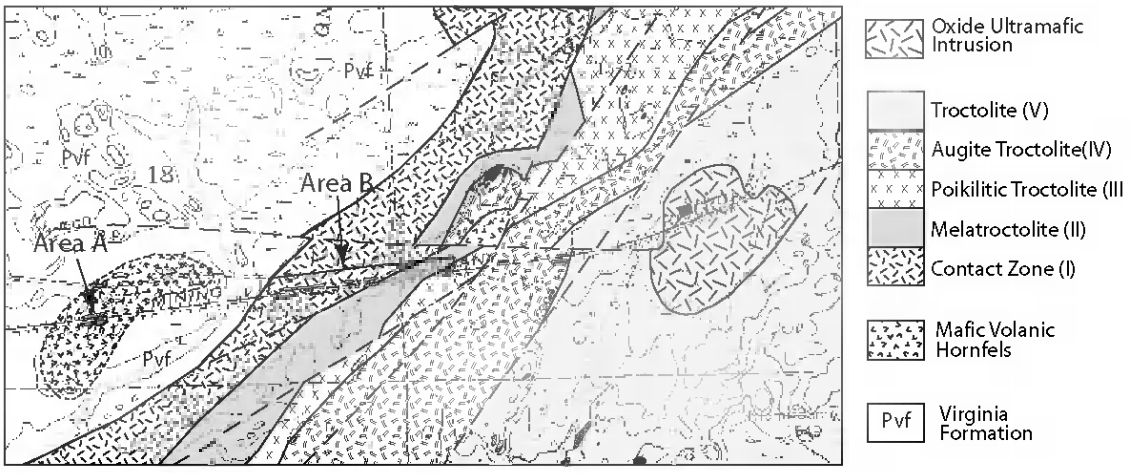


Figure 8-11: Geology of the Partridge River Intrusion in the vicinity of the Wetlegs Cu-Ni deposit showing stops 8-3a (Erie hornfels) and 8-3b (Wetlegs deposit). Roman numerals correspond to stratigraphic units of Severson and Hauck (1990). Modified from Severson and Miller (1999).

Feet (to the East)	**Description**
0 ft.	Railroad culvert – Longnose Creek
100-110 ft. (left side of tracks)	Sulfide/gossan-cemented till consisting mostly of large boulders of a pyrrhotite- and graphite-rich member of the Virginia Formation (the BDD PO unit). Note the rounded granite cobbles beneath the BDD PO boulders. Drilling indicates that the BDD PO subcrops very close to this location.
337-355 ft. (left side of tracks)	Very fine-grained (chilled) gabbronorite at basal contact with 60% plagioclase, 30% orthopyroxene, 5 % clinopyroxene, trace to 2 % biotite, and 3 % oxides. The outcrop contains well-assimilated "streaks" of Virginia Formation inclusions (very hard to see).
400-1025 ft. (both sides of tracks)	Unit I – taxitic medium- to coarse-grained ophitic augite troctolite (POcf) to olivine gabbro (PcOf) with patches and lenses of augite-rich pegmatite. The outcrops are sulfide bearing with the sulfides unevenly distributed along patches, spots, lenses, and joint faces (at numerous orientations!). Trace amounts to 5% uralite due to pervasive deuteric alteration is present.
1245-1355 ft. (two outcrops on left side of tracks)	Unit I – taxitic, pervasively uralitized, medium- to coarse-grained ophitic olivine gabbro (PcOf) to augite troctolite (POcf) with trace to 1% sulfides. Also present within these outcrops are very coarse-grained anorthosite inclusions that vary from 10 cm across to 3x5 meter blocks. The edges of the inclusions are sharp, and straight to highly lobate. Within the inclusions, the plagioclase foliation is subvertical and highly variable.
1385-1685 ft. (many outcrops on	Unit I – taxitic, medium- to coarse-grained ophitic augitic troctolite (POcf) with abundant irregular pegmatitic patches and lenses. The pegmatites

both sides of tracks)	contain variable amounts of saussurized plagioclase, clinopyroxene, oxides, biotite, and uralite, with minor quartz, k-feldspar, graphic granite, and sulfides. Most of the sulfides are either within or adjacent to the pegmatites. Uralite is common throughout the outcrops as pervasive replacement products in irregular patches and along joints.
2355-2480 ft. (four outcrops on both sides of the tracks)	Unit III – mottled, poikilitic/ophitic troctolite (Po(cf)) to augite troctolite (Pocf). This is a major marker bed due to the presence of medium- to high-density olivine oikocrysts up to 10 cm across. This unit is easily recognized in drill core due to the mottled texture, and the relatively finer-grained plagioclase (1-5 mm).

Dunka Road and Babbitt Deposits

As the bus proceeds to the next field stop, we will traverse across the southernmost limits of the Dunka Road deposit (originally drilled by United States Steel, now held by PolyMet Mining and called the NorthMet deposit) and the Babbitt deposit (originally drilled by Bear Creek and Amax, now held by Teck Cominco and called the Mesaba deposit). There is very little to see in the way of mineralized exposures for either of these deposits. However, one point to keep in mind is the immense size of these low-grade deposits. The Dunka Road/NorthMet deposit extends for about 3 miles along the road, and the Babbitt/Mesaba deposit extends for another 2.5 miles along the road (the deposits are separated by undrilled and untested area that is about 1 mile wide).

STOP 8-4: Hornfels mafic volcanic inclusion and oxide ultramafic intrusions near the margin of the Partridge River Intrusion

Location: Erie/LTV RR Grade approximately 2 miles west of the locked gate on the Dunka Road. Babbitt NE 7.5' quadrangle T. 60N., R.13W., Sec 33.; 584510E, 5275715N (NAD83)

Description: The Dunka Railroad hornfels is a large mafic volcanic inclusion located near the eastern margin of the Partridge River intrusion and just south of the Babbitt/Mesaba deposit. Outcrops and drill hole information indicates that the inclusion is about 900 x 1,500 meters across. The hornfels is a fine-grained, granoblastic to poikiloblastic basaltic hornfels that contains variable amounts of plagioclase, augite, hypersthene, inverted pigeonite, and olivine. Both massive and meta-amydaloidal varieties are present.

In addition to the basaltic hornfels, several small bodies of late intrusive OUIs are also present in the railroad cuts. The abundance of OUIs at this locality is related to proximity to the north-trending Grano Fault, which may have served as a feeder vent for the massive sulfides at the Local Boy area of the Babbitt/Mesaba deposit. These OUIs, and a wide variety of granitic rocks, occur as lenses and bodies that cut the troctolitic rocks of the Partridge River intrusion. They are extremely common in drill holes, within a 400-580 meter wide zone, on the west side of the Grano Fault. The OUIs at this locality are characterized by medium- to coarse-grained clinopyroxenite with 75-80% augite, 0-15% olivine, <5% interstitial plagioclase, and 5-15% oxides (ilmenite is dominant). Contact relationships with the basaltic hornfels are sharp but highly irregular and lobate. Granitic dikes and veins, present within a NNE-trending zone, are also evident in the exposures. They are rarely observed in close proximity to the OUI, but where they are, the dikes crosscut the OUI.

STOP 8-5: <u>Cu-Ni sulfide mineralization, Spruce Road deposit, South Kawishiwi Intrusion</u>

Location: Barrow pit 200 feet west of Spruce Road (Forest Rd 181) approximately 3.5 miles north of Highway One. Bogberry Lake 7.5' quadrangle, T.62N., R.11W., Sec 24, NE of SE of SW; 599295E, 5299010N (NAD83)

Description: It all started here! Serious exploration for Cu-Ni deposits at the base of the Duluth Complex began somewhere near this site in 1948 when strongly mineralized and gossanous rocks were uncovered in an excavation into weathered gabbro rubble used to build the Spruce Road. Local prospector Fred S. Childers of Ely, MN, noted copper stains in the material and began searching the outcrops along the basal contact in the vicinity of the Kawishiwi River. Roger V. Whiteside of Duluth, MN, later joined him in further exploration. In 1951, they diamond drilled a 57 meter (188 feet) deep hole located several miles to the southwest of this site in what is now referred to as the Maturi deposit (Fig. 8-3). This hole intersected disseminated sulfides in gabbroic rock that averaged 0.36% Cu and 0.13% Ni (Watowich and others, 1981).

In 1952, both Bear Creek Mining Company and the International Nickel Company (INCO) began intensive exploration efforts along a 61 km-long zone (38 miles) that coincided with the basal contact (an area stretching from south of the town of Hoyt Lakes northeastward to this site at Spruce Road). INCO eventually picked up the Childers-Whiteside properties and began drilling activities in 1954; whereas, Bear Creek concentrated most of their effort near the town of Babbitt and drilled several properties during 1957-1960. During this same period, the Minnesota Geological Survey, in cooperation with the U.S. Bureau of Mines, began an examination of the Maturi-Spruce Road area that culminated in the drilling of three holes in 1953. All of these exploration efforts indicated that large tonnage, but low grade, disseminated Cu-Ni deposits were present along the basal contact.

In 1966, the Minnesota Department of Conservation adopted rules for state mineral leases. The leases were offered through the Department of Natural Resources (DNR) and were awarded to successful bidders. Since 1966, over 20 companies have been actively involved in exploration for Cu-Ni and Fe-Ti-V deposits along the basal contact of the Complex and over 1.700 holes totaling over 1.5 million feet of core have been drilled.

In regards to the Spruce Road deposit, INCO drilled over 150 holes on the property at approximately 200-foot spacings. They calculated that the deposit contains 248 million tons of low-grade material averaging 0.46% Cu and 0.17% Ni. INCO collected at least two bulk samples from Spruce Road: 1) a 1,150 ton sample from several unknown small pits in 1966-67 (Watowich and others, 1981) and; 2) a 10,000 ton sample in 1974. It is uncertain if this barrow pit is one of the 1966-67 INCO bulk sample sites or if this pit was only used for road material. Rock types at this pit consist of sulfide-bearing, taxitic augite troctolite and olivine gabbro of the BH Unit (Basal Heterogeneous Unit) in the Marginal Zone of the South Kawishiwi intrusion. The BH Unit is approximately 360 meters thick (1,200 feet) at Spruce Road. Footwall rocks at Spruce Road consist dominantly of granitic rocks of the Archean Giants Range batholith; but in local areas the Biwabik Iron Formation is still preserved beneath the basal contact. At this locale, the pit is situated about 150 meters (480 feet) above the basal contact. Cu-Ni grades in the top of a nearby drill hole (34801) are 0.51-0.63% Cu and 0.15-0.17% Ni.

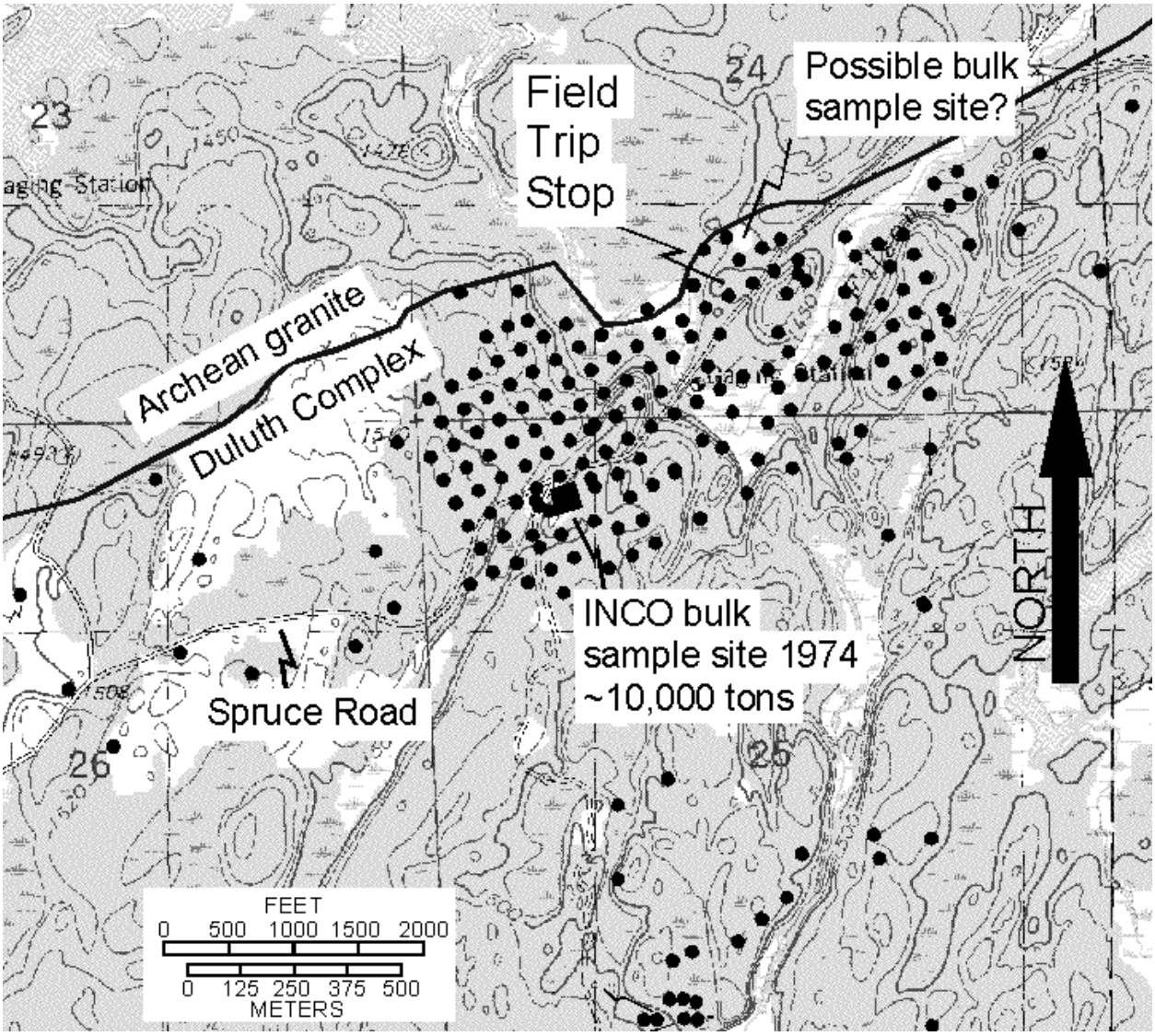


Figure 8-12: Location map of Stop 8-5 at the Spruce Road Cu-Ni deposit. Two possible INCO bulk sample sites are shown (this stop included), as well as, the reclaimed location of a 10,000 ton bulk sample collected by INCO in 1974.

STOP 8-6: <u>Quarry exposure of homogeneous troctolite, South Kawishiwi Intrusion</u>

Location: Approx. 3.0 miles south of Spruce Road on Highway 1; follow old access road to abandoned quarry about 800m S of Highway 1.

Bogberry Lake 7.5' quad; T.61N., R.11W., Sec 11, SE of NW; 597940E 5293550W (NAD83)

Description: This abandoned quarry provides a 3-dimensional look at the remarkable homogeneity of troctolitic cumulates, which compose most the South Kawishiwi (and Partridge River) intrusion. The average rock type is a medium-grained, moderately foliated, augite-poor troctolite. Moderately aligned, cumulus plagioclase makes up 70-75% of the rock, though locally up to 80% plagioclase occurs and the rock would be classified as a leucotroctolite. Subhedral granular olivine composes 20-25% of the average rock, though here too, local olivine enrichment

of up to 30% occurs in thin layers parallel to foliation. Interstitial augite and Fe-Ti oxide compose 3-7% of the rock.

The textural and mineralogic homogeneity of the troctolitic cumulates composing the middle to upper sections of the SKI (and PRI), and displayed here, are matched by limited variation in mineral compositions of plagioclase, olivine and pyroxene. Phinney (1969) reported a range of plagioclase and olivine compositions in troctolitic units of SKI (excluding the marginal rocks) of An_{57-72} and Fo_{50-63}, respectively. Similar results are displayed by unpublished microprobe data from troctolitic rocks from the Babbitt NE quadrangle, which indicate olivine compositions of Fo_{50-65} and augite compositions of En_{68-75}. Many see this lithologic homogeneity and lack of significant cryptic variation as indicating that the SKI and PRI magma systems were open to frequent recharge of relatively primitive magma (Severson, 1994; Lee and Ripley, 1996; Miller and Ripley, 1996). New recharge events are commonly marked by the intermittent occurrences of olivine-rich (melatroctolite) intervals separating otherwise subtle differences in troctolite texture and mode (Severson, 1994). These melatroctolite intervals show a significant increase in An, Fo, and En compositions consistent with the recharge of a more primitive magma. An alternative explanation has been put forth by Chalokwu (Chalokwu and Grant, 1990; Chalokwu and others, 1993). He has suggested that this homogeneity is indicative of single stage emplacement of a viscous semi-crystallized olivine-plagioclase mush that experienced little *in situ* differentiation. However, his sampling routinely overlooked the melatroctolite intervals.

STOP 8-7: <u>Anorthositic inclusion? in troctolite, South Kawishiwi Intrusion</u>

Location: Road cut on Tomahawk Rd. about 7.5 miles west of Highway 1;
Babbitt NE 7.5' quad; T.60N., R.11W., Sec 18, center of NE; 592355E 5282430W (NAD83)

Description: Scattered throughout the entire sequence of monotonous troctolitic cumulates are masses of coarse- to medium-grained troctolitic anorthosite to olivine gabbroic anorthosite. In recent mapping of the Babbitt NE quadrangle (Miller, Severson and Foose, 2002), we (Miller and Severson) have consistently interpreted these anorthositic masses as inclusions. We believe that they represent pieces of the anorthositic series roof pendant that was upwardly displaced and dissaggregated with successive recharge events into the SKI magma chamber. In an earlier detailed field mapping of this mixed anorthosite-troctolite terrane, mainly to the northeast of here, Foose and Cooper (1978, 1981) interpreted many of these anorthositic bodies as comagmatic layers within the troctolite. They saw their elongate and conformable relationship with the internal structure of the enclosing troctolite as evidence that most of these bodies are lensoidal plagioclase cumulate intervals intermittently layered within the troctolite. They used the apparent offset of these layers as evidence for a complex fault pattern (Fig. 8-13).

This stop along the Tomahawk road will examine one such example of a troctolitic anorthosite body within leucotroctolitic cumulates. This interval of the SKI is particularly rich in anorthosite bodies. The relationships displayed here are difficult to reconcile with either the Miller and others (2002) or the Foose and Cooper (1978,1981) interpretations. An irregularly shaped anorthosite block projects into the leucotroctolite and appears to cut across the foliation in the enclosing leucotroctolite (Fig. 8-13).

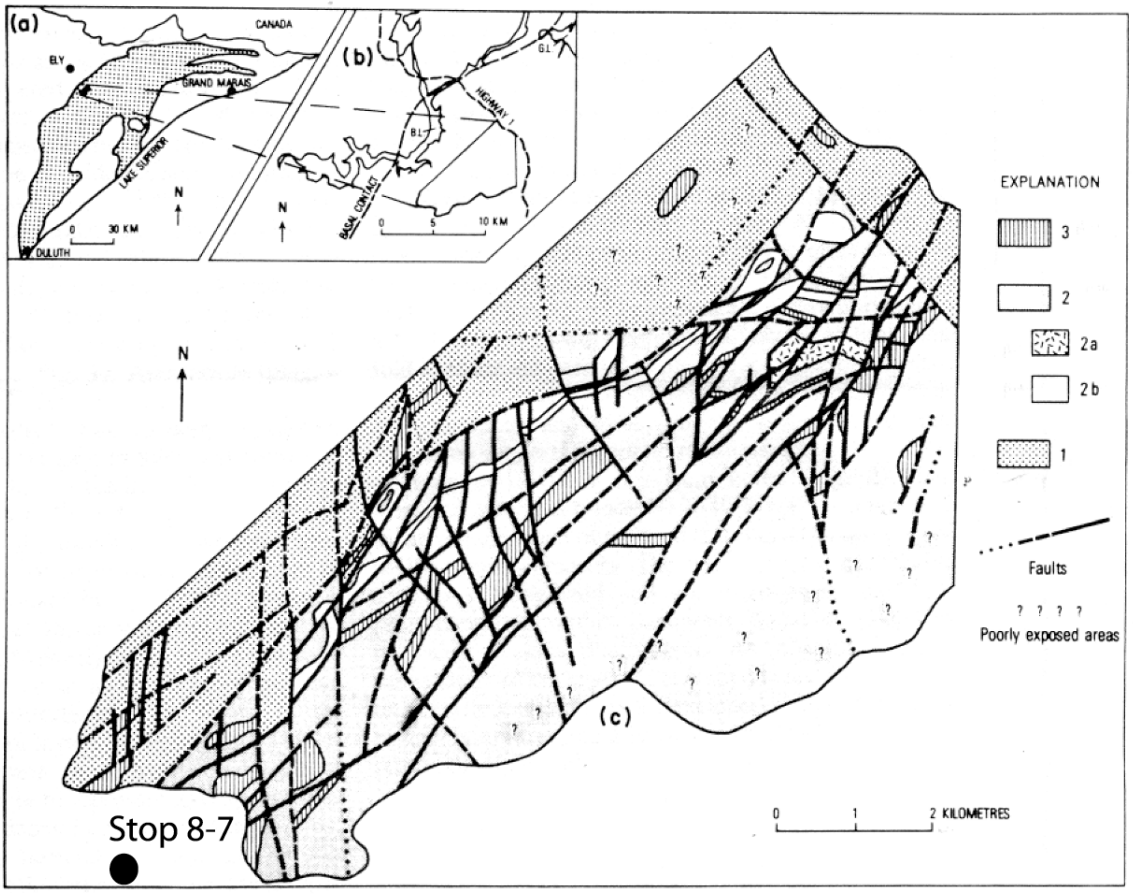


Figure 8-13. Geologic map of the Harris Lake area, which is northeast of Stop 8-7 from by Foose and Cooper (1981). Map units are: 1 - augite troctolite, 2 - troctolite, 2a - ophitic augite troctolite layer, 2b - traceable anorthosite layer, 3 - anorthosite



Figure 8-14. Photo of road cut at Stop 8-7 showing an irregularly shaped troctolitic anorthosite mass in leucotroctolite. Foliation in the leucotroctolite is approximately parallel to the hammer.

STOP 8-8: <u>Grano fault zone, Partridge River Intrusion</u> (optional)

Location: Small exposure north of Forest Rd 113, 100m east ;
Babbitt SE 7.5' quad; T.59N., R.12W., Sec 16, NW of NW; 584486E 5272519W (NAD83)

Description: The Grano fault is a major N-S trending structure that was first recognized in the subsurface by compiling exploration drill hole data related to the Babbitt and Serpentine deposit areas (Severson, 1994). As shown in Fig. 8-15, the fault creates a significant (~60m) displacement (down to the east) of the Duluth Complex footwall and affects both the Partridge River and the South Kawishiwi Intrusions. Drill holes into the fault zone in the southeastern portion of the Babbitt deposit shows that to the immediate west of the fault is a 400-580 meter-wide (up to 2,000 feet) zone that contains abundant subvertical lenses of granophyre and OUI. To the north along the fault trace the zone with late subvertical lenses pinches down appreciably (not present at Serpentine) and the amount of motion along the fault diminishes as well, indicating that the fault is a "scissors-type" fault. The massive sulfides of the Local Boy ore zone of the Babbitt deposit have been inferred to have been "vented" from the Grano fault.

The surface expression of the fault is a prominent linear valley that can be traced over 10 kilometers south from the basal contact (Fig. 8-16). This stop is in the axis of that valley. The rock types exposed on the margins of this valley are a complex mixture of augite leucotroctolite with poikilitic olivine, ophitic augite troctolite to olivine gabbro, poikilitic olivine gabbroic anorthosite, and locally granophyric gabbroic pegmatite. This assemblage represents a complex mix of anorthositic series inclusions and unmineralized marginal zone rocks of the Partridge River Intrusion. The granophyric gabbroic pegmatite is evidently related to a late, volatile-rich mafic magma intruded into Grano fault zone. In the small pavement exposure near the road, a medium-grained poikilitic augite leucotroctolite is cut by gabbroic pegmatite.

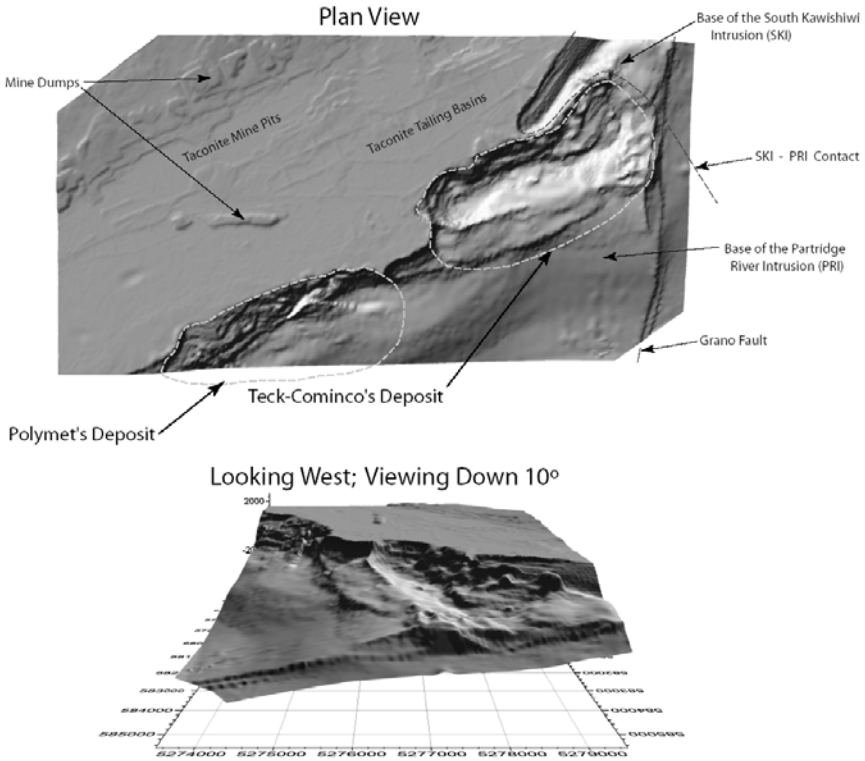


Figure 8-15. Grayscale image of the footwall topography of the Partridge River and South Kawishiwi intrusions near their contact showing the subsurface expression of the Grano Fault (image generated by Dean Peterson)

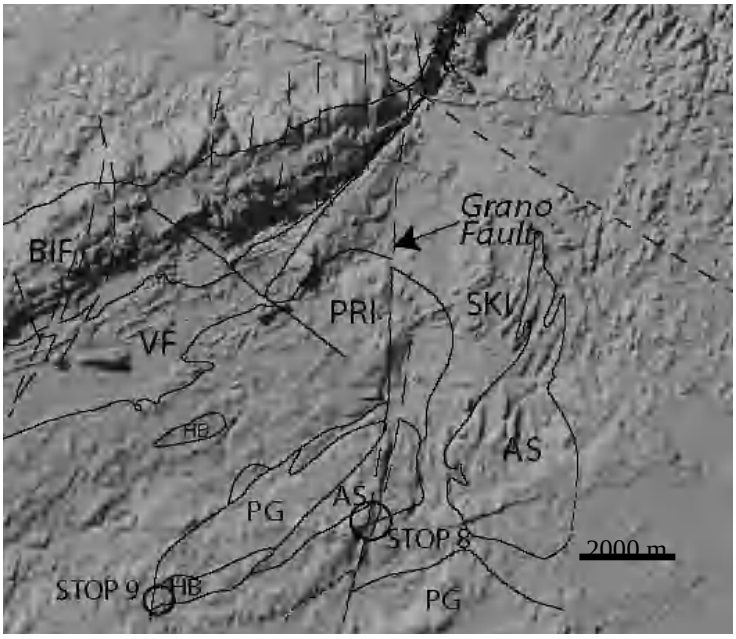


Figure 8-16. Grayscale image of surface topography of the Partridge River and South Kawishiwi intrusions near their contact showing the surface expression of the Grano Fault. Lines show geologic contacts and faults taken from MGS Misc. Map M-119. SKI - South Kawishiwi intrusion; PRI - Partridge River intrusion; AS - anorthositic series; PG - "Powerline Gabbro" of Bonnichsen (1974); HB - hornfels basalt; BIF - Biwabik Iron-formation; VF - Virginia Formation.

STOP 8-9: <u>Volcanic hornfels and plagioclase-phyric leucogabbro, Partridge River Intrusion</u>

Location: Roadcuts along Erie/LTV RR grade approx 1km west of crossing by Forest Rd 113 Babbitt SW 7.5' quad; T.59N., R.12W., Sec 18, SW of SW; 581500E 5271020W (NAD83)

Description: Scattered throughout the homogeneous troctolites of the Partridge River intrusion are kilometer-scale inclusions that are composed of various mixtures of anorthositic rocks, volcanic hornfels and oxide gabbro (Fig. 8-2). This assemblage is thought to represent initial intrusions of anorthositic series magmas into the base of the volcanic pile during the main stage of Duluth Complex magmatism (~1099 Ma). The anorthositic rock lithologies, which dominate the Anorthositic Series throughout the Duluth Complex, are interpreted to have crystallized from plagioclase crystal mushes (evolved basaltic melts laden with suspended plagioclase crystals). However, in this area, these anorthositic rocks are locally intruded by oxide gabbro. Bonnichsen (1974b) speculated that the oxide gabbro, which he termed the Powerline Gabbro, may be a differentiate of the Partridge River intrusion. However, detailed mapping in the Allen quadrangle to the west (Severson and Miller, 1999) demonstrated that this gabbro is intruded by troctolitic rocks of the PRI. We consider the gabbro to be a plagioclase-poor component of the Anorthositic Series.

In a roadside exposure on the forest road, we can see a typical example of the "Powerline Gabbro". It is a coarse-grained, non-foliated, subophitic olivine oxide gabbro composed of 55% plagioclase, 20% granular olivine, 15% subophitic augite, 10%anhedral Fe-Ti oxide, and trace amounts of hornblende and biotite. The mafic phases locally show incipient granoblastic

recrystallization textures and plagioclase and augite display moderate clouding by the exsolution of oxide needles. Both features are indicative of thermal metamorphism.

About 200 m through the woods to the north of the forest road, we come upon the Erie/LTV railroad grade. Exposed along a 500m-long road cut on both sides of the grade is a complex mixture of two general rock types. Most of the exposure is a very fine-grained, dense mafic volcanic hornfels that is sparsely plagioclase porphyritic. In thin section, the hornfels displays a perfect granoblastic texture (Fig. 8-17). In sharp to abrupt contact with this volcanic hornfels is a medium grained plagioclase porphyritic leucogabbro. In thin section (Fig. 8-17), the leucogabbro is composed of strongly zoned plagioclase phenocrysts, which show deep clouding by oxide needles, in a matrix of granoblastic-textured oxide melagabbro (40%Pl, 30% Aug, 30% Fe-Ti Ox). At the contacts between the hornfels and the leucogabbro (best seen on the upper surface of the northern roadcut), the plagioclase phenocrysts in the leucogabbro show a strong alignment parallel to a steep contact with the hornfels. One possible interpretation of this leucogabbro is that it represents a chilled intrusion of plagioclase crystal mush that was emplaced into the volcanic pile early in the emplacement of the anorthositic series. Subsequent intrusions of more voluminous anorthositic series magma and later layered series magma forming the Partridge River intrusion produced the strong thermal metamorphism now evident in both the mafic volcanic and the porphyritic leucogabbro.

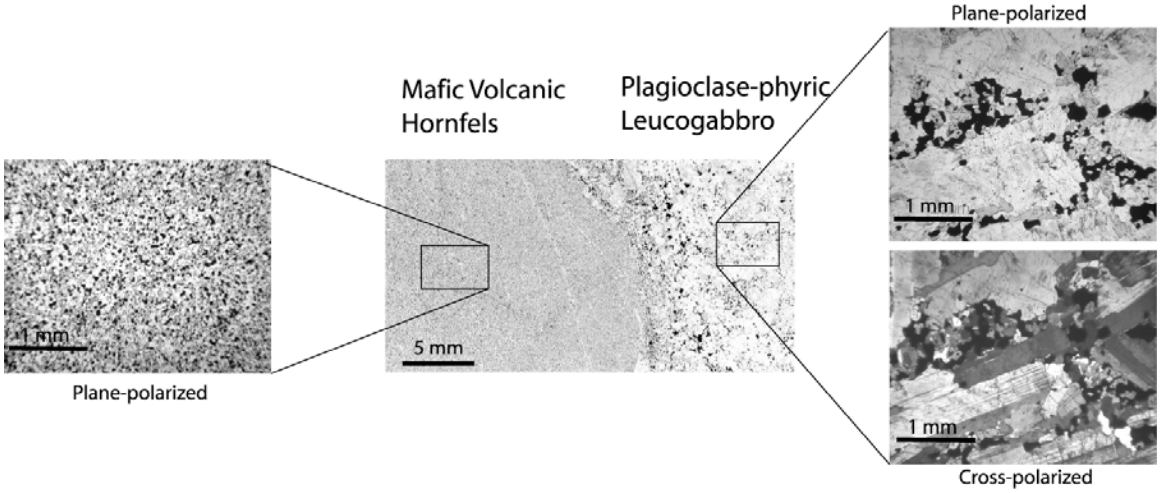


Figure 8-17. Photomicrographs of mafic volcanic hornfels and plagioclase-phyric leucogabbro observed at Stop 8-9.

STOP 8-10: <u>Cross-bedded Colvin Creek hornfels, Partridge River Intrusion</u>

Location: Approximately 0.5 miles south of where Forest Rd 113 crosses the south branch of the Partridge River and 520 meters (1700 feet) east along a flagged trail.

Babbitt SW 7.5' quadrangle; T.59N., R.13W., Sec 25, NE of SW of SE; 580255E 5267943N (NAD83)

Description: Magnetic basalt hornfels inclusions (most notably the Colvin Creek hornfels) within the Duluth Complex were first observed by Bonnichsen (1972) and first described in some detail by Tyson (1976). According to Tyson, the magnetic basalt flows of the Colvin Creek hornfels are characterized by fine-grained granoblastic rocks composed of plagioclase, augite, and magnetite with relict plagioclase- and augite-filled amygdules. Tyson (1976) noted a major difference between the magnetic basalts and the more common non-magnetic basalt inclusions (as in stops 8-3a and 8-4) and theorized that the magnetic basalts were derived from weathered and oxidized basalt flows that were subsequently metamorphosed by the Complex. Tyson's work was of a reconnaissance nature and much more detailed work has since been completed on the Northern Colvin Creek Body by Patelke (1996). The Northern Colvin Creek Body is a large inclusion (2,500 X 800 meters), associated with a magnetic high, that has been rotated to near vertical and exhibits excellent stratigraphic tops to the northwest. Patelke (1996) subdivided the northern Colvin Creek Body into five mappeable units including two metavolcanic units, two intrusive gabbroic sill units, and a cross-bedded sedimentary unit. Within the metavolcanic units, each consisting of multiple basalt flows, Patelke (1996) identified several volcanic features that include: pipe amygdules, sheeted amygdules, and local convoluted flow bases.



Figure 8-18. Photograph of cross-bedded "microgabbroic" sediment (X-BDD Unit) of the Northern Colvin Creek Body.

Enigmatic cross-bedded sedimentary rocks, with near-vertical dips, were first discovered in the Northern Colvin Creek hornfels and reported by Severson and Hauck (1990). They mapped a X-BDD Unit that exhibited a strike length of over 1,600 meters and an exposed thickness of over 240 meters. Overall the cross-bedded rocks are gabbroic in composition and consist of very fine-grained granoblastic concentrations of plagioclase, diopsidic augite, and magnetite with lessor amounts of ilmenite, orthopyroxene, and poikiloblastic pyroxene – the rocks do NOT contain

quartz or biotite, nor a basal conglomeratic unit where it overlies magnetic basalt (Patelke, 1996). Because sedimentary-like features were found in the cross-bedded rocks, and in the overlying gabbroic rocks (both with near-vertical dips), Severson and Hauck (1990) postulated that the cross-bedded rocks were deposited via magmatic density currents (a concept they no longer believe in). More recent studies by Patelke (1996) suggests that these cross-bedded rocks were sedimentary and were most likely deposited in a restricted basin as an eolian sediment that was derived from a strictly basaltic terrain (thus no quartz). Whatever their origin, these rocks exhibit beautiful sedimentary structures that include: bedding, cross-bedding, density-graded modal layering, and scour and fill structures (Patelke, 1996). Patelke (1996) suggests that the cross-bedded rocks of the Colvin Creek hornfels are similar to a thin sandstone unit near Phantom Lake, north of Two Harbors, MN. However, he also cautions that neither of these sedimentary units are analogous to any of the typical interflow sandstones of the North Shore Volcanic Group as described by Jirsa (1984).

Cross-bedded sediments with a gabbroic composition have been found at six locations within the Duluth Complex (most recently shown on a map in Severson, 1995; and in Severson and Miller, 1999). The last stop of this field trip is at one of the more "easily" accessible cross-bedded localities. At this locale, the cross-bedded sediments exhibit shallow dips to the northeast and only about 10 feet of stratigraphic section are exposed. The best exposures are present in the northern Colvin Creek Body which is located about two miles to the west of this stop in sections 27, 33, and 34, T.59N., R.13W. A more complete description of the units exposed at the northern body, and a thorough examination of their origin, can be found in Patelke (1996).

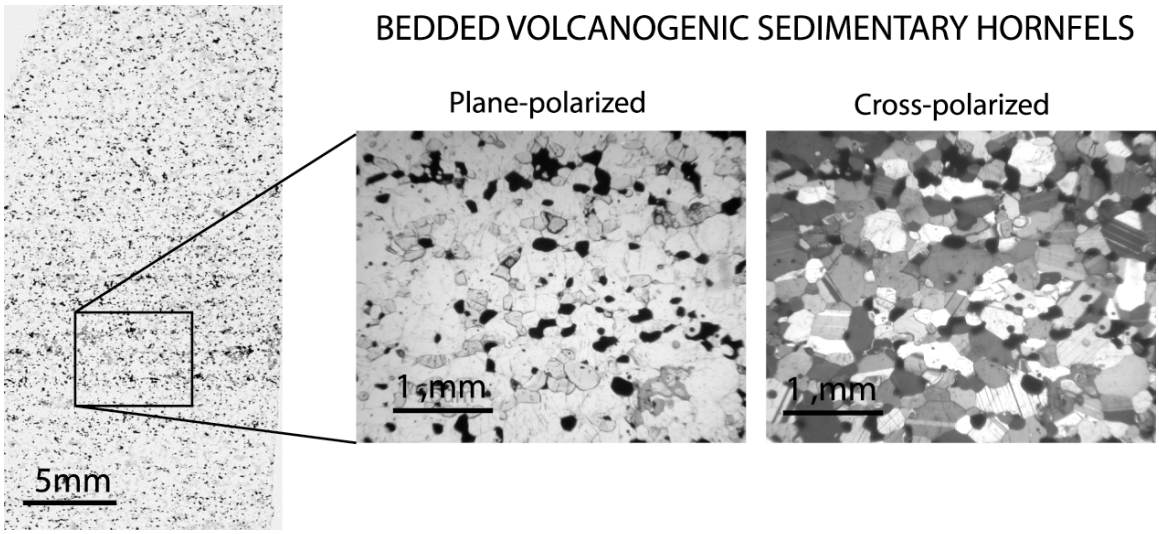


Figure 8-19: Scanned polished thin section image and photomicrographs of X-BDD Unit of the Northern Colvin Creek Body.

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