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FIELD TRIP GUIDEBOOK FOR THE GEOLOGY AND ORE DEPOSITS OF THE MIDCONTINENT RIFT IN THE LAKE SUPERIOR REGION

PREPARED FOR
THE 1995 IGCP PROJECT 336
FIELD CONFERENCE AND SYMPOSIUM,
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FIELD TRIP GUIDEBOOK
FOR THE
GEOLOGY AND ORE DEPOSITS
OF THE
MIDCONTINENT RIFT IN THE LAKE SUPERIOR REGION

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Prepared for the
1995 International Geologic Correlation Program (IGCP) Project 336
Field Conference and Symposium

Petrology and Metallogeny of
Volcanic and Intrusive Rocks of the Midcontinent Rift System

Duluth, Minnesota
August 19 - September 1, 1995

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International Geological Correlation Program
PROJECT 336
PETROLOGY AND METALLOGENY OF
INTRAPLATE MAFIC AND ULTRAMAFIC MAGMATISM

1995 Field Conference and Symposium

PETROLOGY AND METALLOGENY OF VOLCANIC AND INTRUSIVE ROCKS OF THE MIDCONTINENT RIFT SYSTEM

Duluth, Minnesota
August 19 - September 1, 1995

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PREFACE

This guidebook has been prepared for the field conference and symposium on "The Petrology and Metallogeny of Volcanic and Intrusive Rocks of the Midcontinent Rift System" that will be held in Duluth, Minnesota in August of 1995 as part of International Geological Correlation Program (IGCP) Project 336 ("Petrology and Metallogeny of Mafic and Ultramafic Magmatism"). The four-year-long IGCP project was begun in 1992 to investigate the petrology and metallogeny of various intracontinental mafic igneous provinces with an overall goal of establishing geological criteria for targeting mineral deposits in such environments.

The principal goal of the 1995 meeting is to showcase to the international scientific and exploration communities the significant advances in our understanding of the structure, mineralization, and magmatic history of the Midcontinent rift. Toward this end, three field trips have been organized around the three-day symposium to be held August 25-27, 1995 in Duluth. The field trips are designed to provide a broad overview of the geologic framework of the Midcontinent rift, where it is best exposed in the western Lake Superior region. Field trip I (August 19 - 24) looks at the physical volcanology of the flood basalts exposed on Isle Royale. Field trip II (August 22—24) highlights the stratigraphy, structure and mineralization of the volcanic and sedimentary rocks forming the southern limb of the Midcontinent rift in Wisconsin and Upper Michigan. The post-symposium field trip III (August 28—September 1) focusses on the geology and ore deposits associated with the intrusive rocks of the Midcontinent rift exposed along the north shore of Lake Superior in Minnesota and Ontario.

I would like to thank the talented people who contributed to this guidebook —Ted Bornhorst, Bill Cannon, John Green, Jeff Mauk, Suzanne Nicholson, Bill Rose, Ron Sage, Mark Smyk, and Laurel Woodruff—for their expertise in the rocks of which they speak, for their assistance in planning and implementing the field trips, and for submitting their contributions more or less on time. I also want to thank my 1995 IGCP Project 336 co-chairs, Penny Morton and Steve Hauck, for picking up the organizational slack that allowed me to work on this guidebook. I very much appreciate the critical help in logistical planning and registration by the staff at UMD's Continuing Education and Extension Service, especially (Dr.) Lynne Olson and Marge Erickson. Participation in the field trips by many foreign participants would not have been possible without a grant from the IGCP committee and the generous financial support of BHP International, WMC International Ltd., Falconbridge, Kennecott Exploration, INCO Exploration and Technical Services Inc., Cleveland Cliffs Foundation, the Minnesota Exploration Association, and the Department of Geology at the University of Minnesota in Minneapolis. Finally, I would like thank Dave Southwick (director) and G.B. Morey (associate director) of the Minnesota Geological Survey for their full and enthusiastic support of my involvement in this project.

Jim Miller
Field Trip Coordinator
1995 IGCP Project 336
August 14, 1995

THE MIDCONTINENT RIFT IN THE LAKE SUPERIOR REGION

James D. Miller, Jr., Suzanne W. Nicholson, and William F. Cannon

Introduction

For over a century, geologic, geochemical, and geophysical studies in the Lake Superior region have focused on the Middle Proterozoic volcanic, intrusive, and sedimentary rocks that compose the Midcontinent rift (MCR). In the past decade, however, such studies have vastly improved our understanding of the three-dimensional structure and tectono-magmatic evolution of the MCR. This has come about due largely to the acquisition of several kinds of high quality data: 1) high-resolution aeromagnetic data (Chandler, 1990; Teskey and others, 1991); 2) deep-crustal seismic reflection profiles (Behrendt and others, 1988; Chandler and others, 1989; Cannon and others, 1989; Hinze and others, 1990; McGinnis and Mudrey, 1991); 3) petrochemical and radioisotopic data (Brannon, 1984; Green, 1986; Sutcliffe, 1987; Paces and Bell, 1989; Nicholson and Shirey, 1990; Klewin and Berg, 1991; Lightfoot and others, 1991; Shirey and others, 1994); and 4) high precision U-Pb dates of volcanic and intrusive rocks (Davis and Sutcliffe, 1985; Palmer and Davis, 1987; Davis and Paces, 1990; Heaman and Machado, 1992; Paces and Miller, 1993). In addition to the rapid increase in empirical data, conceptual changes as to how continental flood basalts and other large igneous provinces develop and evolve (e.g., White and McKenzie, 1989; Campbell and Griffiths, 1990), particularly with regard to the role of mantle plumes, have played a major role in shaping the ideas about the origin of the MCR. We present here a brief overview of MCR geology, structure, chronostratigraphy, and mineralization, and present some current ideas about its tectono-magmatic evolution so as to provide an empirical and conceptual context for the geology to be viewed on the field trips described in this volume. For more detailed summaries, refer to MCR papers included in special volumes of Tectonophysics (1992, v. 213, p. 1-55) and Canadian Journal of Earth Sciences (1994, v. 31, p 617-720).

Regional Setting

Although buried by younger Paleozoic sediments over most of its length, the arcuate, segmented path of the MCR (Fig. I.1) is easily traceable along a 2500-km-long gravity and magnetic anomaly that projects southwest and southeast of exposures in the Lake Superior region. Along its southwestern arm, the rift crosses several geologic provinces (Van Schmus, 1992). In the Lake Superior region the rift arcs across a Late Archean (2.8-2.6 Ga) granite-greenstone terrane of the southern Superior province. From Lake Superior south to Iowa the rift crosses the east-northeast-trending 1.85 Ga Penokean orogen at a high angle. To the north the orogen is composed of Early Proterozoic metasedimentary rocks deposited at the continental margin of the Superior craton, and to the south, a suture zone (NF and SLT, Fig. I.1), consisting of an island arc assemblage of volcanic, intrusive, and immature sedimentary rocks. Fragments of commonly isotopically reset Archean rocks, including gneiss dated at 3.6 Ga, also occur within the orogen. Based on scattered drill holes through Paleozoic cover in Nebraska and Kansas, a segmented basin of the MCR cuts across the mostly orthogneissic rocks (1.8 to 1.75 Ga) and granitic rocks (1.65 Ga) of the Central Plains orogen (Sims and Peterman, 1986). The rift appears to terminate in northern Oklahoma among anorogenic granite and rhyolite (1.4 Ga) thought to be a thin veneer on the Central Plains orogen.

The eastern arm of the MCR is deeply buried beneath Paleozoic rocks of the Michigan basin and has been encountered in only a few deep drill holes. The gravity and magnetic signature of the MCR ends abruptly where as it crosses the geophysical trace of the Grenville front in southeastern Michigan at nearly a right angle (Fig. I.1). The Grenville front is a tectonic

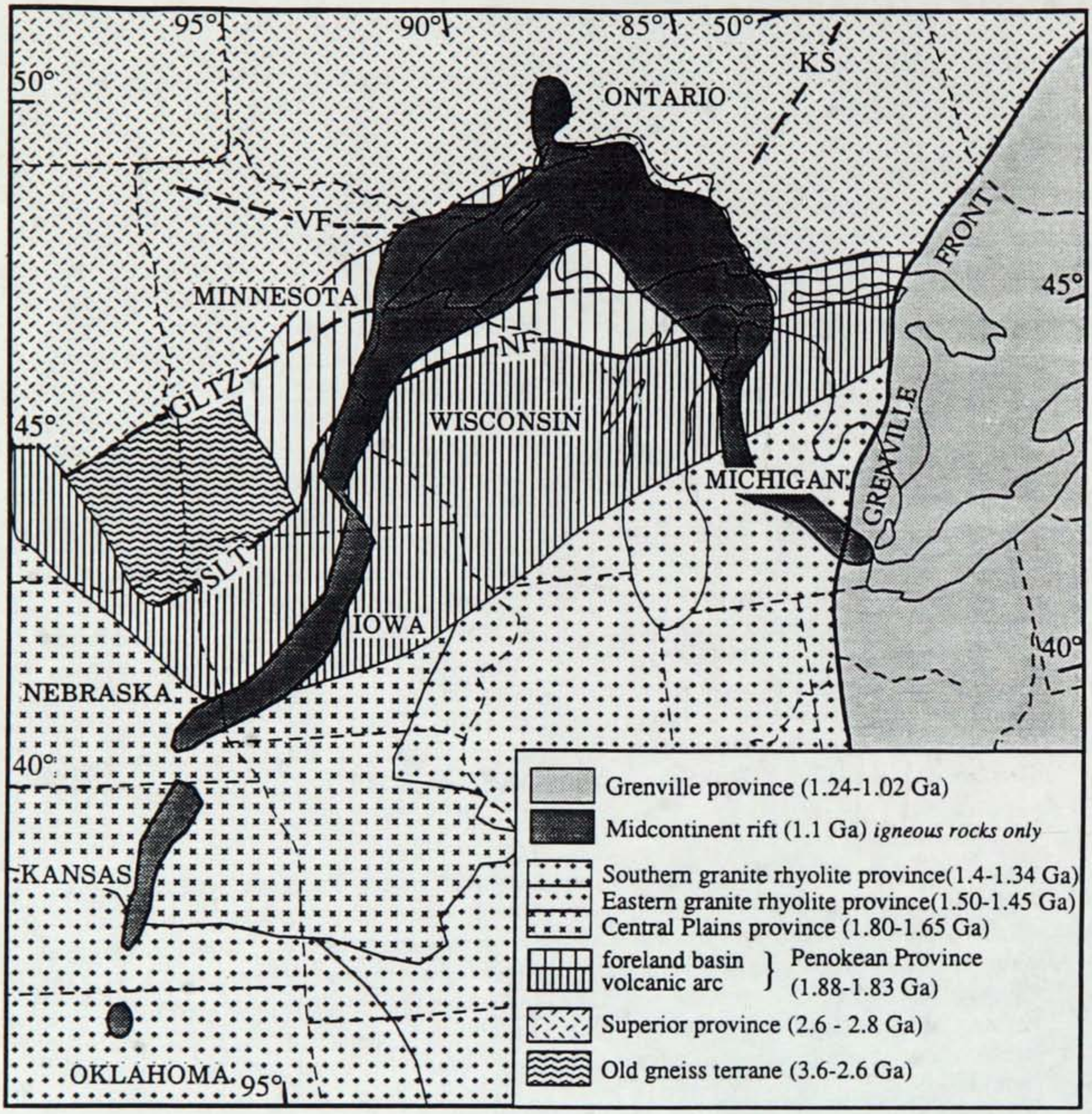

Figure I.1. Basement provinces of the midcontinent of North America (after Van Schmus, 1992) showing the position and extent of igneous rocks of the Midcontinent rift. Some major structural features denoted are the Kapuskasing structure (KS), Vermilion fault (VF), Great Lakes tectonic zone (GLTZ), Niagra fault (NF), and Storm Lake geophysical trend (SLT).

boundary marking the northwestern limit of penetrative deformation and metamorphic effect produced by the Elzevirian (1240 to 1160 Ma) and Ottawan orogenies (1090 to 1025 Ma), which together created the Grenville province (Easton, 1992). These two orogenies bracket the period of igneous activity and rifting of the MCR (1109-1086 Ma; Davis and Paces, 1990). Cannon (1994) has suggested that rifting of the midcontinent occurred during a period of diminished compression within the Grenville province when late tectonic inversion of the rift resulted from renewed tectonism of the Ottawan orogenic phase.

Geology

From exposures around Lake Superior, it has long been recognized that the MCR is composed of three major components: 1) a thick edifice of subaerial lava flows, 2) local concentrations of intrusive rocks, and 3) an upper sequence of sedimentary rocks (Fig. I.2). Recently acquired seismic profiles and gravity data indicate that the deepest part of the MCR beneath western Lake Superior contains as much as 30 km of fill, with volcanic rocks comprising about two-thirds of the total (Cannon and others, 1989; Allen, 1994). These geophysical data also suggest that a volume of magma nearly equivalent to that filling the rift, underplated the crust such that complete crustal separation probably occurred at least in the western Lake Superior area (Behrendt and others, 1988; Allen, 1994). Although faulting and burial beneath Lake Superior precludes study of a continuous and complete sequence of the rift stratigraphy, the main components may be pieced together from exposures around the Lake Superior basin (Fig. I.2).

The volcanic sequence, which is the focus of Field Trips 1 and 2, is composed of predominantly tholeiitic to subalkaline flood basalts, but also includes intermediate and felsic flows and fluvial interflow sedimentary rocks (Green, 1982). All lavas except for a few of the basal flows were erupted subaerially and most have the sheet-like form of flood basalts. Several notable exceptions are the intermediate to felsic composite volcanoes represented by the older Kallander Creek Volcanics of the Powder Mill Group (Cannon and others, 1993) and the younger Porcupine Volcanics of the Bergland Group (Cannon and Nicholson, 1992) in the Wisconsin-Michigan border area and the Michipicoten Island Formation (Annells, 1974) in eastern Lake Superior (Fig. I.2).

Petrochemical and isotopic data indicate that most basalts were derived from a primitive, high-Al olivine tholeiitic primary magma that was generated from a common, chondritic to mildly enriched mantle source (e.g., $\varepsilon_{Nd(i)}$ between -3 and +3; Paces and Bell, 1989; Nicholson and Shirey, 1990; Shirey and others, 1994; Nicholson and others, this abstract volume). Some compositionally distinctive picritic basalts with high *mg#*, low Al_2O_3, and elevated incompatible element abundances form the earliest flows in the Mamainse Point Formation (Klewin and Berg, 1991), the Osler Group (Lightfoot and others, 1991), and diabase intrusions in the Lake Nipigon area (Sutcliffe, 1987). These picritic compositions could not have been derived from the olivine tholeiitic primary magma, but instead appear to have been generated by lower degrees of partial melting of a deep (>100 km) garnet-bearing mantle source. Felsic and intermediate lavas show a broad range of crustal contamination effects and vary in their proportion of the volcanic sequence from less than 2% in the Portage Lake Volcanics of the Bergland Group (Nicholson, 1992) to as much as 25% in part of the North Shore Volcanic Group (Green and Fitz, 1993). A general, nonsystematic progression toward more primitive (higher *mg#*) compositions upward through the volcanic sequences is recognized throughout the rift (Green, 1982; Klewin and Shirey, 1992). Intrusive rocks, the focus of Field Trip 3, are found in three major complexes in the Lake Superior region: 1) the Coldwell Complex and related alkalic/carbonatitic intrusions of northwestern Ontario; 2) the Mellen Intrusive Complex near the Michigan-Wisconsin border; and 3) the Duluth Complex and related intrusions of northeastern Minnesota (Fig. I.2). All complexes were emplaced in multiple intrusive events of varied magma composition. The alkaline gabbroic to syenitic Coldwell Complex and many smaller intrusions, ranging from nepheline syenites to carbonatites, were emplaced as subcircular bodies into Archean crust along two zones extending north of the main rift zone (Currie, 1976). These zones correlate with older crustal structures and may indicate emplacement along zones of crustal weakness or reactivation. In contrast, both the

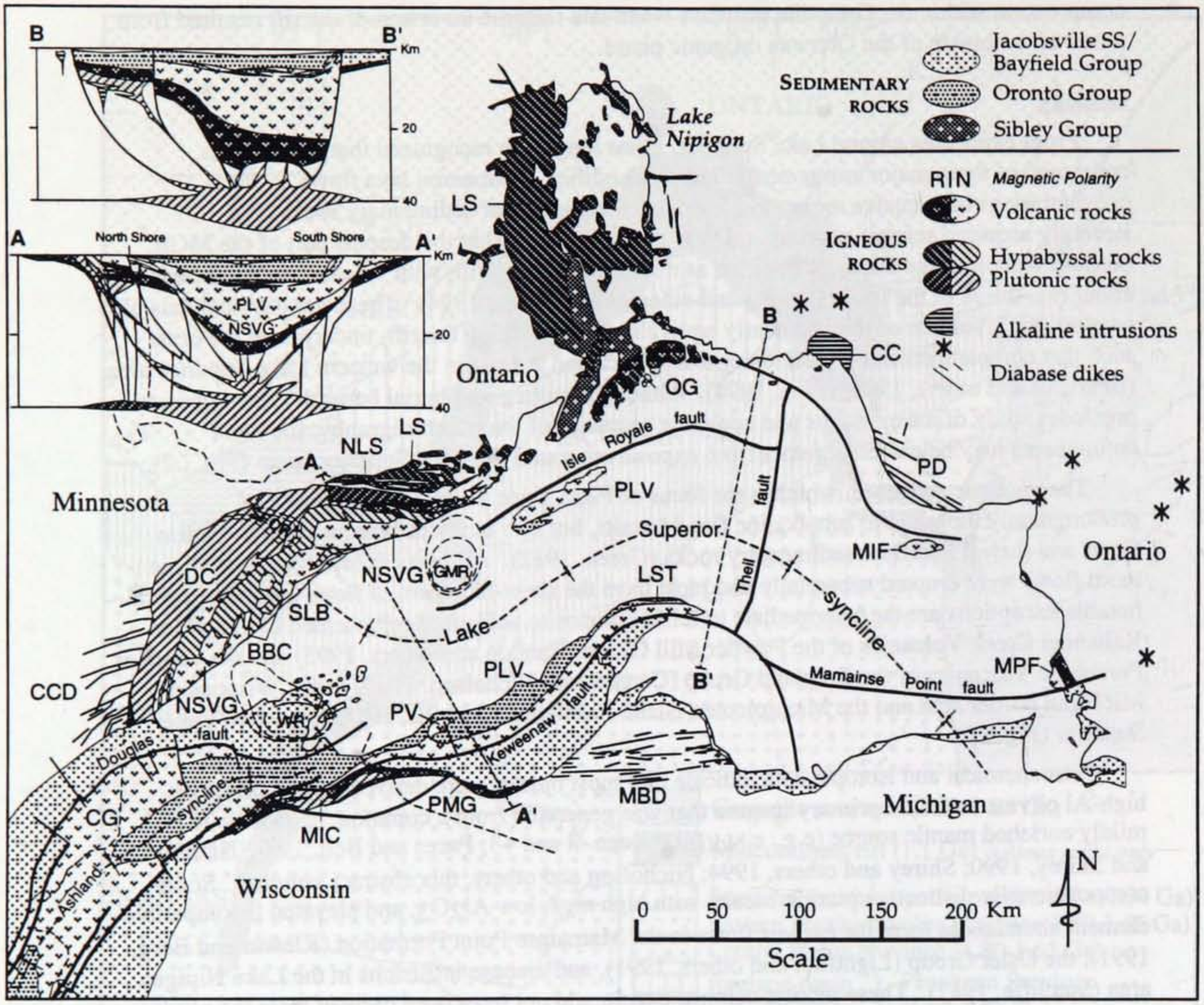

Figure I.2. Generalized Middle Proterozoic geology of the Lake Superior area. Abbreviations are major igneous units discussed in text: BBC- Beaver Bay Complex, CC- Coldwell Complex, CCD- Carlton County dikes, CG- Chengwatana Volcanic, DC- Duluth Complex, NLS- Nathan's (1969) layered series, NSVG- North Shore Volcanic Group, LS- Logan sills, LST- Lake Shore traps, MBD- Marquette-Baraga dikes, MIF- Michipicoten Island Formation, MIC- Mellen Intrusive Complex, MPF- Mamainse Point Formation, OG- Osler Group, PD- Pukaskwa dikes, PLV- Portage Lake Volcanics, PMG- Powder Mill Group, PV- Porcupine Volcanics, SLB- Schroeder-Lutsen basalts. Depth contours of buried crustal blocks are schematically portrayed as concentric dashed rings for White's ridge (WR), Grand Marais ridge (GMR), and Shroeder-Forest Center ridge (SFCR) (after Allen, 1994 and Miller and Chandler, in press). Major reverse faults (solid heavy line) and the axial traces of the deepest parts of the volcanic basins (heavy dashed line) are labelled Lake Superior syncline and Ashland syncline. Schematic cross sections A-A' and B-B' are based on geophysical interpretations of seismic, gravity, and aeromagnetic data by Cannon and others (1989), Sexton and Henson (1994), Teskey and Thomas (1994), and Allen (1994). Vertical exaggerartion in sections is about 2.5X. Chronostratigraphic correlations are displayed in Fig. I.3.

Duluth and Mellen are composed of tholeiitic mafic layered intrusions containing troctolitic, gabbroic, and anorthositic cumulates and some granophyric bodies. These tholeiitic complexes were emplaced as multiple sheet-like intrusions into the basal part of the comagmatic volcanic pile. Although the difference in size between the two complexes (Fig. I.2) reflects in part the steeper rift-ward dip of the Mellen Intrusive Complex, gravity data indicate that the Duluth Complex is significantly larger. The Duluth Complex has a surface area of over 5000 km^2, is situated over two of the largest Bouguer gravity anomalies (>50 mgals) associated with the MCR, and is estimated to be rooted to a depth of as much as 13 kilometers (Allen, 1994).

The largely fluvial redbed sedimentary rocks that fill the rift include four lithostratigraphic groups: 1) thin pre-volcanic quartzose fluvial and lacustrine deposits; 2) syn-volcanic interflow volcaniclastic sedimentary rocks; 3) post-volcanic, immature sedimentary rocks of the Oronto Group; and 4) quartzose sandstone of the Bayfield Group and Jacobsville Sandstone (Ojakangas and Morey, 1982). With the waning of volcanism, Oronto Group sedimentation filled subsiding axial rift basins with as much as 9 km of conglomerate, sandstone, and minor shale. With the formation of a central horst in some segments of the rift basin due to late compression, the Oronto Group sediments were locally reworked and deposited mostly into marginal basins where they accumulated to as much as 6 km in thickness (Figs. I.2 & I.3) to form the mature sandstones of the Bayfield Group and Jacobsville Sandstone (Allen, 1994).

Structure

The regional structure of the MCR has several complicating elements that cause it to deviate from the classic picture of an intracontinental rift (e.g., Rosendahl, 1987). The most obvious feature affecting this complexity is late reverse faulting that created a segmented axial horst of volcanic rocks flanked by thick sedimentary basins along most of the southwestern arm of the MCR (Fig. I.2). Recent seismic profiling has shown that the reverse faults accommodated about 30 km of shortening across most of the southwestern arm (Chandler and others, 1989; Woelk and Hinze, 1991; Anderson, 1992; Cannon and others, 1993). At least 5 km of reverse vertical displacement on the Keweenaw and Isle Royale faults (Fig. I.2) is evident on some seismic profiles (Cannon and others, 1989). Horst geometry is not as well defined on the southeastern arm of the rift. Seismic profiles and potential field data in eastern Lake Superior indicate much less shortening (Thomas and Teskey, 1994; Samson and West, 1994), roll-over structures common to transpressional faulting (Mariano and Hinze, 1994), and reverse faulting transverse to the rift axis (Manson and Halls, 1994). These relationships suggest that late compression was northwest-directed, thereby causing rift closure and inversion by reverse faulting along the southwestern arm and strike-slip or oblique faulting along the southeastern arm (Cannon, 1994).

Removing the effects of the late compression, the extensional structures of the rift also reveal themselves to be complex. Geophysical models across the MCR in the western Lake Superior region and along its southwestern arm, suggest that it was composed of a series of en echelon asymmetric graben that change their polarity across subtle to well-defined accommodation zones (Chandler and others, 1989; Cannon and others, 1989). Most of the volcanic rocks of the rift are confined within these now-inverted graben whose boundaries are commonly defined by the reverse faults (Hinze and others, 1982; 1992). Whereas 10 to 20 km of volcanic rocks commonly are contained within the graben, lava accumulations outside the graben rarely exceed 5 km (e.g., Fig. I.3). This implies that some of the reverse faults must have originally acted as growth faults and perhaps magma conduits to rapidly subsided and infilled axial graben (Cannon, 1992). Moreover, whereas volcanic rocks outside the central graben are predominantly older lava accumulations of reversed paleomagnetic polarity, younger normal polarity lavas comprise most of the volcanic sequences within the graben. This suggests that graben formation began some time after the initiation of volcanic activity (Cannon and others, 1989). Interestingly, in the eastern part of Lake Superior, where an axial graben is not well developed, older reversed polarity lavas comprise most of the sequence (Mariano and Hinze, 1994).

In western Lake Superior, the rift structure is further complicated by the effects of large, isolated crustal blocks within the volcanic basins. Integrated modeling of gravity, magnetic, and seismic data over western Lake Superior (Allen, 1994; Sexton and Henson, 1994) has identified two areas within the axial part of the rift where the volcanic section pinches out. These areas are presumed to be large blocks or ridges of granitic crust, called the Grand Marias ridge and White's ridge (Fig. I.2) by Allen (1994). These blocks may represent detached pieces of crust which did not subside and became isolated during separation. During volcanism, they stood as structural highs and exerted significant control on the shape of the graben into which the lavas accumulated, particularly the Portage Lake Volcanics (PLV). Allen (1994) demonstrated that the axis of the central rift basin (the Lake Superior Sycline, Fig. I.2) is centered between the Keweenaw Peninsula and Isle Royale and then curves around the Grand Marais ridge to the northwest toward the Minnesota coast. Miller and Chandler (in press) have suggested that the western growth fault margin of the PLV-equivalent rocks in Minnesota corresponds to an extensive arcuate dike and sill complex that is part of the hypabyssal Beaver Bay Complex. White's Ridge further divides the PLV basin from another deep trough to the southwest (the Ashland syncline) which hosts the Chengwatana Volcanics (Fig. I.2).

The Duluth Complex and related intrusions in northeastern Minnesota (Fig. I.2) also cause the MCR to deviate from a linear graben form. Modeling of Bouguer gravity data over the Duluth Complex, which is characterized by two broad highs of greater than 50 mgals, indicates that the complex extends to a depth of about 13 km (Allen, 1994). The saddle between the two gravity highs has been attributed to another granitic crustal ridge (SFCR, Fig. I.2; Miller and Chandler, in press) which divided the complex into two intrusive "basins". Although centered off the main axis of the rift, this accumulated thickness of magma is more than half of that which ponded in the central rift graben. Recent modeling of gravity data at the northern apex of Lake Superior by Thomas and Teskey (1994) suggest that a large mafic igneous complex with a thickness of as much as 20 km lies buried beneath the base of the Osler Group (Fig. I.2). The reason that such large volumes of mafic magma ponded along the northwestern margin of the MCR is unclear but may be related to various structural features of the pre-Keweenawan basement (Fig. I.1). The Duluth Complex lies near the projection of the Penokean tectonic front and the Great Lakes tectonic zone (an Archean suture zone), along the projection of major Archean fault zones such as the Vermilion fault, and at the northern shelf margin of the Early Proterozoic Animikie basin (Fig. I.1). More specifically, sheet intrusions of the Duluth Complex and related bodies in Ontario appear to have been emplaced along the nearly concordant interface of subhorizontal Early Proterozoic sedimentary rocks of the Animikie Group and lava flows within the Keweenawan Supergroup.

Chronostratigraphy

Prior to 1985, the principal means of correlating MCR volcanic sequences in the Lake Superior region was by their magnetic polarity and apparent pole position of remnant magnetism. In the western Lake Superior basin, most volcanic rocks were correlated relative to a single magnetic polarity reversal from early reversed to late normal. However, at Mamainse Point in eastern Lake Superior, an additional normal and reversed interval has been noted (Massey, 1983; Klewin and Berg, 1990). In the past decade, high-resolution U-Pb geochoronologic studies of zircons and badellyites have replaced paleomagnetism as the main correlation tool for the volcanic and intrusive rocks of the MCR and have proven to be very important in the rapid expansion of our current knowledge of the rift. Figure I.3 gives the general chronostratigraphy of the rift based on the current collection of over 20 published and unpublished dates. It is now well established that magmatic activity associated with the MCR spanned about 23 million years from 1109 to 1087 Ma. Moreover, there is a clustering of dates from both volcanic and intrusive rocks into three major populations: an early group at 1109-1107 Ma, a main stage group at 1102-1094 Ma, and a late stage group clustered at 1087 Ma.

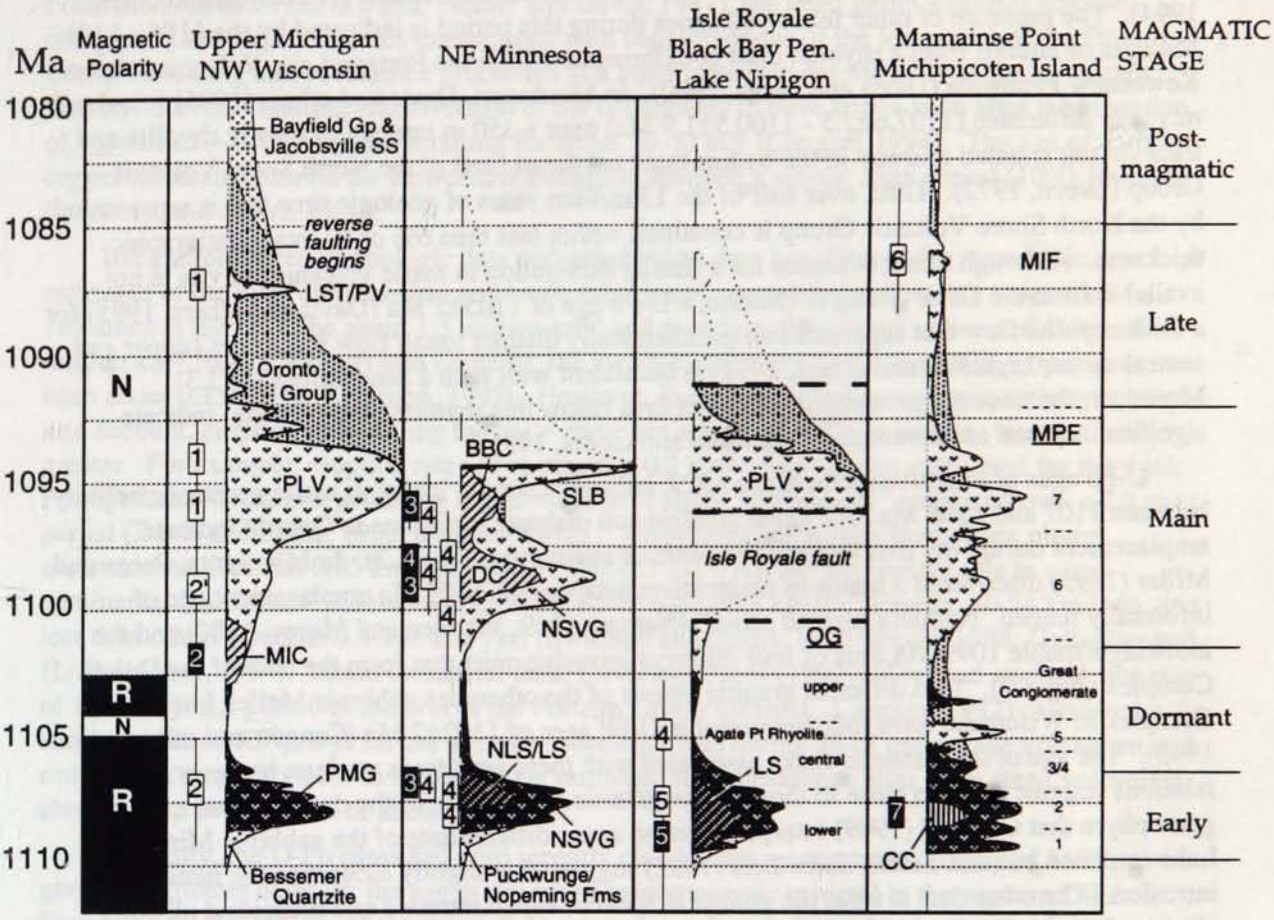

Figure I.3. Chronostratigraphic correlation diagram of magmatic and sedimentary units of the MCR in the Lake Superior basin. Right-handed increase of peak heights schematically portrays the relative levels of magmatic activity and basin subsidence for the various units. These parameters are interpreted from current U-Pb dates, paleomagnetic polarity, and the lithostratigraphy of the various units (see text for discussion). Unit patterns and abbreviations as in Figure I.2. Temporal positioning of flow units composing the Osler Group (OG, lower, central and upper suites, Lightfoot and others, 1991) and the Mamainse Point Formation (MPF, Units 1-7, Klewin and Berg, 1991; Shirey and others, 1994) are largely speculative and based on their chemostratigraphy. Heavy dashed lines mark gaps in the sequence due to loss of exposure, faulting truncation, or erosion. U-Pb dates of volcanic (open boxes) and intrusive (black boxes) rock are from 1) Davis and Paces (1990), 2) Cannon and others (1993) (2' Cannon, pers. comm.), 3) Paces and Miller (1993), 4) Davis and others, 1995, 5) Davis and Sutcliffe (1985), Palmer and Davis (1987), and 7) Heaman and Machado (1992).

An apparent hiatus or diminution of magmatic activity between 1107 and 1102 Ma (Fig. I.3) appears to be real at least among the exposed volcanic and intrusive sequences in the western Lake Superior basin. Most lava flow sequences spanning this interval are relatively thin and dominated by intermediate to felsic compositions. In the Wisconsin-Michigan border area, this time interval is represented by the upper part of the 2- to 4-km-thick Kallander Creek Volcanics (Cannon and others, 1993), which itself forms the upper part of the Powder Mill Group (Fig. I.3). The upper part of the Kallander Creek Volcanics ranges from basalt to rhyolite, but is dominated by andesite, and is thought to represent a partly eroded central volcano (Nicholson and others, 1994). The presence of other felsic volcanoes during this period is indicated by the 1106 - 1101 Ma ages of zircons from a rhyolite clast conglomerate within the Portage Lake Volcanics on the Keweenaw Peninsula (Davis and Paces, 1990). In Minnesota, Davis and others (1995) found a 7 m.y. age difference (1107.6±2.5 - 1100.5±1.9 Ma) over a 350 m interval of mostly rhyolite and trachybasalt situated midway in the 6+ km-thick northeast limb of the North Shore Volcanic Group (Green, 1972). Thus, over half of the 13 million years of geologic time that is represented by the North Shore Volcanic Group is contained within less than 6% of its total stratigraphic thickness. Although direct evidence for a similar diminution in mafic volcanic activity is not available from the Osler Group in Ontario, a U-Pb age of 1105±2 Ma (Davis and others, 1995) for a thick rhyolite flow that separates two geochemically distinct basalt flow sequences (upper and central suites, Lightfoot and others, 1991) is consistent with such a possibility (Fig. I.3). Moreover, the trace element compositions of lava below this rhyolite (central suite) indicate significant crustal contamination (Lightfoot and others, 1991).

U-Pb ages of intrusive rocks of the MCR indicate a similar hiatus in mafic magmatic activity between 1107 and 1102 Ma. All dated alkaline and diabasic intrusions in Ontario indicate emplacement during the first two million years of magmatism (Fig. I.3). In Minnesota, Paces and Miller (1993) discovered a hiatus in magmatism between the 1107 Ma emplacement age of informally named "Nathan's layered series" (Nathan, 1969; Weiblen and Morey, 1980) and the indistinguishable 1099 Ma ages of four different intrusive units that form the bulk of the Duluth Complex (Fig. I.2). Two different granitic phases of the otherwise gabbroic Mellen Intrusive Complex in Wisconsin have indistinguishable U-Pb ages of 1102±2 Ma (Cannon and others, 1993). The age of the gabbroic rocks associated with these granites is unclear; however, field relations suggest they are close to those of the granitic rocks. One of the dates is from a granophyre that Olmsted (1969) interpreted as an upper differentiate of the gabbroic Mineral Lake intrusion but that Seifert and others (1992) suggested is slightly older than the mafic intrusion. The other date is from the granite at Mellen, which intruded and brecciated all other phases of the Mellen Intrusive Complex (Olmsted, 1969).

Although the reduction in magmatic activity between 1107 and 1102 Ma appears to be real for the volcanic and intrusive suites mentioned above, it is not clear whether their magmatic histories are representative of the MCR as a whole. Seismic profiles over the western basin show that the Powder Mill Group and North Shore Volcanic Group cannot be directly correlated with thick early accumulations which apparently are buried beneath the main rift axis (Cannon and others, 1989). This may indicate that the presently exposed volcanic sequences accumulated marginally to a now-hidden axial graben basin and are therefore not representative of all volcanic activity in the western basin during this early stage. If this is the case, volcanism became focussed into axial graben earlier in the evolution of the western basin of the MCR than present exposures would suggest (Cannon, 1992). In the eastern basin, where graben formation was minimal throughout volcanic activity, a greater thickness of reversed lavas beneath eastern Lake Superior suggests either a greater rate or longer period of basaltic eruption in the early evolution of the MCR (Mariano and Hinze, 1994). The recognition of two extra paleomagnetic reversals in the Mamainse Point Formation may indicate more or less continuous eruption in the eastern basin, during the period between 1107 and 1100 Ma, while the western basin was in a relatively dormant stage (Fig. I.3). Nevertheless, a period of volcanic diminution is indicated in the medial part of the Mamainse sequence as well by the occurrence of thick polymict conglomerates (Fig. I.3) and an abundance of felsic rocks (Annells, 1973). Thorough U-Pb dating of this most complete section of MCR magmatism (Shirey and others, 1994) is clearly needed.

The late stage of volcanism (Fig. I.3) is bracketed between the upper part of the Portage Lake Volcanics (1094.0±1.5 Ma, Davis and Paces, 1990) and the age of the Michipicoten Island Formation (1086.5+1.3/-3.0 Ma, Palmer and Davis, 1987), the latter being a central volcano complex. Late stage volcanics interdigitate with conglomerates of the lower part of the Oronto Group indicating that subsidence proceeded at a significant rate despite the waning of volcanic activity. Reverse faulting and inversion of the rift appears to have begun soon after the cessation of volcanism (~1080 Ma) and persisted for about 30-40 m.y (Cannon, 1994). The age of native copper mineralization on the Keweenaw Peninsula spanned a similar time period (1060-1045 Ma, Bornhorst and others, 1988).

Integration of geochronologic data and geophysical data has allowed for reasonable estimates of the rates of volcanism and basin subsidence. Assuming the present volume of volcanics in the rift to be about 1.5 million km^3 and an originally erupted volume of at least 2 million km^3 (taking erosion into account), the average eruption rate for the MCR would have been about 0.15 km^3/yr (Cannon, 1992). However, if a 5 m.y. hiatus in volcanic activity is taken into account, eruption rates during the early and main stages of volcanic activity were probably greater. For example, eruption rates of as much as 0.2 km^3/yr have been calculated for the thick Portage Lake Volcanics, assuming a total volume of about 500,000 km^3 emplaced over a 2.2 m.y. period (Davis and Paces, 1990; Cannon, 1992) . These rates are greater than ocean plume environments (Iceland 0.05 km^3/yr; Hawaii 0.03-0.1 km^3/yr), but are comparable to some continental flood basalt provinces (Columbia River 0.07-0.28 km^3/yr; Parana, >0.24 km^3/yr), and less than others (Deccan 0.4-1.0 km^3/yr) (Swanson and others, 1975; Baski, 1988; Gallagher and Hawkesworth, 1994). Based on similar data, Davis and Paces (1990) calculated a subsidence rate of 1.3 mm/yr for a 2850 m interval of the Portage Lake Volcanics. Cannon (1992) calculated a similar value of 1.5 mm/yr for the average subsidence rate during all of main stage volcanism and estimated a range from 0.5 to 5 mm/yr. He estimated that subsidence after main stage volcanism slowed to an average rate of about 0.2 mm/yr.

Ironically, the U-Pb dates had until recently clouded the magnetostratigraphy of MCR, giving conflicting dates for the age of the major paleomagnetic reversal in the western Lake Superior region. The age of the major R-N reversal was first estimated to have occurred around 1098 Ma based on a date of 1097±4 Ma for the uppermost reversely polarized lava in the Osler Volcanic Group (Davis and Sutcliffe, 1985) and an age of 1096.2±1.8 Ma for the lower part of the normally polarized Portage Lake Volcanics (Davis and Paces, 1990). However, recently Davis and others (1995) reanalysed zircons from the Osler sample and acquired an age of 1105±2 Ma. With the oldest normally polarized age in the western Lake Superior basin being the 1102±2 dates for the granitic rocks of the Mellen Complex, the age of the major reversal is now pegged between 1105 and 1102 Ma (Fig. I.3).

Mineralization

The MCR hosts two major classes of mineral deposits, magmatic and hydrothermal. All important mineral production in this region has come from the world-class hydrothermal deposits, whereas mineral deposits related to magmatic rocks have provided only a small fraction of the total mineral production from the rift (Fig. I.4). The following discussion draws heavily on the compilation by Nicholson and others (1992) and references cited therein.

Rift-related hydrothermal deposits include four main types: 1) native copper deposits in basalts and interflow sediments; 2) sediment-hosted copper sulfide and native copper; 3) copper sulfide veins and lodes hosted by rift-related volcanic and sedimentary rocks; and 4) polymetallic (five-element: Ag-Ni-Co-As-Bi) veins in the surrounding Archean country rocks. The scarcity of sulfur within the rift rocks resulted in the formation of vary large deposits of native metals. Where hydrothermal sulfides occur (ie., shale-hosted copper sulfides), the source of sulfur was local sedimentary rocks.

The native copper deposits of the Keweenaw Peninsula have a long history of prehistoric mining, and modern production began in the late 1800's. These deposits were the principal source of copper for the United States for many years, yielding more than 5 million metric tons of refined Cu from 1845 until the 1960's when the last mines closed. In addition to the Keweenaw Peninsula deposits, native copper is found in virtually all exposed mafic volcanic rocks in the area, and unsuccessful attempts to develop profitable mines, mostly in the late 1800's, were

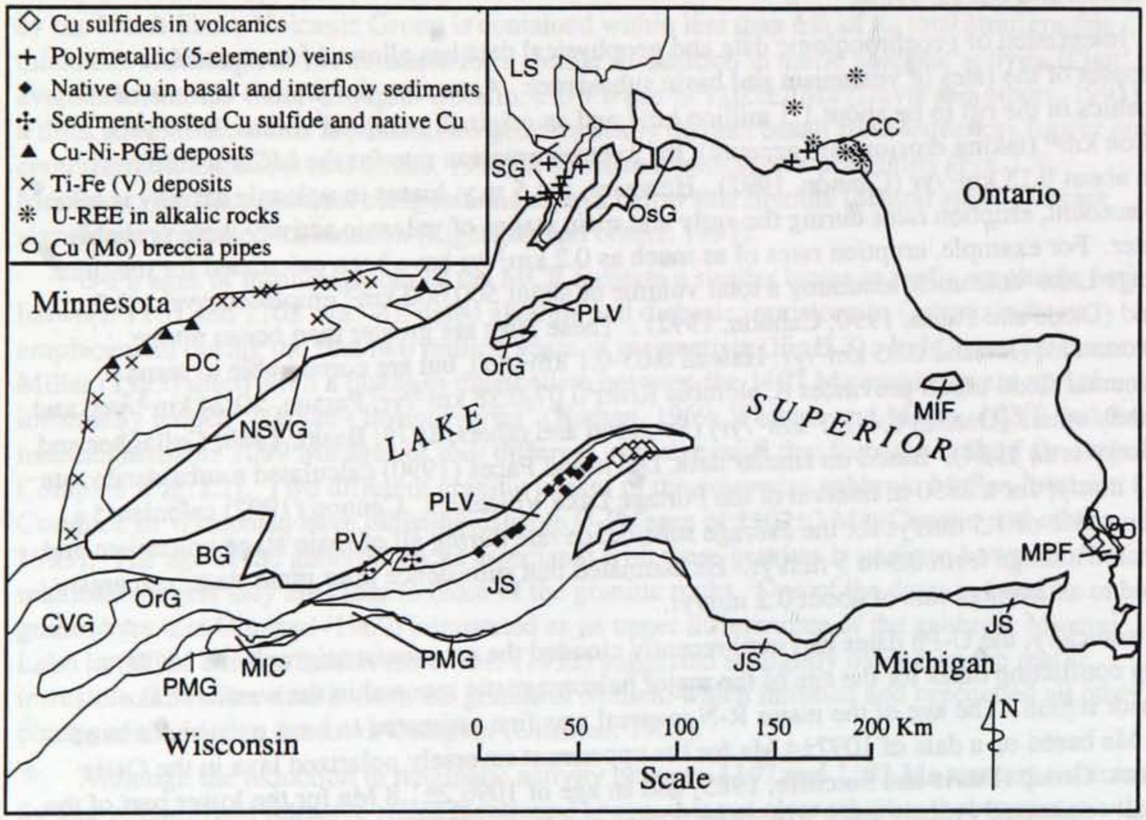

Figure I.4. Map of Lake Superior region showing locations of major ore deposits occurring with Middle Proterozoic rocks. Detailed geology given in Figure I.2. Abbreviations of major units include: BG - Bayfield Group, CC - Coldwell Complex, CVG - Chengwatana Volcanic Group, DC - Duluth Complex, JS - Jacobsville Sandstone, NSVG, North Shore Volcanic Group, LS - Logan Sills, MIF - Michipicoten Island Formation, MIC - Mellen Intrusive Complex, MPF - Mamainse Point Formation, OrG - Oronto Group, OsG - Osler Group, PLV - Portage Lake Volcanics, PMG - Powder Mill Group, PV - Porcupine Volcanics.

widespread. Native copper, locally accompanied by small amounts of native silver, occurs in the brecciated and vesicular tops of basalt flows (amygdular lodes), in interflow sedimentary rocks (conglomerate lodes), and in cross-cutting veins (fissures) (White, 1968, 1971). The copper typically fills open-spaces or replaces basalt. The major native copper deposits occur in zones where prehnite and pumpellyite are the major alteration minerals and epidote is absent or scarce (see Stoiber and Davidson, 1959; Jolly, 1974; and Livnat, 1983, for discussions of alteration patterns).

At the White Pine mine in Michigan (Field Trip 2, Day 2), sedimentary copper sulfide and native copper deposits in the lowermost part of the Nonesuch Formation comprise one of the world's largest shale-hosted copper deposits (Ensign and others, 1968; Brown, 1971; Kelly and Nishioka, 1985; Mauk and others, 1989a, 1989b; Seasor and Brown, 1989). In the area of active mining, the ore body is estimated to contain about 200 million tons (181 million mt) of ore with an average grade of 1.1% Cu and 0.25 oz (9 g) Ag/ton (Mauk and others, 1989a). An additional area in the Presque Isle syncline has been extensively drilled and also contains substantial mineralization. At White Pine, the copper is mostly stratabound within the most organic-rich beds of the Nonesuch Formation. Two stages of mineralization are recognized (Mauk and others, 1989a, 1989b): 1) chalcocite layers formed during the main stage of mineralization, probably as a result of late diagenesis, accounting for 80-90% of the contained copper; and 2) during the second stage, native copper was deposited largely in structurally disturbed zones, mostly the result of thrust faulting.

The four different types of rift-related hydrothermal deposits can be related to a regional model that includes heating of basinal brines, leaching of metals, and movement of fluids upsection (Nicholson and others, 1992). More than 1 million km^3 of mafic magma was erupted in the rift and a comparable volume of mafic intrusions is inferred beneath the rift, providing a ready and structurally confined supply of mafic source rocks that were available for leaching of metals by basinal brines. These brines were heated by a steep geothermal gradient that resulted from the melting and underplating of magma derived from the mantle plume. Hydrothermal deposits were emplaced at least 30-40 m.y. after rift magmatism and extension ceased. This time lag may reflect either the time required to heat deeply buried rocks and fluid within the rift, or may be due to timing of post-rift compression that may have provided the driving mechanism for expulsion of hydrothermal fluids from deep portions of the rift.

Of far less economic importance historically, the magmatic deposits occur in intrusions exposed near the margins of the rift and include: 1) Cu-Ni-PGE mineralization in the Duluth Complex and other smaller intrusions; 2) Ti-Fe (V) in the Duluth Complex; 3) U-REE in small carbonatites; and 4) Cu (Mo)- bearing breccia pipes resulting from local hydrothermal activity around small felsic intrusions. The ages of the magmatic rocks, where known, fall between 1109 and 1086 Ma. Mineralization associated with some magmatic bodies resulted from concentration of incompatible elements during fractional crystallization (e.g., Ti-Fe (V), REE, U, TH, Ba, F) . Most of the sulfide deposits in intrusions, however, contain sulfur derived from country rocks: interaction between magma and country rocks was important in generation of the magmatic Cu-Ni sulfide deposits (e.g. Ripley and Al-Jassar, 1987).

The copper-nickel sulfide mineralization of the basal portion of the Duluth Complex in Minnesota has been delineated along a 50 km belt southeast of Ely, Minnesota and estimated at more than 4 billion metric tons of mineralized rock containing 0.66% Cu and 0.2% Ni (Listerud and Meineke, 1977) with additional large resources of lower-grade rock in the northeast portion of the complex (Martineau, 1989). Associated platinum-group-element (PGE) mineralization in the complex has just begun to be evaluated as a resource (Morton and Hauck, 1987; Severson, 1994; Ripley and Chryssoulis, 1994). The Cu-Ni mineralization occurs at or near the base of the Duluth Complex, hosted mostly by troctolite intruded between slightly older anorthositic series units and the country rocks. Sulfide mineralization consists of mostly pyrrhotite, chalcopyrite, cubanite and minor pentlandite, generally occurring as disseminated grains among silicate phases in the troctolite, but also as veinlets, inclusions in silicates, intergrowths with hydrous minerals, and as rare massive sulfide segregations (Weiblen and Morey, 1976; Ripley and Watowich, 1982; Severson, this abstract volume). The copper-nickel mineralization resulted from the interaction of the metal-bearing intruding magma and sulfur-rich fluids derived from desulfurization and dehydration of country rocks and was accompanied by at least some local partial melting and assimilation (Ripley, 1981; Weiblen and Morey, 1980; Foose and Weiblen, 1986; Ripley and Al-Jassar, 1987; Ripley and others, this abstract volume; Zanko and others, this abstract volume).

Tectono-magmatic Evolution

Recent geochemical, geophysical and geochronologic data indicate that the rift evolved in several stages (Fig. I.5), with each stage characterized by distinctive magma compositions, structural styles, and rates of eruption and subsidence. Moreover, these data suggest that, at particular stages, these characteristics varied over the breadth of the MCR in the Lake Superior region.

Prior to the onset of volcanism, a gentle regional basin developed over most of the western Lake Superior area into which quartz-rich fluvial and lacustrine sediments were deposited. Between 1109 and 1107 Ma during a period of reversed magnetic polarity, the early volcanic stage was characterized by eruption of mostly moderately differentiated plateau basalts into broad basins, intrusion of tholeiitic diabase sills and alkaline bodies, and the formation of regional dike swarms (Fig. I.5A). Earliest magmatism along the northern and eastern margin of the rift basin involved emplacement of minor amounts of picritic basalt. Mafic to syenitic magmas of alkaline and carbonititic affinities were emplaced along the northern flank of the rift. Most reversely polarized lava sequences and intrusions in the western Lake Superior area were emplaced as variably evolved subalkalic to tholeiitic basalt with minor amounts of andesite and rhyolite.

After 1107 Ma, volcanism in at least the margins of the western rift basin had slowed significantly, and eruption of intermediate to felsic volcanic rocks increased, commonly localized in flanking shield volcanoes (Fig. I.5B). Basaltic eruptions diminished in the western rift basin, or were erupted into a central basin that now lies buried beneath later volcanic accumulations and beneath Lake Superior. Until U-Pb dates are acquired on the Mamainse Point Formation, it is difficult to determine whether basaltic magma production waned in the eastern part rift basin during this period as well.

The main stage of magmatic activity, which created the bulk of the total igneous fill of the MCR, began about 1102 Ma and continued until at least 1092 Ma (Figs. I.3 & I.5C). During this interval, rates of magma production and basin subsidence were comparable to those estimated for many large igneous provinces. By the end of this stage, rifting had thinned the crust to less than 25% of its original thickness in the Lake Superior area. The main stage is represented by the nearly 10 km thickness of normally polarized North Shore Volcanic Group lavas in Minnesota and the at least 5 km of Portage Lake Volcanics in Michigan (and possibly the upper part of the Kallander Creek Volcanics in Michigan and Wisconsin as well). The first few m.y. of main stage magmatism involved a differentiated range of tholeiitic basalt and minor felsic compositions. Over the last 2 m.y. of the main stage, however, volcanism came to be dominated by primitive to mildly evolved, high-Al olivine tholeiitic basalts so characteristic of the Portage Lake Volcanics (Paces, 1988). Along most of the southwestern arm of the MCR, lava flows accumulated in a series of central growth fault-bounded grabens. In the western Lake Superior area, this geometry was more complex due to the effects of large isolated crustal blocks and the development of the "intrusive basins" into which the Duluth Complex and related buried complexes further east were emplaced. By 1096 Ma, however, most magmatism in the western basin area was focussed into a

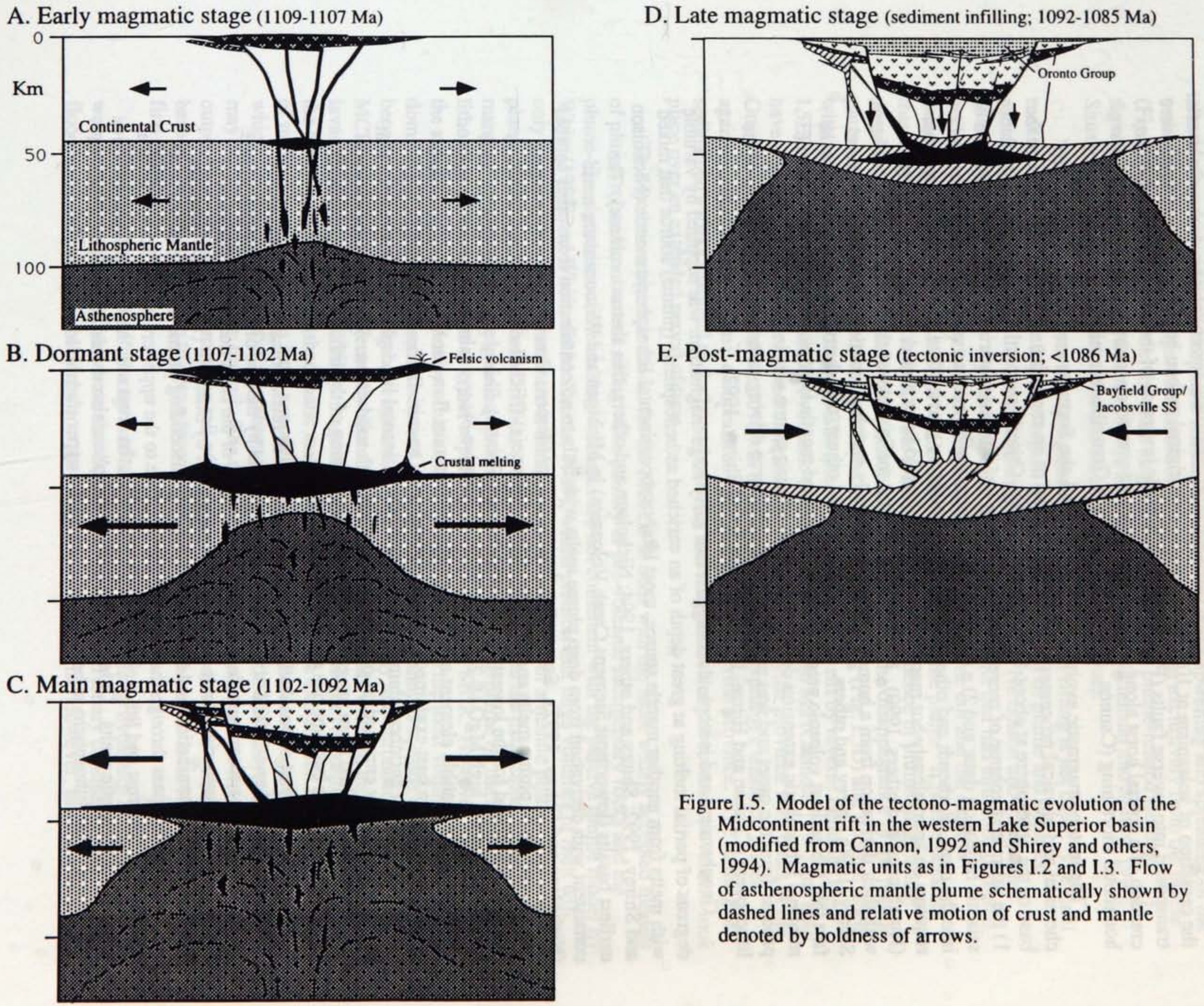

Figure I.5. Model of the tectono-magmatic evolution of the Midcontinent rift in the western Lake Superior basin (modified from Cannon, 1992 and Shirey and others, 1994). Magmatic units as in Figures I.2 and I.3. Flow of asthenospheric mantle plume schematically shown by dashed lines and relative motion of crust and mantle denoted by boldness of arrows.

central, albeit curved, graben. Along the southeast arm, a thinner sequence of lavas accumulated in broad basins.

Between 1092 and 1085 Ma, volcanism waned while subsidence of the central grabens continued (Fig. I.5D). Consequently, the grabens became filled with a thick accumulation of fluvial and some lacustrine sedimentary rocks and progressively fewer volcanic rocks. Soon after the cessation of volcanism at 1085 Ma, many of the original graben-bounding normal faults were transposed into reverse faults (Fig. I.5E). The tectonic inversion of the rift that resulted in the creation of a central horst along most of the southwest arm was accomplished by about 30 km of horizontal shortening (Cannon, 1994), in response to Grenville tectonism.

The tectonomagmatic evolution of MCR in the Lake Superior area can be explained by chemical and physical processes attending the arrival of an anomalously hot mantle plume at the base of the lithosphere at about 1109 Ma (Fig. I.5A). Criteria supporting a plume model include: 1) the estimated volume of more than 2 million km^3 of erupted material (Cannon, 1992) based on seismic reflection data; and 2) a nearly equivalent amount of magma underplated and intruded into the crust (Berhendt and others, 1990) as suggested by gravity data. This volume of magma requires an anomalously hot mantle source such as would be expected from a mantle plume (Hutchinson and others, 1990). Isotopic data on most mafic rocks of the MCR are also consistent with their derivation from a homogeneous, isotopically enriched mantle plume (Nicholson and Shirey, 1990; Shirey and others, 1994; Nicholson and others, this abstract volume). Indeed, the fact that the earliest volcanic products show undepleted mantle isotopic signatures at a time of minimal lithospheric extension demonstrates that a rising plume was involved in the rifting process from the onset and was probably an integral force driving extension and thinning of the lithosphere (Cannon and Hinze, 1992; Nicholson and Shirey, 1992).

Geochemical and isotopic data suggest that early stage magmatism was supplied by variable degrees of partial melting at great depth of an enriched asthenospheric (plume) source that mixed with melts from another mantle source, most likely subcontinental lithospheric mantle (Nicholson and Shirey, 1992; Shirey and others, 1994; Nicholson and others, this abstract volume). The earliest basalts (lowermost Siemens Creek Volcanics) in Michigan and Wisconsin are most consistent with derivation from deep plume melts with little to no contribution from other sources (Nicholson and Shirey, 1992). Succeeding basalts (upper Siemens Creek Volcanics) resulted from a mixture of plume melts and melts from other source(s) (Nicholson and Shirey, 1992) as has been postulated for the lowest groups at Mamainse Point by Shirey and others (1994), who concluded that the high MgO (6-20 wt %), variably negative $e_{Nd(i)}$ values (-0.5 to -6.0), and distinctive incompatible element abundances of the lowermost flows at Mamainse Point (their Groups 1 and 2) are best explained as being derived from an initially deep (~100 km) source which combined an enriched plume source with a subcontinental lithospheric mantle source that had been previously enriched by processes associated with subduction. These earliest volcanic rocks probably resulted from progressive melting and erosion of the lithospheric mantle in conjunction with the buoyant rise and spread of the plume.

The diminishment of basaltic volcanism and increase of felsic volcanism in parts of the MCR between 1107 and 1102 Ma are not readily explained by the rise of a single mantle plume. Perhaps this period reflects ponding of magma at the base of the crust due to delamination of subcontinental lithospheric mantle (Fig. I.5B) as suggested by Sutcliffe (1987). Early on, the absence of such delamination and the overpressure of an ascending plume would have caused magmas to readily pass through the crust-mantle interface to the upper crust (Fig. I.5A). With a decoupling of the crust and lithospheric mantle, however, the crust may have acted as an effective density filter to mantle melts. Delamination of the lithosphere also would have allowed most of the thinning of the lithosphere to be accomplished by the upper mantle, leaving the crust only moderately attenuated as the relatively thin edifice of early volcanics suggests. Wholesale ponding of magma could have caused extensive melting of the lower crust to give rise to the felsic magmatism so prevalent during this period of reduced volcanic activity.

The onset of main stage magmatism and graben development probably reflects the complete shouldering aside of the lithospheric mantle such that continued upwelling and spreading of the asthenosphere acted with vigor to thin the crust and supply magma exclusively from the plume (Fig. I.5C). Most basalts from the main stage of volcanism show plume-related isotopic signatures (Brannon, 1984; Dosso, 1984; Paces and Bell, 1989; Nicholson and Shirey, 1990; Shirey and others, 1994; Nicholson and others, this abstract volume).

The reason for the cessation of magmatic activity and crustal rifting is not clear. The final melts generated in the Lake Superior region appear to be mixtures of plume and depleted asthenospheric (N-MORB-like) mantle (Shirey and others, 1994; Nicholson and others, this abstract volume). This may represent melting of parts of the plume that had entrained depleted asthenospheric mantle or perhaps melting of both the narrow plume tail and the shallow asthenospheric mantle that infilled behind the dissipated head of the plume. It is also possible that magmatism ended due to drift of the plume: however, no evidence of plume magmatism postdating 1086 Ma is known in the Midcontinent region. With the disappearance of the mantle plume, the rift experienced thermal collapse and sediment loading (Fig. I.5D) and became filled with as much as 8 km of sediments. The tectonic inversion of the rift by reverse faulting (Fig. I.5E) probably began with the last volcanic eruptions at about 1086 Ma (Cannon, 1994) and may have been due either to rejuvenated compression within the Grenville province with onset of the Ottowan orogenic phase at about 1090 Ma or simply to the loss of plume's tensional forces against long standing regional compression provided by the Grenville province. The folding of sedimentary rocks marginal to the horst-bounding faults indicates that the reverse movement took place over an extended period of time.

This tectono-magmatic model for the MCR is at odds with the more widely accepted models of plume-related continental rifting. The paucity of pre-volcanic sedimentary rocks and the plume-like signatures of the earliest volcanics are not consistent with the passive plume model by White (1992) and White and McKenzie, 1989), which predicts that large volumes of melt will only be generated if the lithosphere is thinned prior to the arrival of a mantle plume. The active plume model by Campbell and Griffiths (1990), which emphasizes the extensional force of the mantle plume on the lithosphere and the erosive capabilities of the plume on the subcontinental lithospheric mantle, appears to better explain the evolution of the MCR. However, their model of the structure of the plume predicts that early volcanism should be the most voluminous and dominated by pure plume component and subsequent volcanism should subside in rate and become more geochemically heterogeneous. The lithostratigraphy and chronostratigraphy of the MCR volcanic sequences generally demonstrate the opposite trends with the most voluminous lavas generated late in the 23 m.y. magmatic history of the rift. Moreover, neither model explains the 5 m.y. period of diminished mafic magmatism. An alternative explanation for the diminution of mafic magmatism comes from a model recently proposed by (Bercovici and Mahoney, 1995), which suggests that as plumes pass through the upper-lower mantle boundary, the plume head may become detached from the tail. The detached head will initiate the first flood basalt outpouring, whereas the remaining tail will develop a second head, which, when it arrives at the base of the lithosphere, can trigger a second outpouring of flood basalt at least 10 m.y. after the first pulse.

The antiquity of the MCR may lead to the conclusion that plume dynamics in the Proterozoic were somehow different from those that influenced Mesozoic to Tertiary continental rifting and flood basalt volcanism and which form the basis of current models. However, as Shirey and others (1994) pointed out, the magmatic progression seen in the MCR is similar in many ways to other continental flood basalt provinces. Because the MCR presents one of the most completely exposed sections of flood basalts, by virtue of its incomplete rifting and later tectonic inversion, its tectono-magmatic evolution should also be seriously considered as a test and constraint of general models of plume-generated continental rifting and magmatism.

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FIELD TRIP 1

VOLCANIC GEOLOGY OF ISLE ROYALE, MICHIGAN

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OVERVIEW

Preface (or why people don't go to Isle Royale)

Isle Royale has very few visitors, especially considering that it is a national park and is, for many people, within a couple of days travel. This lack of tourism can be explained partially by the park's island location and by the fact that a trip to Isle Royale seems to require a deeper commitment, of both time and money, than other vacations might. The very people whom you would most expect to want to visit Isle Royale don't go.

When you compare the popularity of various national parks, the public's avoidance of Isle Royale is obvious, perhaps even more obvious to me because of my position. As a professor at Michigan Technological University for more than 20 years, I have had direct contact with hundreds of students, many of them geology majors, who are committed to the outdoors and to field experiences. However, very few of these students go to the park, even though they live in Houghton, MI, which is the home of the Ranger III, one of the principal transporters of visitors to and from Isle Royale.

Likewise, many of the geologists I have known have visited all of the geological sites around Lake Superior and the other Great Lakes, but only a few of them have been to Isle Royale. This is a remarkable contradiction, something I'm at a loss to explain. It seems to attest to America's addiction to the automobile; maybe people just can't stomach the thought of being separated from their car for a few days!

At any rate, I hope that this guidebook will encourage more geologists, as well as other people, to visit the park. Besides the fact that Isle Royale has outstanding geological sites, a trip there can be made at moderate expense, and the park offers comfortable facilities and logistics that most geologists would find agreeable. I recommend taking a week to visit (September is the very best time) and renting a motor boat from the park concession to allow access to the many wave-washed outcrops.

Introduction

This guidebook is meant to be used with The Geologic Story of Isle Royale National Park by N. King Huber (1983) and does not try to duplicate what occurs in that volume. Instead, the goal is to focus on the volcanic geology that can be observed in the wave-washed shorelines of the Rock Harbor area, emphasizing the physical volcanology of basaltic lava flows. The field trip described here offers the participant a chance to see excellent exposures of large and thick, tholeiitic lavas and to observe the combination of physical features that are associated with slow solidification and segregation.

Background and Key References

Huber's work offers a complete background to this trip. It includes a bedrock geological map (Huber, 1973) and United States Geological Survey (USGS) Professional Papers on the Portage Lake Volcanics (PLV) (Huber, 1973a), the Copper Harbor Conglomerate (CHC) (Wolff and Huber, 1973), and the glacial and postglacial geology (Huber, 1973b). Since 1973, when Huber's work was published by the USGS, there have been a number of further studies published that are important to the context of Isle Royale. Here, I will cite those that pertain to the Keweenaw section of the Midcontinent rift and to solidification of lavas in particular for those who wish to read more of the current literature.

A detailed study of the composition of the lavas of the PLV, as determined by study of a complete section on the Keweenaw Peninsula, was completed by Paces (1988). In this study, Paces provides a general description of the flows with respect to texture and thickness (Figure

1.1); petrography (Table 1.1); mineral chemistry (Figure 1.2); and chemical composition (Table 1.2), which is described in the following:

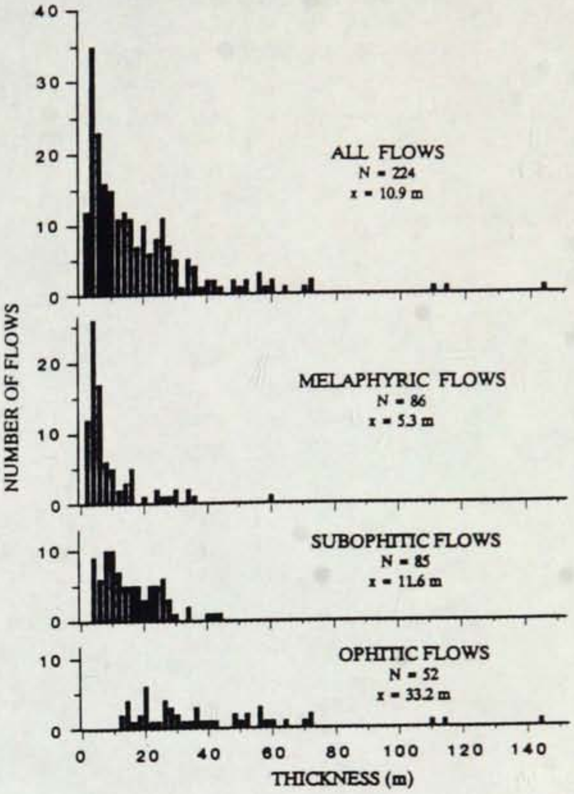


Figure 1.1: Histograms of PLV lava flow thickness (Paces 1988). N=number of lava flows; X=geometric mean thickness in meters. The coarsely-ophitic Greenstone flow is not shown. Textural categories are based on macroscopic observations.

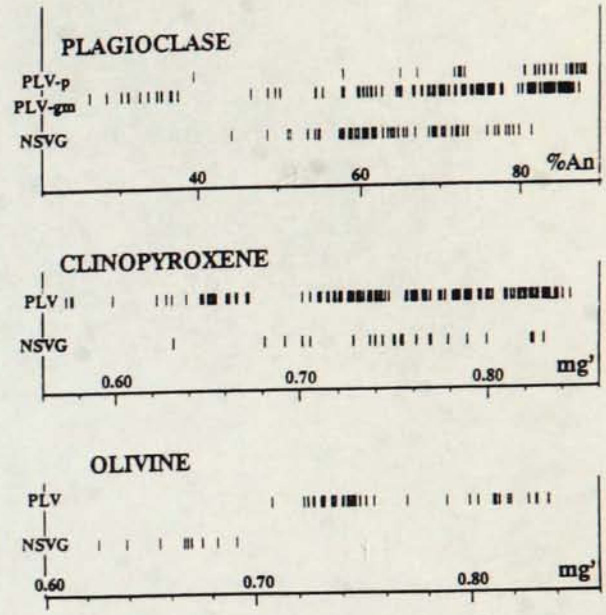


Figure 1.2: Compositions of minerals in PLV primitive olivine tholeiites compared to those available from North Shore Volcanic Group primitive olivine tholeiites (Paces 1988). Each line represents a single analysis. The North Shore Volcanic Group data was taken from BVSP (1981) and Brannon (1984). PLV plagioclase compositions are sub-divided into groundmass (PLV-gm) and phenocryst (PLV-p) crystals.

Table 1.1: Petrographic modes, in volume percent, of representative PLV olivine basalts and basaltic andesites (Paces 1988). Model analyses are based on 500-spot point counts on a 0.6 x 1.2 mm grid.

	Olivine Basalts N=8		Basaltic-andesites N=3	
	Typical	Range	Typical	Range
Groundmass				
Plagioclase	41	(39-56)	53	(50-55)
Clinopyroxene	23	(19-29)	18	(16-22)
Olivine	18	(11-19)	—	—
Fe-Ti Oxides	4	(2-13)	7	(6-10)
Glass	10	(4-11)	19	(15-25)
Vesicles	4	(1-9)	3	(1-10)
Phenocrysts				
Plag ± Olv	<1	(<1-9)	<1	<1

Table 1.2: Average major and trace element compositions for eight groups of PLV lavas (Paces 1988). Tholeiites were grouped by their Ni concentrations with criteria listed in the first row and the number of samples in each group in the second row. POT=primitive olivine tholeiite; OT1=olivine tholeiite 1; OT2=olivine tholeiite 2; IOT=intermediate olivine tholeiite; FOT=Fe-rich olivine tholeiite; AND=andesite; DAC=dacite; and RHY=rhyolite. The last three columns are individual analyses. Oxides are given in weight percent, trace elements in ppm.

	POT	OT1	OT2	IOT	FOT	AND	DAC	RHY
Ni(ppm)	400-300	300-250	250-200	200-100	100-15	D58-	83JP-7	85MH-1
	n=5	n=9	n=14	n=8	n=6	1297		
SiO_2	47.82	47.34	48.03	48.55	49.94	56.39	68.44	77.89
Al_2O_3	15.89	15.27	15.32	15.12	13.28	13.78	15.17	12.77
FeO_t	9.77	11.82	12.32	12.86	14.91	9.87	4.46	1.11
MgO	12.44	11.69	9.85	9.06	7.78	5.52	1.14	0.17
CaO	10.58	10.24	10.16	9.65	6.64	5.10	1.40	0.01
Na_2O	2.04	2.10	2.25	2.31	2.91	3.94	4.74	3.67
K_2O	0.19	0.22	0.33	0.42	1.43	2.27	3.86	4.28
TiO_2	0.98	1.13	1.35	1.60	2.34	1.83	0.51	0.08
P_2O_5	0.16	0.19	0.22	0.25	0.36	1.00	0.19	0.01
MnO	0.14	0.16	0.16	0.18	0.24	0.30	0.08	0.01
V	189	230	255	305	431	263	59	0
Cr	286	260	213	203	113	16	6	5
Mn	1090	1316	1250	1365	1804	2282	627	105
Ni	326	279	231	172	54	10	7	5
Cu	37	51	73	86	126	5	13	61
Zn	64	80	79	85	106	146	81	25
Rb	12	6	5	8	30	51	62	129
Sr	358	253	267	283	245	335	88	36
Y	20	20	20	23	31	47	45	35
Zr	78	85	101	126	212	430	573	145
Nb	4	6	7	8	14	28	55	68
Ba	108	154	200	265	514	757	879	1127
La	6.7	8.2	10.5	13.4	25.8	48.6	54.4	39.9
Ce	15.0	18.4	23.7	30.6	58.7	117.2	139.0	96.2
Sm	2.49	3.13	3.74	4.50	7.91	13.78	14.00	8.57
Eu	0.96	1.14	1.33	1.51	2.12	3.84	3.44	0.95
Tb	0.46	0.61	0.71	0.88	1.58	2.51	2.77	1.97
Yb	1.47	1.79	2.00	2.46	4.43	6.94	9.41	6.70
Lu	0.21	0.27	0.30	0.38	0.64	1.04	1.39	1.00
Ta	0.43	0.48	0.55	0.66	1.14	1.77	2.64	3.52
Hf	1.74	2.09	2.57	3.11	5.72	10.44	16.13	7.29
Th	0.55	0.77	0.97	1.31	3.39	5.94	10.32	16.30
Sc	24.86	28.19	28.25	30.62	34.61	21.06	11.43	2.73

PLV lava flows display a relatively limited number of textures based on the relationships between dominant mineralogical constituents. These components originally included groundmass plagioclase, olivine, clinopyroxene, iron-titanium oxide, volcanic glass or mesostasis, occasional phenocrysts or microphenocrysts of plagioclase, and sometimes olivine. Textures that developed within the coarsest portion of different lava flows range from fine-grained intergranular through subophitic and ophitic. This same range in textures can be observed in individual, thick lava flows which grade from intergranular chilled flow margins to a coarsely ophitic flow interior. True quench textures (Lofgren 1980), including skeletal, dendritic or spherulitic olivine and pyroxene, have not been observed in PLV basalts.

PLV lava flows do not preserve evidence of an extensive pre-eruptive crystallization history. Chilled margins are generally aphanite. Occasionally, lavas contain minor amounts (usually less than 1%) of small euhedral phenocrysts of plagioclase (often with melt inclusion-rich cores) and sometimes olivine. When present, both of these phases commonly exhibit glomeroporphyritic tendencies. Neither the plagioclase nor olivine phenocrysts show obvious evidence of disequilibrium with the liquid. Except for rounded plagioclase cores, both olivine and plagioclase phenocrysts are in apparent textural equilibrium with the liquid. Slightly porphyritic lavas frequently exhibit serrate textures.

The dominant textural element in all lavas is the framework of groundmass plagioclase laths. This framework is a randomly-oriented, felt-like structure of interlocking euhedral to subhedral laths. Rarely, the partial alignment of laths forms crude trachytic fabric, indicating movement of magma after at least partial crystallization.

The second most prominent textural element is defined by clinopyroxene crystals and their relationships to the plagioclase lath framework. In all cases, clinopyroxene has clearly crystallized later than olivine and plagioclase. Clinopyroxene crystals exhibit intergranular to ophitic textures depending both on the size of the clinopyroxene crystals as well as the size of the plagioclase laths. Melaphyric flows and chilled flow margins contain small, blocky clinopyroxene crystals intergranular to the plagioclase framework. In many (but not all) thicker flows, clinopyroxene grains begin to enclose subophitically, and eventually ophitically, plagioclase and olivine crystals as the massive flow interior is approached. The boundary between subophitic and ophitic textures is gradational and is exceeded when a significant number of plagioclase laths are completely enclosed by the surrounding clinopyroxene oikocrysts. Absolute size of the oikocryst is not definitive: a large clinopyroxene grain may only subophitically enclose large groundmass plagioclase laths, however the same sized grain may ophitically enclose plagioclase laths of smaller dimensions.

Thus, over half, 60-70% (volume basis), of most PLV lava flows are typically composed of a plagioclase lath framework with loosely packed clinopyroxene oikocrysts. The remaining interstitial space within the plagioclase framework and between oikocrysts is filled with variable proportions of intergranular olivine, iron-titanium oxides, and intersertal volcanic "glass." Evidence of gas exsolution is preserved in some flow interiors as vesicular cavities of ellipsoidal to highly irregular shapes. Diktytaxitic textures, however, are not apparent. Vesicles are particularly well preserved in thinner flows which quenched rapidly; however, they are observable in some thicker flow interiors as well.

--Paces 1988

Two conclusions emerge from Paces work: (1) the lavas are compositionally similar throughout the section and generally are high magnesium, olivine tholeiites and (2) the flows range from less than 10 m (33 ft.) to more than 100 m (330 ft.) thick, and the thicker ones are more likely to have ophitic textures (Fig. 1.3).

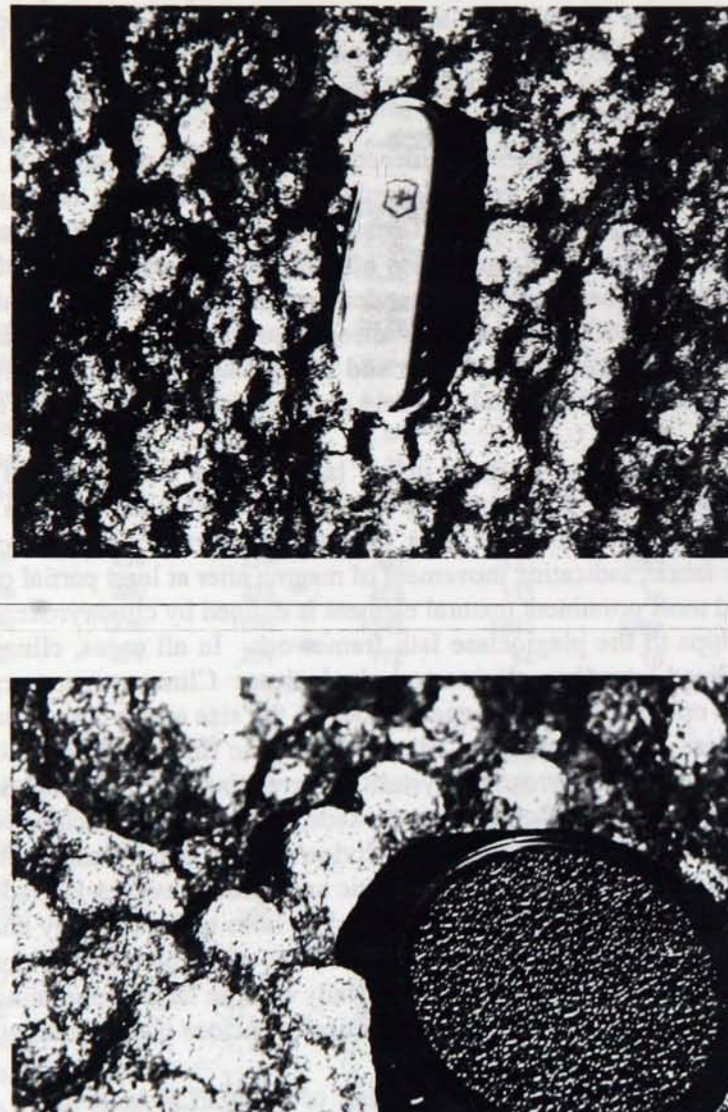


Figure 1.3: Ophitic texture in Isle Royale lavas. Weathered outcrops are especially advantageous to observe the ophitic texture because pyroxenes develop positive form. The scale of oikocrysts is typically between 1 and 5 cm, with observable changes in size sometimes observable on individual outcrops.

Lane (1911) describes the first recognition of the mirror image geology and lithological similarity of the PLV and the CHC on both sides of the Lake Superior syncline (Fig. I.2) and further suggests that the great lava flow of the Keweenaw Peninsula and the large flow of Isle Royale are the same. Huber (1973a) strongly supports Lane's correlations. Longo (1984), after extensive field mapping and sampling at Isle Royale and the Keweenaw, gives field observations and geochemical data that strongly confirmed the correlation of the Greenstone flow. Interpretation of seismic reflection profiles across Lake Superior (Figs. 1.4 and 1.5) provide confirmation of the rift basin synclinal geometry and the correlation of Isle Royale stratigraphy with the Keweenaw Peninsula. This correlation means that the Greenstone flow is one of the earth's largest lava flows; according to Longo (1984), it has an aggregate volume of 1650 km^3 (396 mi^3), comparable to the Roza flow of the Columbia River Flood basalts, which is estimated to be 1500 km^3 (360 mi^3) by Swanson and others (1975). The areal extent of the Roza, 40,000 km^2 (15,450 mi^2), is much larger than the Greenstone flow, 5000 km^2 (1930 mi^2), a comparison which results from the ponding of the Greenstone within the rift basin. Thus, the solidification of the Greenstone flow is a kind of magma ocean experiment, the likes of which is rare on this planet.

Paces and Davis (1990) acquired U-Pb dates of zircons from the Greenstone flow and two others in the PLV sequence on the Keweenaw Peninsula (Fig. 1.6). Their data show that PLVwas largely erupted between 1096 and 1094 Ma, which corresponds to the main stage of volcanic activity (Fig. I.6) Based on the estimated areal extent of the PLV basin in the western Lake Superior area, magma eruption rates of 0.02- 0.05 km^3/y (Davis and Paces, 1990) and 0.2 km^3/y (Cannon, 1992) have been calculate. The rate estimated by Cannon (1992) is comparable to that calculated for many Phanerozoic flood basalt provinces.

Studies of jointing, internal structures, and mineral textural have led to better understanding of the conditions and rates of eruption and solidification. The formation of columns in lavas are particularly important to the appreciation of the Isle Royale occurrences. The recognition of the role of water infiltration in the formation of certain kinds of entablature jointing (Fig. 1.7) in the Columbia River flood basalts by Long and Wood (1986) was an important insight, as was the detailed work on column formation by DeGraff and Aydin (1993) and DeGraff and others (1989). The work by Goff (1977) provides information on vesicle cylinders in lavas for the Isle Royale occurrences. The processes that operate during the solidification of tabular magma bodies are reviewed by Marsh and others (1991). Worster and Huppert (1993) present models developed to predict the patterns of solidification in time and space (Figs. 1.8 and 1.9); fundamental work to understand solidification in other kinds of materials has been done and is presented by Hellawell and others (1993). Green's (1989) report on the physical volcanology of the North Shore Volcanics gives a valuable comparison with rocks associated closely in time and space with the Isle Royale PLV.

The works cited above, along with others referenced in this guide, contribute more information to our understanding of the geology of Isle Royale. All of these works are valuable as supplements to this field guide.

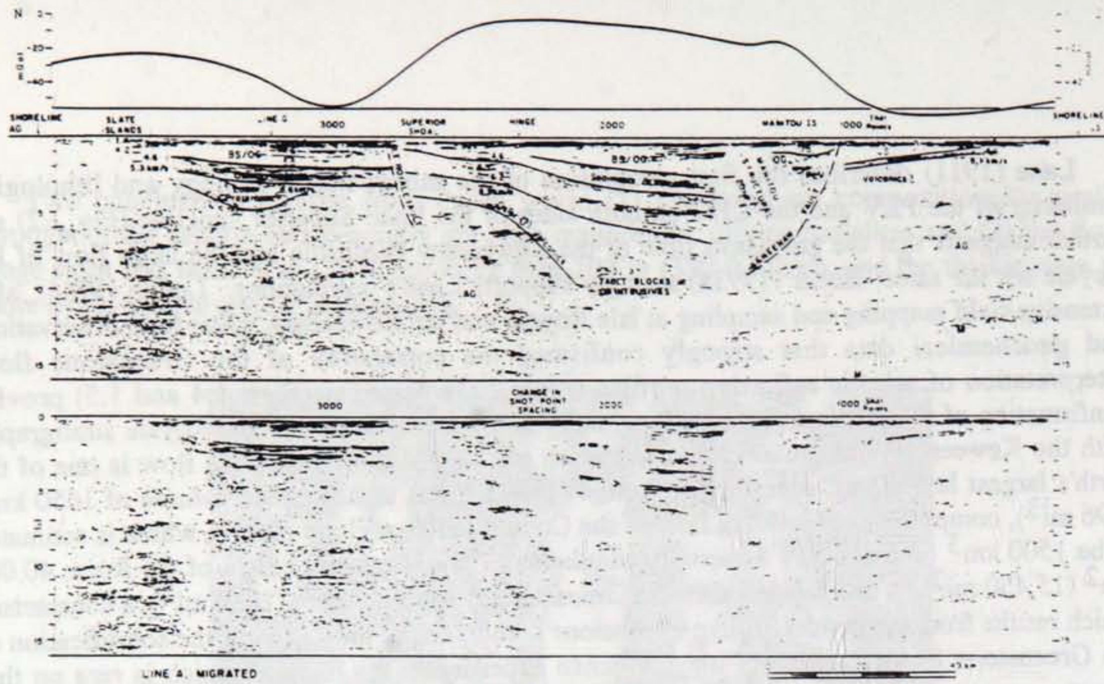


Figure 1.4: *Top:* Interpreted reflection profile along line A showing the subsurface geology beneath central Lake Superior (Cannon and others 1989). *Bottom:* Migrated seismic reflection profile along line A (Cannon and others 1989). Vertical scale is seconds of two-way travel time. Vertical exaggeration is 1:1 for average velocity of 6 km/s.

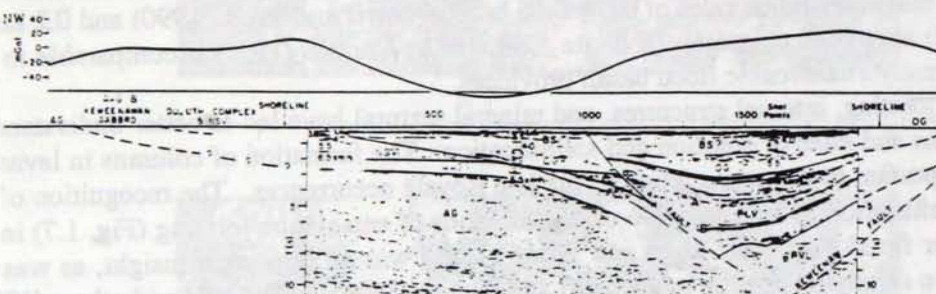


Figure 1.5: *Top:* Interpreted reflection profile along line C, showing subsurface structure beneath W Lake Superior (Cannon and others 1989). Inferred subsurface units are projected updip to their exposed extensions on land in N Michigan and Minnesota. *Bottom:* Migrated seismic reflection profile along line C (Cannon and others 1989). Velocity/depth profiles determined from refraction surveys were used to calculate depths to reflectors. M=approximate location of Moho; AG=Archean Gneiss; PLV=Portage Lake Volcanics; pPLV=pre-Portage Lake Volcanics; NSV=North Shore Volcanics Group; OG=Oronto Group; BS=Bayfield Group; and AnG=Animikie Group. Vertical scale is seconds of two-way travel time. Vertical exaggeration is 1:1 for average velocity of 6 km/s.

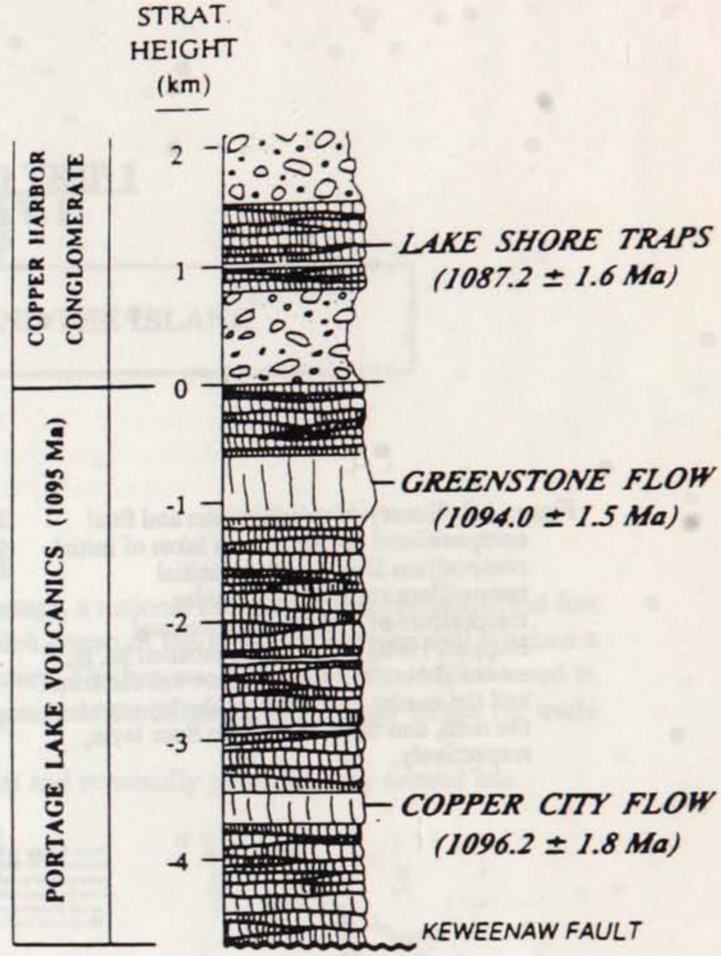


Figure 1.6: Schematic stratigraphic column for the Portage Lake Volcanics and part of the Copper Harbor Conglomerate exposed on the Keweenaw Peninsula showing U-Pb zircon ages (from Davis and Paces, 1990).

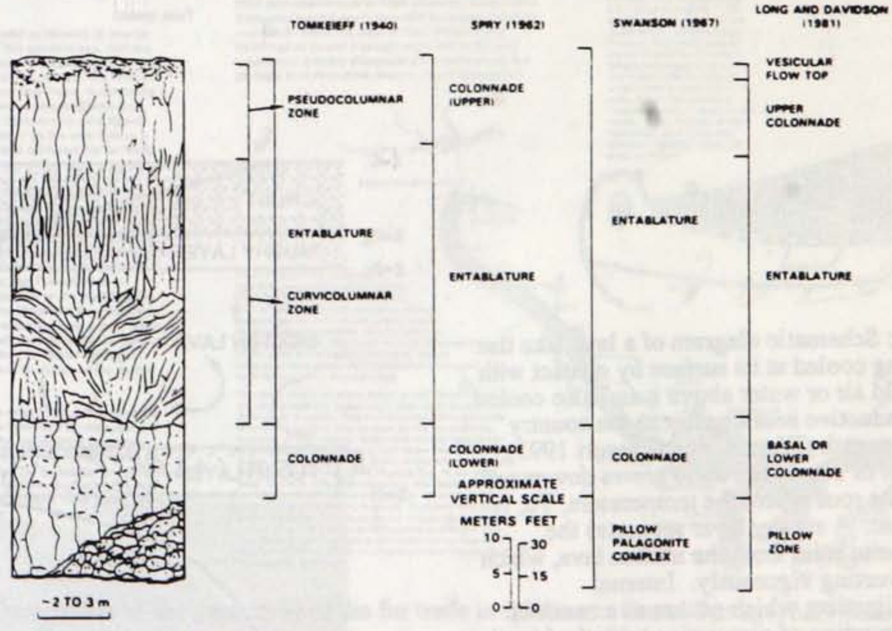


Figure 1.7: Typical intraflow structures present in Grande Ronde Basalt flows (Long and Wood 1986). Fractures are represented in a stylized manner, and fracture widths are not to scale.

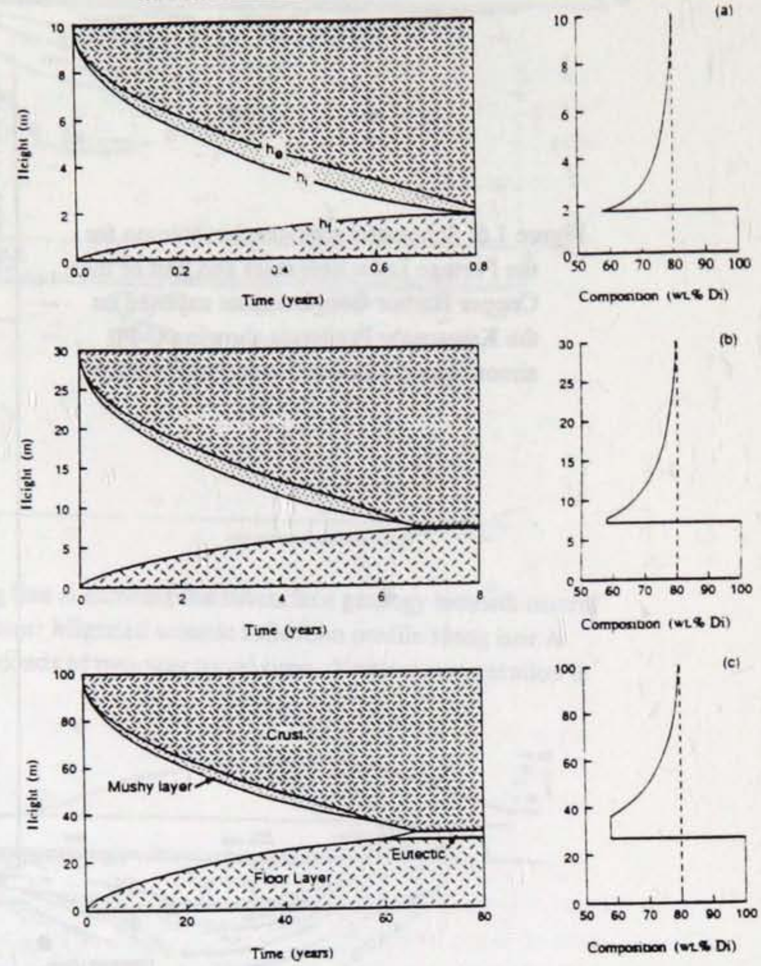


Figure 1.8: History of solidification and final compositional profile in lava lakes of initial composition $Di_{80}An_{20}$ and initial temperature equal to the liquidus temperature of 1352°C (Worster and Huppert 1993). The curves labelled he, hi, and hf show the interfaces between the crust and the mushy layer, the mushy layer and the melt, and the melt and the floor layer, respectively.

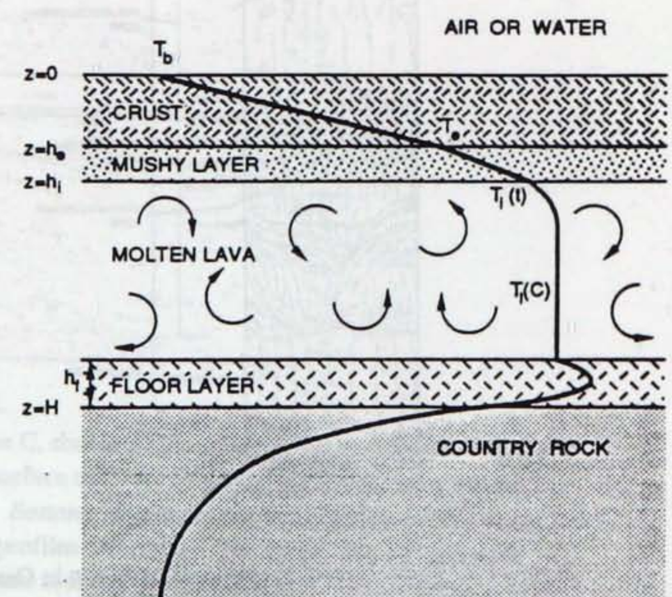


Figure 1.9: Schematic diagram of a lava lake that is being cooled at its surface by contact with the cold air or water above it and also cooled by conductive heat transfer to the country rock beneath (Worster and Huppert 1993). A crust of composite solid grows downward from the roof where the temperature, Tb, is constant. A mushy layer separates the composite solid from the molten lava, which is convecting vigorously. Internal crystallization, which occurs as a result of the interaction of convection with the kinetic undercooling at the interface with the mushy layer, forms a solid layer near the floor of the lake. A sketch of the temperature field is indicated on the diagram.

FIELD TRIP 1
DAY 1

A TRIP AROUND THE ISLAND

STOP 1-1: DEPARTURE

Location: Grand Portage, Minnesota

Duration: brief

Description: Our trip begins at Grand Portage, a national monument and reconstructed fort that commemorates the fur trade era, which began in the sixteenth century and reached a peak in the later part of the eighteenth century. The fort was built in 1778 and abandoned in 1803. Figure 1.10, from the National Monument, describes the geography of the fur trade and the significance of this site.

From Grand Portage we will travel east and eventually go all the way around Isle Royale in a clockwise fashion (Fig. 1.11).

The Great Carrying Place

Only three water passages to the Northwest are scored into the broad rock face of the Laurentian Shield which forms the western shore of Lake Superior. They are the rivers known today as the St. Louis, the Kaministiquia, and the Pigeon. The last of these is navigable except for a few kilometers at its mouth, but a narrow, muddy trail links the lake with the navigable waters of the Pigeon. Indians had used this trail for ages before the first European explorer, a Frenchman, recorded it in 1722.

The French explorers, who continued to search for an easy passage to the western sea, and the French missionaries, who sought converts in the wilderness, gave the trail its name—"Le Grande Portage," The Great Carrying Place. It remained, however, for the Highland Scots and their partners in the North West Company to give Grand Portage its place in history as the vital link in a network of waterways that nurtured the fur trade empire.

The fur trade of North America developed in the 16th century between French fishermen and Indians along the banks of Newfoundland and the mainland coast. Furs gathered by the French for sale in Europe soon became the cash crop of New France. When fur-bearing animals, particularly beaver, became scarce, the French traders searched westward. Blocked to the north by the British fur domain of the Hudson's Bay Company, the French traders were forced over a long water and land route from Montreal through the Ottawa River and the Great Lakes to Grand Portage—gateway to the untapped fur riches of the Northwest.

Under British auspices after 1783, the route over the Grand Portage was inherited by independent traders who founded the North West Company. Theirs was a fur trade empire based primarily on the exportation of beaver fur to supply the particular tastes of European fashion. Beaver pelts, from which the finest quality felt hats were made, were in tremendous demand.

In July the post at Grand Portage was the scene of the North West Company's annual rendezvous. From the east came canoe loads of trade goods from the warehouses in Montreal destined for the interior. From the northwestern outposts came wintering traders with loads of furs en route to Montreal.

Though it was the North West Company's most important inland trading center, the Grand Portage post was short-lived; it was founded about 1778 and abandoned in 1803 when the North West Company moved north to avoid American taxation and established the post of Fort William. The buildings at Grand Portage were left to the elements and quickly disappeared. Traffic along the portage trail dwindled; then it, too, disappeared.

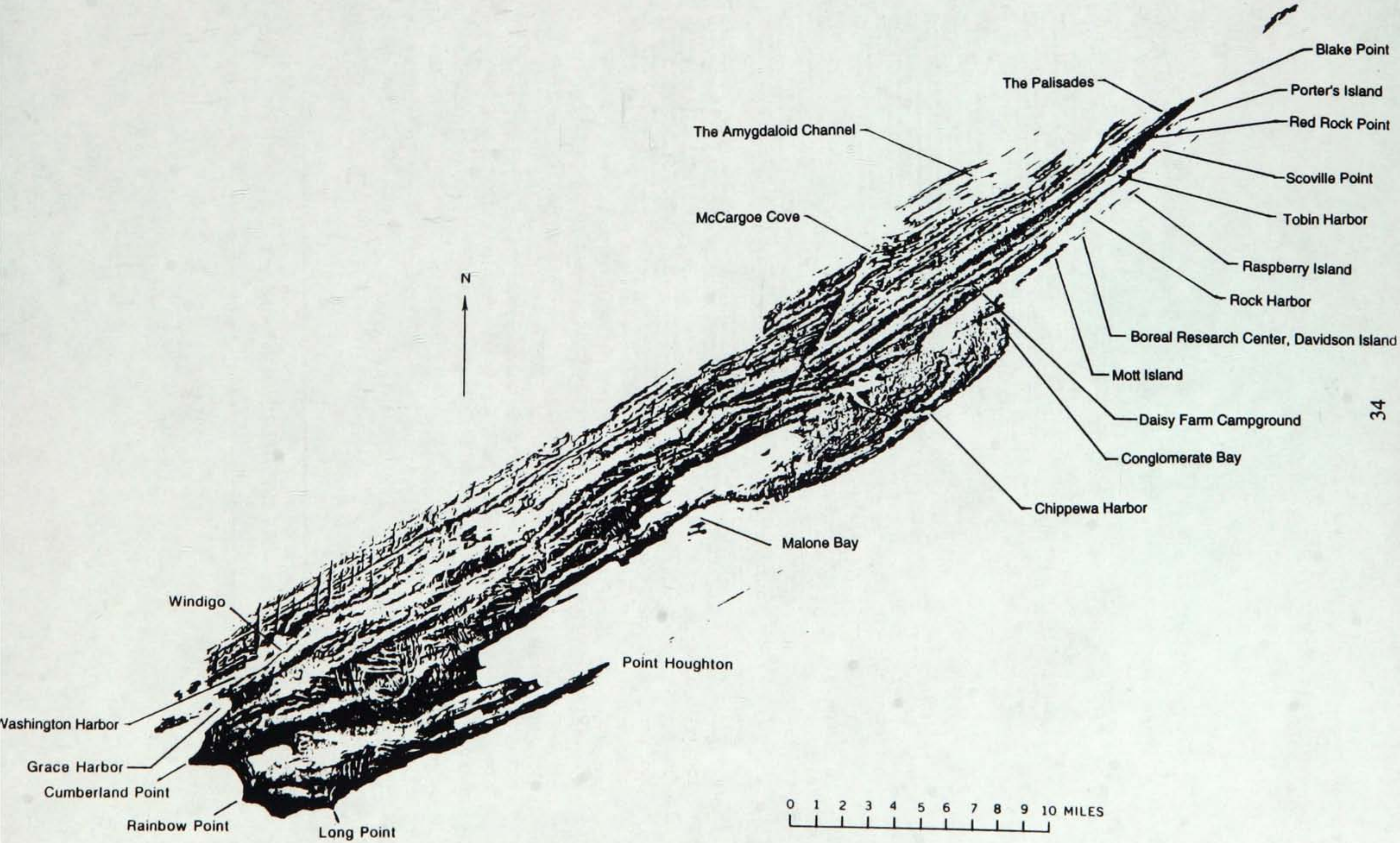
For at least 200 years the Indian birch bark canoe served as the main transport in the fur trade. The biggest ones, the "canot du maître," traveled the Great Lakes on the Montrealers' route to Grand Portage. They were usually about 11 meters (36 feet) in length and could hold up to 3.5 metric tons (4 tons) of cargo, plus a crew of 8 to 16 men. The north canoes or "canot du nord," used on the narrow water courses of the northwest were somewhat smaller, about 8 meters (26 feet) in length.

From Montreal the route of the fur trade swept westward for nearly 5,000 kilometers (about 3,000 miles) over a network of rivers and lakes, linked by portages in those places where travel by canoe was impossible. The far northern terminus was Fort Chipewyan on Lake Athabaska. At the peak of the trade hundreds of tons of pelts and trade goods were paddled and portaged each season along this waterway.

The short season between break up and freeze-up of the ice spurred the voyageurs in their dash to Grand Portage. In 5 months they had to cross half a continent and return. In early May as soon as the ice broke up, brigades of lake canoes, laden with several tons of cargo, moved out from Lachine, just above Montreal at the mouth of the Ottawa River. Eight weeks and 36 portages later they glided into Grand Portage Bay. In the northwest, crews of seasoned voyageurs, the "homme du nord," waited for the ice to break up, usually around May 15. Then Athabaska brigades, burdened with furs, began traveling down from the fur country to rendezvous in mid-July with traders at Rainy Lake and Grand Portage. Return journeys had to get underway by August 1 to allow northmen to be in Chipewyan and voyageurs in Montreal by October.

Figure 1.10: Description of the geography of the fur trade in relation to Grand Portage, MN (Grand Portage National Monument brochure, 1986).

Figure 1.11: Shaded relief map of Isle Royale National Park, MI (Huber, 1973b). The map was modified to include locations that will be visited during this field trip.



STOP 1-2: THE ISLE ROYALE FAULT

Location: East of Grand Portage, west of Rock of Ages (on Lake Superior)

Duration: 2 hours

Description: Between the Minnesota shoreline and Isle Royale, the strike of Keweenawan rocks, known as the North Shore Volcanics in Minnesota, changes from E-W to about N 55 E, where the PLV formation at Isle Royale begins. This discontinuity could be partially related to the Isle Royale Fault (Fig. 1.2), which the Voyageur crosses between Grand Portage and Isle Royale. This is a thrust fault, apparently associated with the inversion of the Midcontinent rift and was detected in the GLIMPCE profile collected on a N-S line east of the tip of the Keweenaw Peninsula (Fig. 1.4). It is thought to extend west to at least the southwest end of Isle Royale, where Isle Royale is mantled with a much thicker portion of glacial till and the glacial features are much more prominent (Huber, 1983; see pp. 20-21 and 41-54).

STOP 1-3: WASHINGTON HARBOR

Location: Windigo 7.5' quadrangle (T63N, R39W, Sec. 1-3)

Duration: 15 minutes

Description: From Grand Portage, we will take the Voyageur east to Washington Harbor and Windigo, about 35 km (22 mi) offshore. A schematic columnar section of the PLV on Isle Royale is shown in Figure 1.12. The bedrock geology of the Washington and Grace Harbor areas includes four large flows that continue all the way to the other end of the island. The Greenstone flow (pg) crosses the center of Washington Island and outcrops in several places south-southeast of Windigo. The Tobin Harbor flow (pth) outcrops at South Rock, southwest of Washington Island. The Minong flow (pm) outcrops south of McGinty Cove, and the Scoville Point flow (psp) outcrops near Middle Point on the south side of Grace Harbor. Figure 1.13 shows the variations in thickness of the PLV.

The thickest flows in this area are the Washington Island flow (pwi) and the Grace Island flow (pgi). Both of these flows occur only locally, from the end of Washington Island to a point between Windigo and Sugar Mountain, a distance of about 14.5 km (9 mi) along strike. The lava flows here dip at 15-20~SE, an attitude that is similar almost everywhere on Isle Royale. Vertical N-S trending fractures, with little offset, cut across the bedrock strata near Washington Harbor (Fig. 1.11). Huber (1983) interprets these as structures related to the warping of the Lake Superior Syncline. South of Grace Harbor, the bedrock of the island is buried by till (Fig. 1.34).

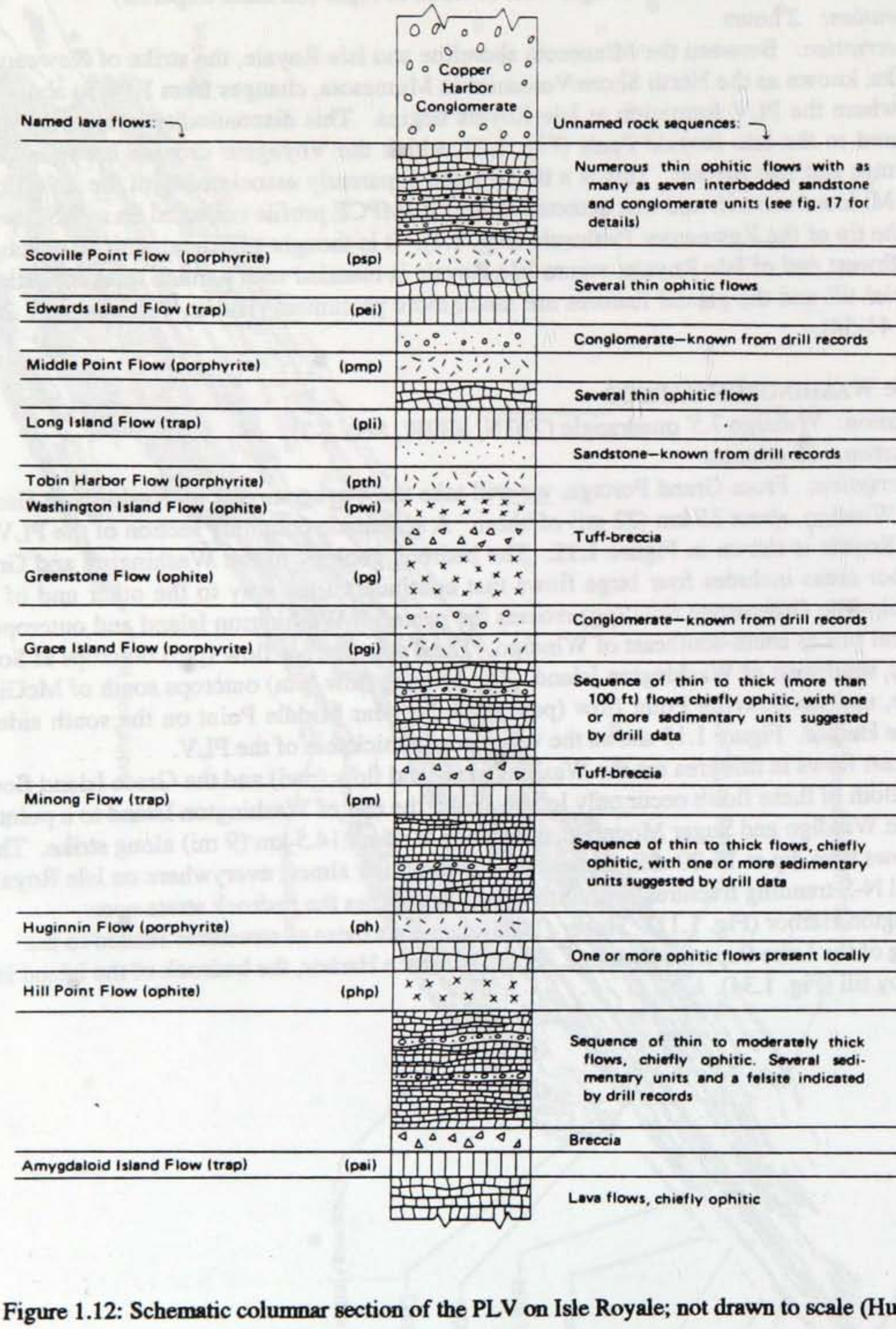


Figure 1.12: Schematic columnar section of the PLV on Isle Royale; not drawn to scale (Huber 1973).

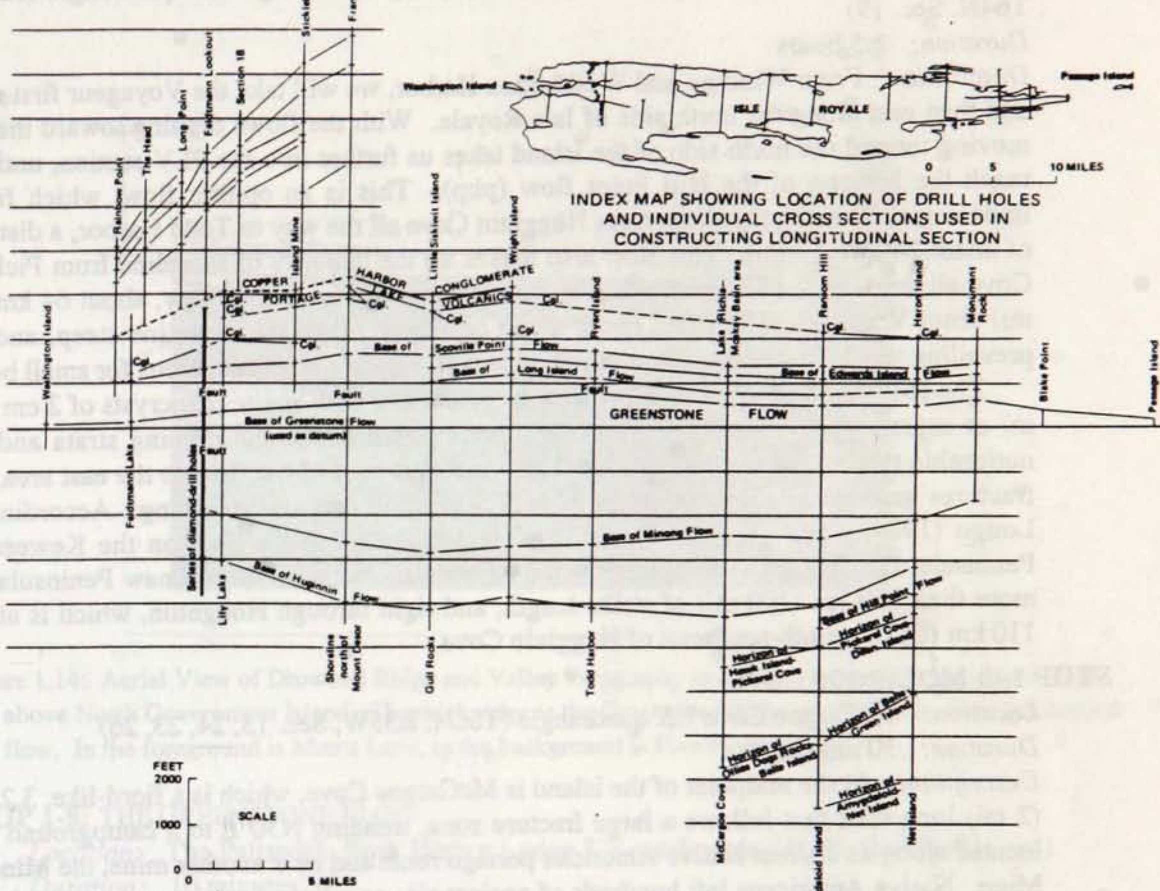


Figure 1.13: Longitudinal stratigraphic section showing variations in thickness of the PLV on Isle Royale (Huber 1973a).

STOP 1-4: WINDIGO

Location: Windigo 7.5' quadrangle (T64N, R38W, Sec. 29)

Duration: 30 minutes

Description: The Windigo area (Fig. 1.11) was the site of the last serious mining on Isle Royale, from 1890 to 1892. After failure and closure of mines further east, the Wendigo Copper Company (renamed from the Isle Royale Land Corporation) founded a mining venture on 8000 acres of land at Washington Harbor, under the leadership of Jacob Houghton, who was the brother of Douglass Houghton. The town site was named Ghyllbank and was located near the present site known as Windigo. The mine site, about 2 km (1.25 mi) inland to the northeast, was named Wendigo. People built roads all around the west end of Isle Royale, and 135 people lived at the mine site. The company did diamond drill exploration, as well as extensive trenching. In 1892, the miners gave up and left. When mining stopped, the company tried to sell land to tourists and resort owners (Rakestraw, 1965).

STOP 1-5: THE HILL POINT FLOW

Location: The north side of Isle Royale, beginning in Windigo 7.5' quadrangle (R38W, T64N, Sec. 19)

Duration: 1.5 hours

Description: From Windigo and Washington Harbor, we will take the Voyageur first north and then east along the north side of Isle Royale. With the flows dipping toward the SE, moving toward the north side of the island takes us further into the PLV section, until we reach the horizon of the Hill Point flow (php). This is an ophitic flow, which forms imposing cliffs along the shore from Hugginin Cove all the way to Todd Harbor, a distance of about 24 km (15 mi). This flow also makes up the majority of shoreline from Pickerel Cove all the way to Hill Point itself, at the west end of Five Finger Bay, about 64 km (40 mi) from Windigo. The tilted strata along the shore make the shoreline steep, and the prevailing winds from the north-northwest can make conditions treacherous for small boats.

The Hill Point flow is a coarse-grained, ophitic unit with augite oikocrysts of 2 cm (0.8 in) or more. The vertical fractures are superimposed across the dipping strata and are noticeable throughout the entire flow. From the west area of the flow to the east area, the fractures gradually begin to change from N-S to more northeast-trending. According to Longo (1984), the Hill Point flow may correlate with a large flow on the Keweenaw Peninsula, the Scales Creek ophite, which extends all along the Keweenaw Peninsula for more than 160 km (100 mi) of strike length, and right through Houghton, which is about 110 km (68 mi) south-southeast of Hugginin Cove.

STOP 1-6: MCCARGOE COVE

Location: McCargoe Cove 7.5' quadrangle (T66N, R35W, Sec. 13, 24, 23, 26)

Duration: 30 minutes

Description: At the midpoint of the island is McCargoe Cove, which is a fiord-like, 3.2 km (2 mi) long inlet that follows a large fracture zone, trending N30°E to a campground site located along an ancient Native American portage route and near another mine, the Minong Mine. Native Americans left hundreds of ancient pits as relics of mining over centuries at this site, and in 1874 three companies were formed in Detroit to exploit the potential here. They built a dock and a warehouse and started to build a railroad. Some large masses of copper were successfully mined, and the community here grew for several years in spite of difficult winter conditions. But mining did not last beyond 1885 (Rakestraw, 1965).

STOP 1-7: THE AMYGDALOID CHANNEL

Location: McCargoe Cove and Belle Harbor 7.5' quadrangle (T66N, R34W, Sec. 4,5, 6)

Duration: 30 minutes

Description: From McCargoe Cove, we will continue to the northeast, passing through the Amygdaloid Channel (Fig. 1.11). Amygdaloid Island is composed of the oldest lavas of the PLV on Isle Royale and is supported by a large flow, the Amygdaloid Island flow (pai), which is a fine-grained basalt (termed "trap"). At the west end of Amygdaloid Island is the National Park Service (NPS) ranger station near Kjaringa Kjeft. Crystal Cove is 3.2 km (2 mi) east of the station, which was, beginning in 1906, a private residence and fishery.

As we travel through the Amygdaloid Channel, the drowned ridge and valley topography of Isle Royale will become very visible (Fig. 1.14), with more resistent lava flows holding up linear islands. Shipwrecks are numerous on the many "reefs" found all around the northeast end of Isle Royale. Opposite of Crystal Cove on the south side of Amygdaloid Island is Belle Isle, which is a beautiful campground, accessible only by boat and canoe, located on the site of a resort that operated in the 1920s, serving the grand lake steamers of that period.

Figure 1.14: Aerial View of Drowned Ridge and Valley topography at Isle Royale, seen looking west from above North Government Island. The thick ridge is the Greenstone Ridge, underlain by the Greenstone flow. In the foreground is Merrit Lane, in the background is Five Finger Bay.

STOP 1-8: THE GREENSTONE FLOW

Location: The Palisades, Rock Harbor Lodge 7.5' quadrangle (T67N, R33W, Sec. 24)

Duration: 10 minutes

Description: From this point, we will head toward the tip of Isle Royale to Blake Point (Fig. 1.15), moving up in the stratigraphic succession. We will first cross the Hill Point flow (php) at Hill Point, then the Minong flow (pm) near Locke Point, and finally the Greenstone flow (pg) at the Palisades. The Greenstone flow is perhaps the earth's largest lava flow. The following are comments by Longo (1984):

> Similarities in the stratigraphic sequence of Isle Royale and the Keweenaw Peninsula of Michigan were recognized by numerous workers prior to 1851. The first thorough study of both areas, conducted by Lane (1893, 1911), resulted in the correlations of specific rock units. One unit in particular, due to its persistence as a prominent ridge on both Isle Royale and the Keweenaw Peninsula, became Lane's most convincing evidence for a correlation across this section of the Lake Superior syncline. Lane (1893) states, "The backbone ridge thus agrees in every way with the great corresponding ridge on the Keweenaw Point." Outcrop and drill core data by Lane (1893) reveal this unit as a single immense, differentiated lava flow. Lane (1893) refers to the flow as "the Greenstone, the `backbone' and biggest ophite of all, with the bed at its base we correlate as the Allouez Conglomerate."

The Greenstone's great thickness and differentiated nature led some workers to consider it as an intrusive sill (Seaman and Seaman 1944; Van Hise and Leith 1911). However, convincing data have proven this unit to be a lava flow (Lane 1893, 1911; Butler and Burbank 1929; Broderick 1935, 1946; Cornwall 1951), and henceforth known as the Greenstone flow. Huber (1973a) confirms the similarities of the

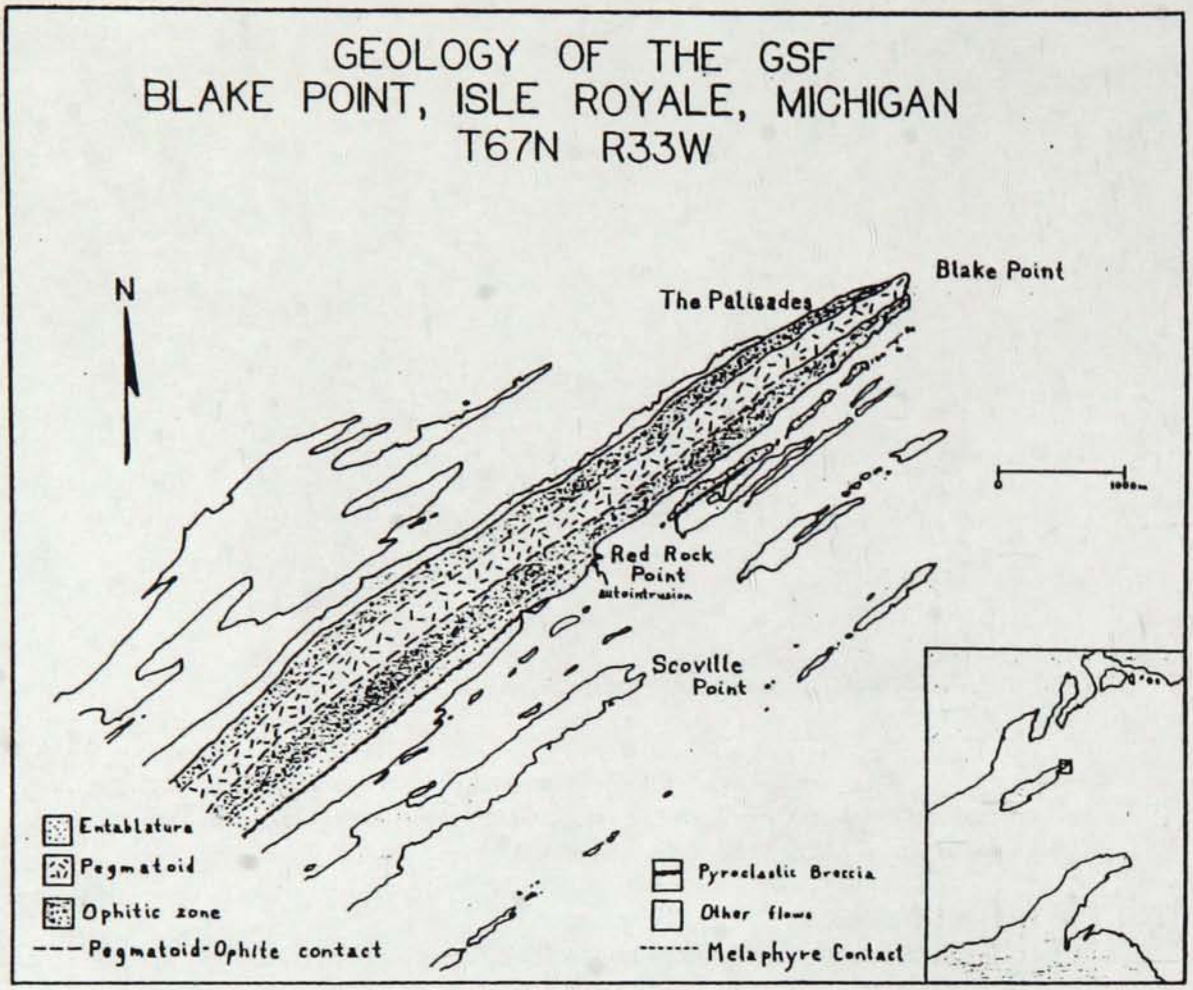


Figure 1.15: The geology of the Greenstone flow at Blake Point on Isle Royale (Longo 1984).

Figure 1.16: Columns in the Greenstone Flow at the Palisades, seen from the Northwest. The columns here are up to 5 m across.

Greenstone flow on Isle Royale and the Keweenaw Peninsula, and he supported the correlation.

--Longo 1984

The columnar character of the exposure seen at the Palisades (Fig. 1.16) is interpreted by Longo (1984) as a colonnade, following the terminology of Tomkeieff (1940). The cliffs are formed in the lower part of the great lava flow, which is 122 m (400 ft.) thick at the Palisades. The unit that comprises the cliffs is called the lower ophite and is about 30 m (98 ft.) thick. The rude columns are typically a few meters in diameter, and the rock is a coarse ophite with large pyroxenes enclosing many tiny plagioclases. The pyroxene oikocrysts increase in size from the bottom until they reach about 1.5 cm (0.6 in) in diameter near the top of the lower ophite. The lower ophite commonly forms steep anti-dip slopes on the north side of the Greenstone ridge (Huber, 1973a). We will return to the Palisades and will also walk around to Blake Point later in the trip to look at this exposure more carefully.

STOP 1-9: BLAKE POINT

Location: Rock Harbor Lodge 7.5' quadrangle (T67N, R33W, Sec. 24)

Duration: 10 minutes

Description: As we take the Voyageur around Blake Point, we will see that the layers of the Greenstone flow are clearly delineated (Fig. 1.15). Under the navigation light at the water line is the contact between the lower ophite and the pegmatoid (or pegmatitic) zone, about 23 m (75 ft.) thick. The pegmatoid represents the flow interior and has a texture that is strikingly different from the ophite, consisting of a mat of interlocking plagioclase laths as long as 2 cm (0.8 in) each. The pegmatoid is a bit less resistant to weathering, causing an indentation of the shoreline south of the light that marks the pegmatoid zone.

Farther S, a cliff face with columns marks the beginning of a third layer of the Greenstone flow, the upper ophite. Like the Palisades, this part is also marked by colonnade columnar jointing. We will have an opportunity to examine this location on the ground later in the trip.

The uppermost parts of the Greenstone flow consist of an entablature jointed part, which is locally prominent, and a fragmental flow top with abundant agates. The flow top is not well exposed because it erodes much more readily than the interior.

Around the point, we will pass between Third Island and North Government Island. On the north side of Edwards Island, exposures of entablature style columnar jointing are spectacular in the Edwards Island flow (pei). North of Edwards Island are exposures of the Long Island flow (pli), recognized by characteristic blue agate amygdules. Both of these flows can be traced along Tobin Harbor and through scattered outcrops as far as 50 km (31 mi) to a point near Hay Bay (Huber 1973).

STOP 1-10: ROCK HARBOR

Location: Rock Harbor Lodge 7.5' quadrangle (T66N, R33W, Sec. 3, 4)

Duration: evening and all night

Description: Heading southwest to Rock Harbor, we will pass Scoville Point, which is the site of the Scoville Point flow (psp), a porphyritic basalt with fine, equant, millimeter-sized plagioclases. This flow extends all the way across Isle Royale and is very resistant to erosion, forming shoreline exposures and topographic highs. We will see this flow on numerous occasions throughout the next several days. The exposure shows the asymmetrical nature of most of Isle Royale, with pronounced dipslopes and antidip slopes (Fig. 1.17).

Rock Harbor is the National Park's main visitor center and the location of the only hotel on Isle Royale, Rock Harbor Lodge. Lodging will be in the housekeeping cabins at the lodge.

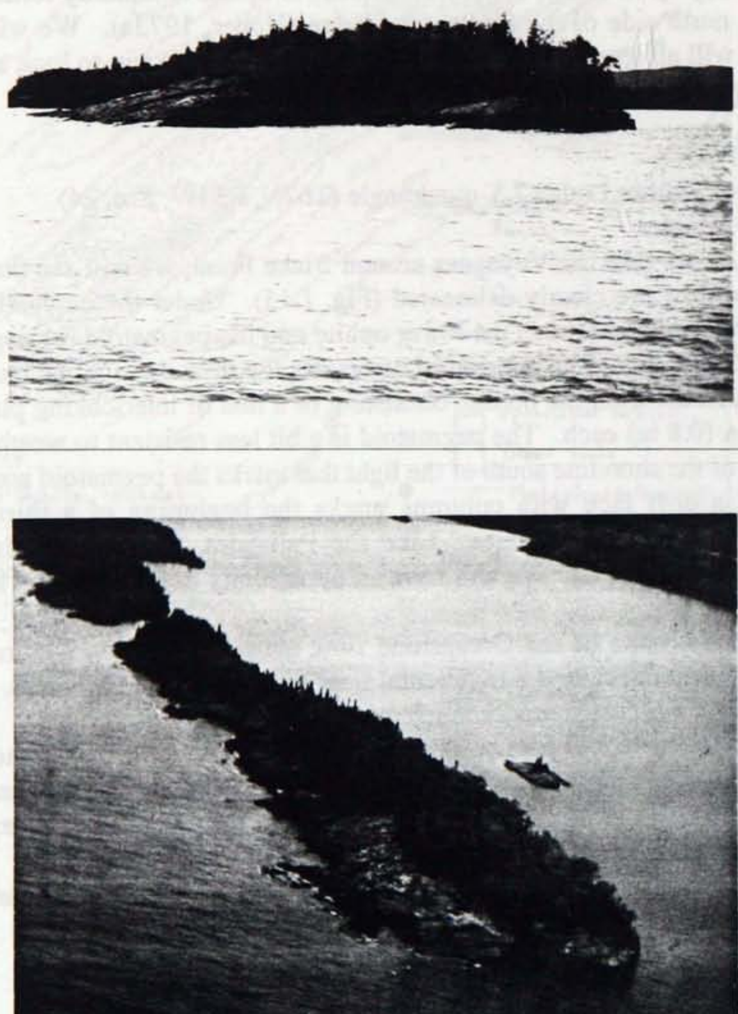


Figure 1.17: Views of asymmetrical linear islands near Rock Harbor. a) Smithwick Island seen looking southwest, with the dipslope to the left, and b) aerial view of South Government I, seen looking southwest, dip slope to the left.

FIELD TRIP 1
DAY 2

AN OVERVIEW DAY

STOP 2-1: MOUNT FRANKLIN

Location: Belle Harbor 7.5' quadrangle (T66N, R34W, Sec. 12)

Duration: 1.5 hours

Description: The day begins with a morning hike. We will take the Mount Franklin Trail, which begins 0.3 km (0.2 mi) west of Three Mile Campground, which we will reach by motor boat along Rock Harbor. The trail immediately climbs a ridge supported by the Scoville Point flow (psp), then levels off and descends. We will cross a boardwalk over a swamp and arrive at a valley where there is a junction with the Tobin Harbor Trail, 0.8 km (0.5 mi) from Three Mile Campground. We will continue on the Mount Franklin Trail, straight ahead, crossing the Tobin Creek swamp and then climbing a ridge underlain by the Tobin Harbor flow (pth). From here we will descend to cross another swamp and then begin the 300 ft. ascent of the Greenstone ridge. The entire swamp and ascent is underlain by the great Greenstone flow (pg). At the top of the ridge there is a junction with the Greenstone Ridge Trail, which we will take left to go about 0.5 km (0.3 mi) to Mount Franklin, elevation 330 m (1080 ft.).

Here there is a good view of the north side of the island, including Five Finger Bay, Lane Cove, and Amygdaloid Island, as well as of the Canadian Shoreline, including the Logan Sills and the Sleeping Giant. The Greenstone flow is indeed the backbone of the island, forming the most prominent ridge all along; only at Blake Point, however, is a reasonably complete section through the flow exposed. The contact between the pegmatoid and the lower ophite units of the Greenstone is mainly located near the crest of the Greenstone ridge. The lower ophite underlies the north slope, which is a steep, anti-dip slope, and the pegmatoid armors the gentler dip slope to the south.

STOP 2-2: THE MOUNT OJIBWAY TOWER

Location: Lake Ritchie 7.5' quadrangle (T66N, R34W, Sec. 15)

Duration: 1 hour

Description: From this point, still following the same trail, we will descend sharply to a wooded area and level off for about 0.4 km (0.25 mi) before climbing again. We will then reach the ridge crest and follow it for another 3.2 km (2 mi), with occasional outstanding views, to the Mount Ojibway tower. This structure was built in 1962 and was used initially as a fire tower. Now it is used for monitoring acid rain, along with other environmental monitoring. We can climb the tower stairs for full views of the surroundings, both to the north and south.

STOP 2-3: DAISY FARM CAMPGROUND

Location: Lake Ritchie 7.5' quadrangle (T66N, R34W, Sec. 22)

Duration: 1 hour

Description: From the tower we will descend to the Daisy Farm Campground via the Mount Ojibway Trail. We will go down from the ridge to the first level spot and then begin to rise over a smaller ridge. The beginning of this small ridge is the approximate location of the top of the Greenstone flow; the ridge top and the dip slope to the south is underlain by the Tobin Harbor flow (pth). At the base of this ridge we will cross a swamp fed by Tobin Creek. Then we will ascend Ransom Hill, which has the Long Island flow (pli) on its anti-dip (N) side and the Edwards Island flow (pei) on its dip slope (S) side, where there is some entablature jointing. From Ransom Hill, the trail descends to Daisy Farm Campground.

Daisy Farm is located on the site of an old mining community, called Ransom, which was founded in 1847 with the clearing of land and the construction of a smelter. The mining prospects dimmed quickly, however, and the mining activity ended only two years later in 1849. Then, in 1866 all the buildings burned down. In later years, the place was the site of a sawmill, a garden that supplied vegetables to Rock Harbor Lodge, and a Civilian Conservation Corps (CCC) camp, which was a foundation for youth employment, developed by Roosevelt during the depression (Rakestraw 1965).

STOP 2-4: EDISON'S FISHERY AND THE LIGHTHOUSE

Location: Lake Ritchie 7.5' quadrangle (T66N, R34W, Sec. 26)

Duration: 1.5 hours

Description: From Daisy Farm we will travel by boat across Rock Harbor to Edison's Fishery. The fishery itself is a restored camp that is occupied each summer by a retired Lake Superior fisherman and his family; this man is employed by the park to interpret what life was like here during the heyday of Isle Royale Fishing camps, from before the establishment of the park in 1936 until the sea lamprey invasion of the 1950s.

The lavas that underlie the site of the fishery and the lighthouse are a sequence of 45-50 ophitic flows (Fig. 1.18), which occur between the Scoville Point flow (psp) and the overlying CHC. As we walk around the point we will see several flow tops exposed, good examples of cellular amygdaloids. This is an excellent place to find greenstone, a nodular, compact form of pumpellyite that is prized as a semi-precious gemstone (Huber 1983, see pp. 58-59).

The geological purpose of stopping here is to look at the flow sections along the wave-washed shoreline, following it from this point to Tonkin Bay. We can also look at the amygdule mineral suite, which can be found on the pebble beaches. The amygdules of Isle

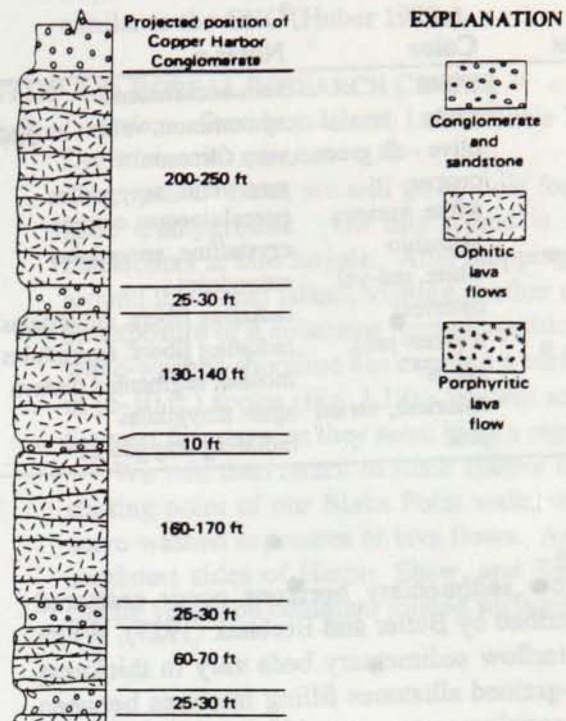


Figure 1.18: Columnar section of upper part of Portage Lake Volcanics showing distribution of sedimentary units in the Chippewa Harbor area (Huber, 1973). All the ophitic flow zones consist of multiple flows.

Royale's flows contain a variety of secondary minerals, listed alphabetically (by Huber) as barite, calcite, chlorite, copper, datolite, epidote, laumontite, natrolite, prehnite, pumpellyite (chlorastrolite or greenstone), quartz (agate), and thomsonite (Table 1.3). The prehnite is unusual in that it contains disseminated native copper inclusions and has a pink color, which has caused some to confuse it with thomsonite (Huber 1969). Overall, the assemblage is zeolite facies and prehnite-pumpellyite facies, representing a slightly lower grade than much of the Keweenaw Peninsula area. This may partially explain the lower abundances of native copper on Isle Royale than the amount found on the Keweenaw Peninsula.

STOP 2-5: MOTT ISLAND

Location: Lake Ritchie 7.5' quadrangle, NE corner (T66N, R34W, Sec. 24)

Duration: 1 hour

Description: From the lighthouse, we will return to the fishery and take the boat across the Middle Island Passage to Mott Island (about 3.2 km (2 mi) NE). We will stop here to visit one of the best exposures of sedimentary units within the PLV, found at the southwest end of the island, facing East Caribou Island near the park headquarters complex. There are seven such units mapped by Huber (1973) in the Chippewa Harbor area. Most of them are remarkably constant in thickness and lithology throughout their lateral extents, which are 65 km (40 mi) or more. Paces (1988) reports the following about interflow sediments in the PLV:

> Occasionally, lava flows are separated by intervening sheets and lenses of terrigenous clastic

Table 1.3: Characteristics of secondary minerals at Isle Royale.

Mineral	**Formula**	**Hardness**	**Color**	**Notes**
Barite	$BaSO_4$	2.5-3.5	white	vein occurrences
Calcite	$CaCO_3$	3	white	vy common, veins , amygdules
Chlorite	$(Mg,Fe)_5Al_2Si_3O_{10}(OH)_8$	2-2.5	olive - dk green	very common
Copper	Cu	2.5-3	copper	rare, veins, amygdules
Datolite	$HCaBSiO_5$	5-5.5	white, various	porcelaineous, opaque
Epidote	$Ca_2(Al,Fe)_3Si_3O_{12}(OH)$	6-7	pistachio	crystalline, amygdules
Laumontite	$CaAl_2Si_4O_{12}.4H_2O$	3.5-4	white, red-yel	amygdules
Natrolite	$Na_2Al_2Si_3O_{10}.2H_2O$	5-5.5	colorless	radiating fibers, amygdules
Prehnite	$Ca_2Al_2Si_3O_{10}(OH)_2$	6-6.5	lt green-pink	radiating fibers, amygdules
Pumpellyite	$Ca_2MgAl_2Si_3O_{11}(OH)_2.H_2O$	5.5	green	mosaic, segmented pattern
Quartz	SiO_2	7	colorless, varied	agate amygdules
Thomsonite	$NaCa_2Al_5Si_5O_{20}.6H_2O$	5-5.5	pink	radiating amygdules

sediment. Twenty two major interflow sedimentary horizons occur scattered throughout the PLV section and are described by Butler and Burbank (1929), White (1952), and Merk and Jirsa (1982). Interflow sedimentary beds vary in thickness from less than 1 cm (0.4 in) thick fine-grained siltstones filling fractures between flow top fragments to coarse boulder conglomerates over 100 m (330 ft.) thick locally. Typically, interflow sediments are poorly sorted, lithologically immature conglomerates and sandstones derived from a nearby volcanic source of some relief and deposited in an alluvial fan-type environment (Merk and Jirsa 1982). Transportation was generally from the southeast to northwest, or from basin margins towards the center of the subsiding graben (White 1952). Although the interflow sediments are volumetrically insignificant within the PLV (3% of the total lithologic volume) (Merk and Jirsa 1982; White 1971), they form distinct and relatively continuous stratigraphic marker horizons within an otherwise monotonous volcanic pile.

The occurrence of occasional interflow sediments implies that rates of lava flow extrusion, sedimentation, and/or tectonic subsidence were not constant during the formation of the PLV. White (1960) shows that a subsidence-depositional equilibrium was established so that both lava flows and sediments were deposited on near-horizontal surfaces. Most lava flows were deposited directly on top of the underlying lava flow top indicating a more-or-less constant and relatively short repose period between eruptions. The infrequent presence of sedimentary beds between lava flows may indicate occasional hiatuses in magma extrusion, which allowed for alluvial fans to transgress out towards the center of the basin. Conversely, interflow sedimentary horizons may mark brief periods of increased depositional rates possibly related to episodic normal faulting and basin subsidence.

--Paces 1988

The pebble conglomerate unit at Mott Island is thicker than most, about 12 m (40 ft.). The pebbles are essentially all of volcanic rock, with mafic varieties about twice as abundant as felsic ones. The sandstones are very feldspathic, plagioclase being most abundant. Fine-grained hematite is ubiquitous. The rocks have scour and fill structure, suggesting an origin in braided stream environments. Desiccation cracks and raindrop impressions have also been noted. Paleocurrent data suggest transport to the southeast, similar to the CHC (Huber 1973a).

STOP 2-6: BOREAL RESEARCH CENTER

Location: Davidson Island, Lake Ritchie 7.5' quadrangle (T66N, R33W, Sec. 8)

Duration: 30 minutes

Description: Next, we will go by boat for a brief stop at Davidson Island, opposite Three Mile Campground. On this island is the Boreal Research Center, a residence for researchers at Isle Royale. After stopping at the dock in front of the center, we will walk around this small island, visiting another exposure of the epiclastic sedimentary rocks and an exposure of a columnar jointed, ophitic flow on the southeast corner of the island. The wave-washed shoreline has exposed a surface perpendicular to the columns, which are 2-3 m (6-10 ft.) across (Fig. 1.19). We will see many exposures of large columns like these in the next few days, as they seem to be a regular feature of ophitic flows at Isle Royale.

We will then return to Rock Harbor by motor boat. On the way to Rock Harbor, the starting point of our Blake Point walk, we will pass several islands that have interesting wave-washed exposures of lava flows. Among the sites we may see from the boat are: the southeast sides of Heron, Shaw, and Smithwick Islands, where there are exposures of calcite veins and columnar jointed surfaces like the one we visited at Davidson Island.

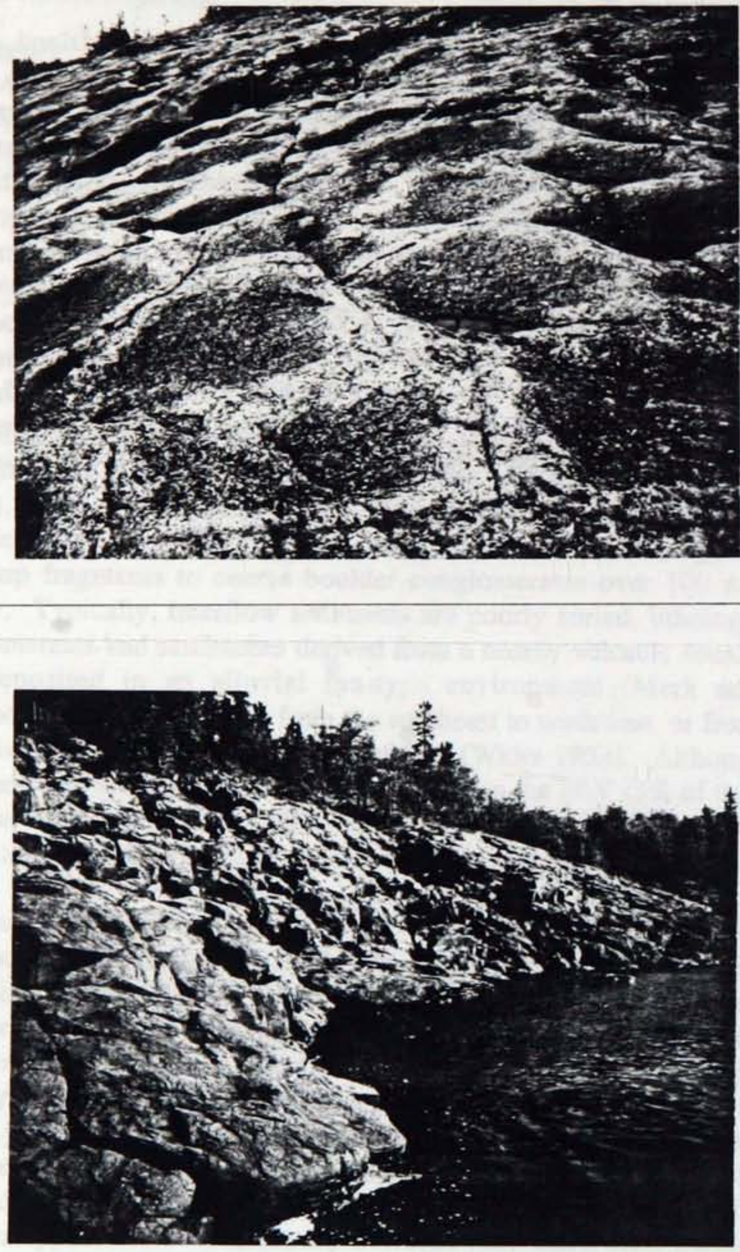


Figure 1.19: Columnar Joints in Ophitic Flows. a) Differential erosion perpendicular to the columns frequently depicts the columnar form by selectively eroding the center of columns, such as these on Smithwick Island. b) In other places the columns emerge as prisms, such as here on Raspberry I.

FIELD TRIP 1
DAY 3

A GOOD LOOK AT THE GREENSTONE FLOW

STOP 3-1: BLAKE POINT

Location: Rock Harbor Lodge 7.5' quadrangle (T67N, R33W, Sec. 24)

Duration: 4 hours

Description: Weather permitting, this is a day to visit one of the most dramatic places on Isle Royale, Blake Point. The shoreline around Blake Point offers the best view of the Greenstone flow, better than any other sites at Isle Royale or the Keweenaw Peninsula. The northwest side of Edwards Island has good exposures of entablature columnar joints in the Edwards Island flow (pei).

The boat will take us to the Merrit Lane Campground for our walk to Blake Point. We will follow the shoreline from Merrit Lane around the point, remaining close to the wave-washed rocks, yet trying to keep our feet dry. Most of the walk is on the upper ophite unit of the Greenstone flow. (The entablature part of the Greenstone and its flow top is underneath Merrit Lane, and we will see parts of this from the boat later). The upper ophite exhibits a rude columnar structure all along the walk, with the columns perpendicular to the bedding. The size of the oikocrysts increases from top to bottom.

After rounding the corner, we will cut through the bushes to descend a cliff that marks the lower anti-dip face of the upper ophite. At the base of this cliff, we will see wave-washed exposures of the pegmatoid, here about 23 m (75 ft.) thick. The contact here appears to be quite sharp, although Huber (1973a) says it is frequently gradational. The following description of the pegmatoid is from Longo (1984):

> Lacroix (1928, 1929) coins the term "pegmatitoide" to describe the coarse-grained zones considered to represent the final stages of differentiation in basaltic lavas of France. The lavas of Michigan's Copper Country show similar differentiates for which Lane (1893) applies the term "doleritic." Cornwall (1951) adopts the textural term "pegmatite" from the usage of Butler and Burbank (1929). He changed the confusing "doleritic" term to "pegmatitic facies," and subsequently described such units in the Greenstone flow, Big Trap, and several other large flows within the PLV on the Keweenaw Peninsula. For the present study, the term "pegmatoid zone" from Lindsley and others (1971) is adopted to encompass the portion of the Greenstone flow with numerous en echelon, lens-shaped pegmatoids, associated granophyric phases, and subophitic layers.
>
> Texturally, pegmatoids are coarse grained when compared to ophitic zones. Coarse plagioclase laths dominate with interstitial, subhedral clinopyroxene and abundant interstitial to somewhat poikilitic magnetite and ilmenite. Consequently, the

pegmatoids are strongly magnetic compared to ophitic units. This suggests that a higher titaniferous magnetite/ilmenite ratio for magmatoids than for ophites. Visual inspection generally reveals a greater overall opaque (oxide) concentration in the pegmatoids.

Subophitic layers are often found hosting the en echelon pegmatoids. These layers, like pegmatoids, are strongly magnetic and very coarse grained stratiform features, but contain less abundant, smaller sized pyroxene. The contacts between pegmatoids and subophitic units are usually sharp, although instances of gradational contacts have been observed. Subophitic layers grade into the ophites and seem to occupy the greatest volume of the pegmatoid zone. They have been observed to pinch out within pegmatoid units and may not be continuous planar features throughout the flow. Perhaps pegmatoid units are not only lens-shaped but also flattened amoeboid-like features interfingering with subophitic layers.

The frequency of pegmatoids and subophitic layers increases proportionally with increasing flow thicknesses. Both vary in thickness and shape and typically occur in the upper half of a lava flow. Pegmatoids have also been observed as autointrusions, such as in the entablature on Isle Royale and the upper ophite on the Keweenaw Peninsula. The stratiform pegmatoids are usually found armoring the tops of cliffs formed of the lower ophite. The extension of weak vertical joint patterns into the pegmatoid (forming crude large columns) suggests that pegmatoids may be part of the colonnade. In most cases pegmatoid zones separate a basal colonnade from an upper colonnade.

Pegmatoids are not unique to thick flows of the PLV. Lindsley and others (1971) assert that three of the thicker flows from the Picture Gorge Basalt contained pegmatoid lenses. Santin (1969) discusses the presence of pegmatoids in horizontal basalts of the Lanzarote and Fuerteventura Islands in the Canadian Archipelago.

--Longo (1984)

The pegmatoid underlies the low shoreline and also the area under the light tower. A section of the Greenstone flow is exposed on Passage Island, a 2 km (1.2 mi) long island that can be seen about 4 km (2.5 mi) offshore from Blake Point. Around the corner from the tower and vertically down about 4 m (13 ft.) is the contact with the lower ophite (which is too difficult for us to reach safely). Longo (1984) describes the contact as a gradation over about 1 m of thickness.

From here, we will return by the same route to Merrit Lane. Weather permitting, we will travel around the point in boats to examine the lower ophite cliffs along the Palisades. The columns exposed on the anti-dip slope are up to several meters across. The base of the Greenstone flow is not exposed here.

STOP 3-2: PORTER'S ISLAND

Location: Rock Harbor Lodge 7.5' quadrangle (T67N, R33W, Sec. 26)

Duration: 45 minutes

Description: Upon returning to Merrit Lane, we will head south to Porter's Island, which includes exposures of a fragmental rock that Huber (1973a) interprets as pyroclastic. However, according to Longo (1984), these exposures may represent a fragmental top of the flow. The breccia unit, which is about 1-5 m (3.3-16.4 ft.) thick, contains rounded and semi-rounded fragments of the Greenstone flow set in a finer matrix that has amoeboid-shaped, agate amygdules. Longo did an extensive perographic study but could not find any evidence of shards or pumice. He did, however, find bow-tie spherulitic plagioclases in the matrix, which suggests an undercooled texture for the basaltic material there. This unit occurs at the top of the Greenstone flow along about 15 km (9.3 mi) of strike length (approximately to Mt Ojibway), according to Huber's map. Similar units are found at the top of the Greenstone flow on the Keweenaw Peninsula (Longo, 1984). Figure 1.20 shows an old drawing of a cross section through Lake Superior, including Isle Royale and the Keweenaw peninsula.

STOP 3-3: RED ROCK POINT

Location: Rock Harbor Lodge 7.5' quadrangle (T67N, R33W)

Duration: 1 hour

Description: At Red Rock Point, we will pass excellent examples of entablature jointing of the upper part of the Greenstone flow. The basalt of the entablature is melaphyre, very fine grained (Fig. 1.21). The curvicolumnar nature of a few of the columns resembles some of the Columbia River basalt descriptions (Fig. 5). Long and Wood (1986) suggest that entablature jointing results when extensive floods that are created from deranged drainages cause dramatic quenches of solidifying flood basalts.

Around the corner of Red Rock Point is an autointrusion of the Greenstone's entablature zone by material that is apparently from the pegmatoid zone. Longo (1984) describes the feature:

> A large autointrusive dike was found intruding (north 20°W, 65°E) the columnar-jointed melaphyre at Red Rock Point. Despite an apparent lack of aplites, the dike is texturally similar to the stratiform pegmatoid. It is composed of randomly oriented, euhedral plagioclase laths with interstitial, subhedral augite and pigeonite (no poikilitic textures occur). The plagioclase laths are immense by comparison to the microlites of a typical ophitic unit.
>
> Three characteristic features of the dike are: (1) the abundant plagioclase phenocrysts (up to 1 cm (0.4 in)), (2) a blue-green hue from plagioclase altered to chlorite in the dike, and (3) alignment of plagioclase laths parallel to the dike contact, forming an igneous lamination. Amygdules are more abundant along the dike contact also.
>
> The process of autointrusion is similar to the mechanisms of pegmatoid formation, except that after the residual liquid is pressed out of the hosting crystal mesh, the differentiated magma is squeezed up into the vertical tensional fractures.
>
> --Longo 1984

Longo (1984) interprets the auto-intrusion to be related to a sag flowout structure, described by McKee and Stradling (1970) as: a large structure that develops as the crust of a partly solidified flow founders and causes the upward escape of the flow's fluid interior (Fig. 1.22).

Below the water level at Red Rock Point is an occurrence of coarse-grained granophyric rock, which can be found in beach cobbles and boulders and may occur within the Greenstone flow itself.

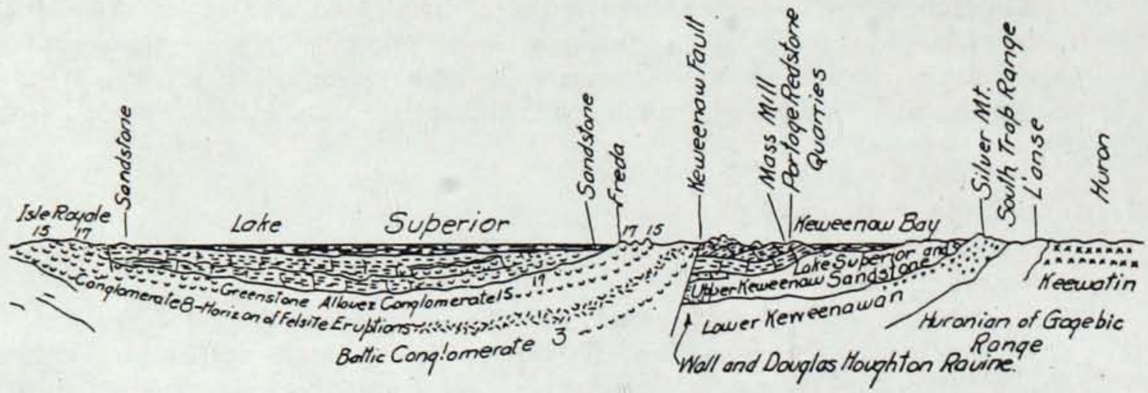


Figure 1.20: An older drawing of a cross section through Lake Superior, from the Huron Mountains to Port Arthur (Lane, 1911).

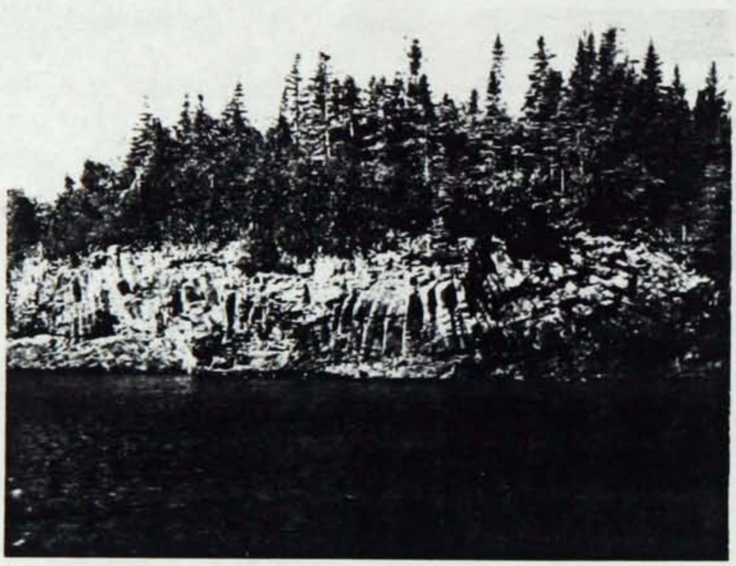


Figure 1.21: Entablature Jointing as seen in the upper portion of the Greenstone Flow at Red Rock Point.

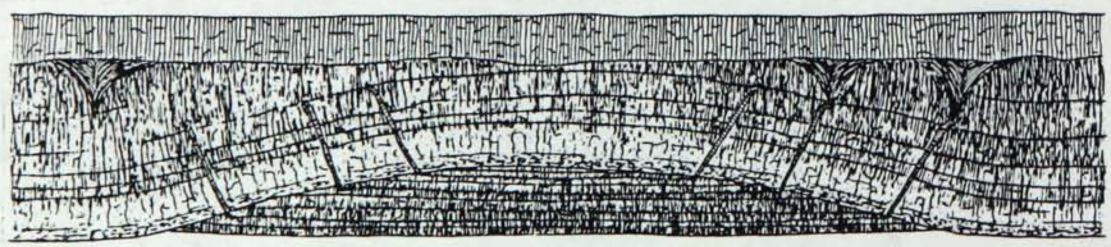


Figure 1.22: Idealized cross section of a sag flowout structure (McKee and Stradling, 1970). Following erosion, the dikes commonly remain as topographic highs in a circular pattern around the central sag area.

STOP 3-4: TOBIN HARBOR

Location: Rock Harbor Lodge 7.5' quadrangle (T67N, R33W, Secs. 26, 27, 34)

Duration: 5 minutes

Description: Continuing to the southwest along Tobin Harbor, we will pass Newman Island. Opposite this island on the north shoreline of Tobin Harbor are outcroppings of the volcanic breccia that may represent the fragmental top of the Greenstone flow. We will continue southwest to Hidden Lake dock.

STOP 3-5: LOOKOUT LOUISE

Location: Rock Harbor Lodge 7.5' quadrangle, NW corner (T67N, R33W, Sec. 34)

Duration: 1 hour

Description: From Hidden Lake to Lookout Louise, we will hike about 1.6 km (1 mi) long and 85 m (280 ft.) up. We will begin on the Tobin Harbor flow, but after passing the lake we will on the Greenstone flow, following a dip slope up to Lookout Louise. At about the halfway point, the trail passes Monument Rock (Fig. 1.23), an individual column from the colonnade of the upper ophite that is exposed as an erosional remnant. Huber (1983, see especially pp 47-55) suggests that Monument Rock was formed by wave cut shoreline processes along a former "raised" shoreline, which he associates with glacial Lake Minong, about 10 Ka. From Lookout Louise we will look over the steep anti-dip slope of the lower ophite and see Five Finger Bay, Duncan Narrows, and Amygdaloid Island.

The plan is to return from Lookout Louise to Hidden Lake and take the boat back to Rock Harbor for the night.

Figure 1.23: An old drawing of the northeast view of Monument Rock on Isle Royale (Foster and Whitney, 1851).

FIELD TRIP 1
DAY 4

SCOVILLE POINT

STOP 4-1: SCOVILLE POINT VIA THE STOLL TRAIL

Location: Rock Harbor Lodge 7.5' quadrangle (T67N, R33W, Secs. 26, 35)

Duration: 3.5 hours

Description: We will take the Stoll Trail, which goes along the shore of Rock Harbor. Along here, we will see Nipissing shorelines and outcrops of the Scoville Point flow. Also, we will be able to see the ophitic flows above and below the Scoville Point flow along the way. Huber (1973) describes the basalt of this flow as containing "fine, equant, millimeter sized, plagioclase crystals distributed uniformly through a fine grained matrix." He says the thickness is 30-60 m (100-200 ft.). There are not many features that can be seen in outcrop, but the flow is very resistant to erosion and buttresses the shoreline. About 0.8 km (0.5 mi) from the lodge lie ancient mine pits, attributed to Native Americans who occupied this area from about 2500 BC during the period of the Nipissing stage. The mining was apparently informal and quite limited in any one place, but there are more than 1000 such pits all over Isle Royale according to Rakestraw (1965).

As we near the Scoville Point, the Scoville Point flow dominates the shoreline and has steep smooth exposures. At the point itself, we will look at the excellent exposures of the Scoville Point flow, the ophitic flows below it, and the Edwards Island flow (pei), which underlies the companion point located just to the northwest of Scoville Point. There is a good exposure of cellular amygdaloid in one of the ophitic flows and the Edwards Island flow shows well developed entablature jointing, which may be viewed in an interesting variety of perspectives.

We will return to the lodge and to Three Mile Campground via the Tobin Harbor Trail, which is easier to hike. It stays near the shore of Tobin Harbor, mostly atop the Edwards Island flow. Just northeast of the Rock Harbor Lodge on the return trail is the site of the Smithwick Mine remains; this mine was discovered in 1843 and actually operated in 1847 and 1848. The work done here mostly consisted of exploratory shafts and excavations, and it is unclear whether or not much ore was found (Rakestraw, 1965).

STOP 4-2: SUZY'S CAVE

Location: Belle Harbor 7.5' quadrangle (T66N, R33W, Sec. 8)

Duration: 1 hour

Description: The trail to Suzy's Cave parallels the shore of Rock Harbor and lies on thin ophitic flows, which are stratigraphically below the Scoville Point flow. At Snug Harbor, the Scoville Point flow crops out only within the recessed part of the harbor, and the point where the Americas Dock is located is underlain by ophitic flows above the Scoville Point flow.

We will walk on the rocky dip slope of the Scoville Point flow (psp), facing Rock Harbor along the shore. At Suzy's Cave, about 1.8 km (1.1 mi) east of Three Mile Campground along the Nipissing shoreline, waves cut a sea arch at a contact between two lava flows about four thousand years ago, selectively eroding the region of the flow top.

STOP 4-3: PASSAGE ISLAND

Location: Passage Island 7.5' quadrangle (T67N, R32W, Secs. 3, 4, 9)

Duration: 4 hours

Description: Weather permitting, we may take a boat excursion to Passage Island using the park concessions boat. Passage Island is located 6 km northeast of Blake Point. It is underlain by the Greenstone Flow.

FIELD TRIP 1
DAY 5

RASPERRY AND EDWARDS ISLANDS

STOP 5-1: RASPBERRY ISLAND

Location: Rock Harbor Lodge 7.5' quadrangle (T66N, R33W, Sec. 3)

Duration: 5 hours

Description: We will go by boat to Raspberry Island, about 0.5 km (0.3 mi) southeast of Rock Harbor Lodge, and spend the entire day looking at a remarkable set of exposures nearby that provide an impression of some of the solidification features of an ophitic flow (approximately 20-30 m thick).

Raspberry Island, which is one of many small islands along the south side of Rock Harbor, is a sequence of three ophitic flows dipping 15°SE. The uppermost of these flows is extensively exposed on a wave-washed dip slope along the southeast shore. This shoreline receives strong storm waves and, therefore, has extensive wave-washed exposures about 1 km (0.6 mi) long. These expose the flow interior, although neither the top nor the base of the flow is clearly exposed.

A loop trail goes around the west half of the island, marked by informative signs about the unique ecosystem of this island, which features frequent fog and damp, moss-rich swamps. Among the unusual plants is the pitcher plant (*Sarracenia puerperia*), which is an insectivorous plant that flourishes in the swamp along the loop trail.

First, we will visit the most west point of the island, where the regional attitude of the lava flows is seen in the view along strike toward Smithwick Island across the Smithwick Channel. The point facing the channel on Raspberry Island is underlain by the oldest of the three flows on Raspberry Island. We will walk on a dip slope that shows some of the jointing pattern we observed on the southeast sides of Davidson and Smithwick Islands. Next, we will head to the southeast corner of the island to observe some rude columns in the uppermost Raspberry Island flow.

The west Part of the southeast Shore: Vesicle Cylinders

We will next move out to the wave-washed southeast shore, where there are two zones of exposures of vesicle cylinders (Fig. 1.24). Paces (1988) describes vesicle cylinders (Goff 1977) in the PLV:

Figure 1.24: Vesicle Cylinders at Raspberry Island. a. In cross-section the cylinders display an equidimensional aspect while b. in cross section they are elongate.

> Vesicle pipes are elongated, tube-like structures, 10-30 cm (4-12 in) in diameter and 0.5-2 m (1.6-6.6 ft.) in length, containing somewhat coarser and more prismatic crystals compared to the adjacent groundmass. They are oriented vertically and occur predominantly in the bottom half of the flow. The origins and dynamic behavior of vesicle cylinders are poorly understood; however they appear to represent an accumulation of exsolved magmatic gas bubbles which migrate upwards through the magma during the period when the cooling magma behaves as a Bingham plastic (i.e., possesses a finite yield strength, Walker 1987).
>
> --Paces 1988

Here at Raspberry Island, exposures of vesicle cylinders show a regular spacing between them, 1-3 m (3-10 ft.) apart, and a marked variety of textures; some were evidently preserved almost as voids, while others are filled with material that closely resembles vesicular pegmatoid. An interesting aspect of the exposures here is the relationship between the ophitic textures of the flow and the vesicle cylinders. The grain size of oikocrysts seems to be influenced by the proximity to the vesicle cylinder.

Vesicle cylinders are found mainly in only two areas along this shoreline. This may reflect their restricted occurrence in a thin part (less than a few meters thick) of this flow. Based on limited field examination, this thin part seems to be in the lower part of the flow. The comparisons between this occurrence and written descriptions, one by Paces (1988) of the PLV on the Keweenaw (Fig. 1.25) and one by Marsh and others (1991) of solidification in sheet-like basaltic bodies (Fig. 1.26), are illuminating.

Both descriptions show the vesicle cylinders as developing in the upper part of the lower capture front of a solidifying body and continuing upward to the liquid zone. This suggests that the segregations formed during solidification--when the region around the cylinders was mushy. The vertical movement of the cylinders was accommodated by the positive thermal gradient and by more fluid material that overlay the mush region.

The process forming the cylinders could possibly be common to the solidifying process of a wide variety of materials, such as metals, aqueous solutions, and organic mixtures; this process is described as channel convection in partly solidified systems (Hellawell and others, 1993). In experiments with aqueous solutions, Hellawell and others (1993) present observations of plumes that formed above the channels and extended into the liquid zone above (Fig. 1.27).

The Central Part of the Southeast Shore: Slickenside Surfaces

Featured conspicuously along the east shore of Raspberry Island are slickenside surfaces (Fig. 1.28). A study of fault slickenfibers allowed Witthuhn (1993) to use geometrical and statistical methods to define the kinematics of the closing of the rift. In Witthuhn's study, Raspberry and Edwards Islands offered one of the largest populations of measurements (Fig. 1.29). The measurements revealed two consistent stress fields, for each limb of the syncline, that would satisfy the conditions envisioned for the opening and closing of the Midcontinent rift. Most of the faults on Isle Royale, including both normal and reverse faults, trend northeast. This suggests that the reverse faults represent reactivated normal faults. The orientation of reverse faults at Isle Royale differs significantly from the predominately N-S trending structures measured in the PLV on the Keweenaw Peninsula (Fig. 1.30).

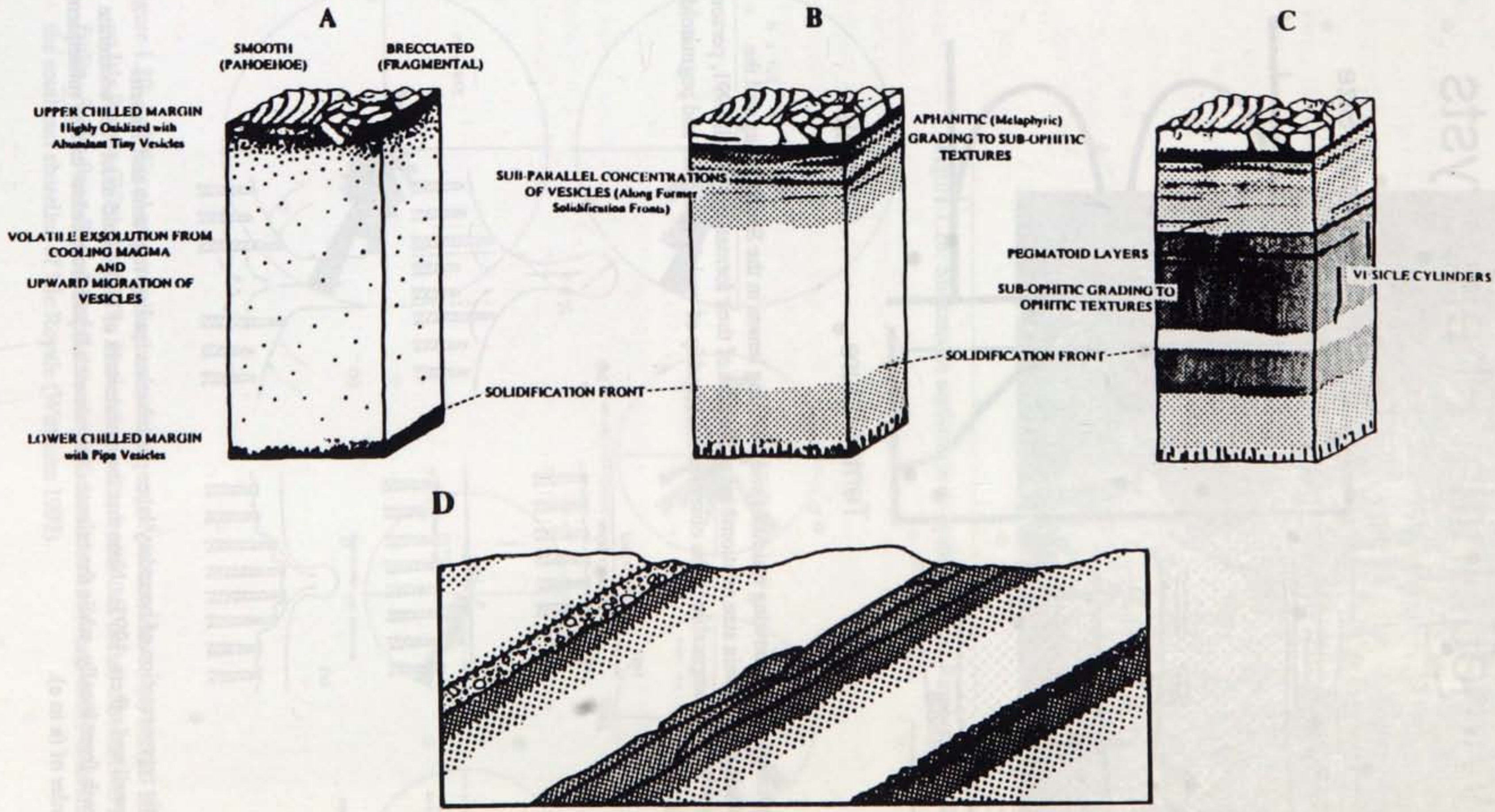


Figure 1.25: Illustration of the development of volcanic rock textures and flow structures typical in sub-aerially erupted PLV flood lavas (Paces 1988). Unstippled areas within the cooling lava flow interior represent molten material. (a) ponded lava flow after emplacement (b) partially solidified (c) solidification near complete (d) illustration of a portion of four successive lava flows summarizing the hydrodynamic characteristics of the volcanic pile. Very high permeability is present in interflow sediments and in highly vesicular and/or brecciated flow tops (heavy stippling). Permeability decreases gradually as the degree of vesicularity diminishes (light stippling) towards the center of the flow. Massive interiors of thicker flows (unpatterned) are largely impermeable to migrating ground water and secondary hydrothermal fluids. Lava flows were originally deposited in a near-horizontal orientation and have been subsequently tilted by both depositional and compressive tectonic mechanisms.

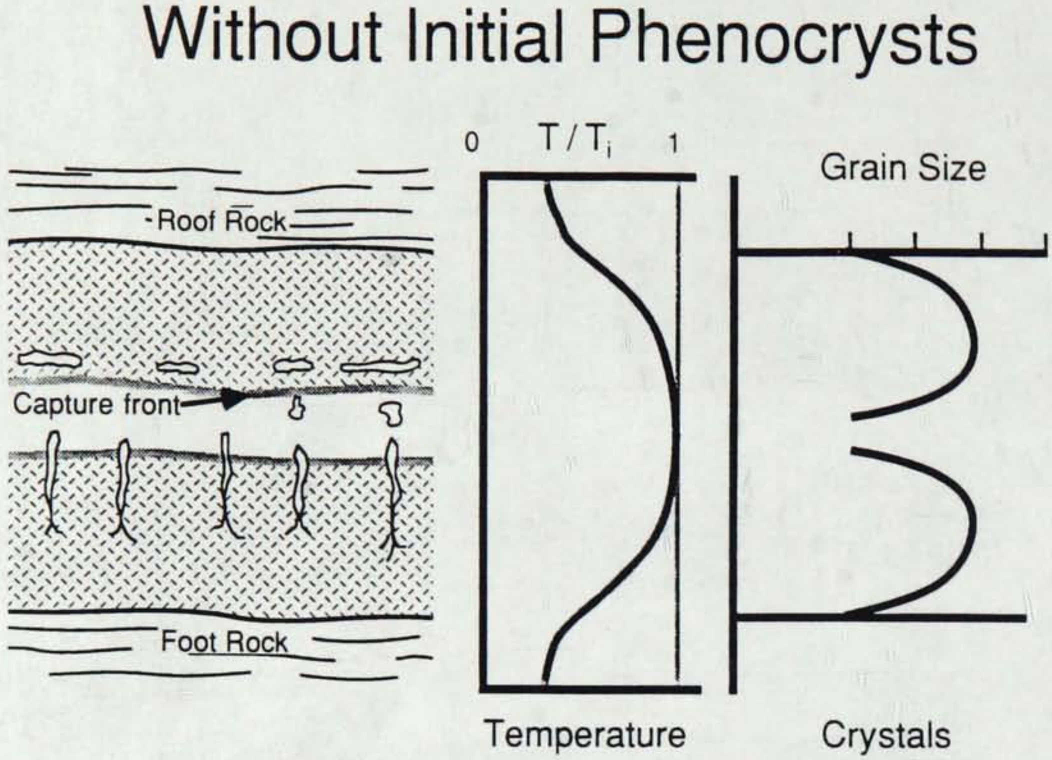


Figure 1.26: Conceptual diagram, showing residual fluids forming lenses in the Shonkin Sag and the schematic temperature and grain size relationships at the time of their formation (Marsh 1991, personal communication). This is comparable to the observed relationship of vesicle cylinders and pegmatoids on Raspberry Island.

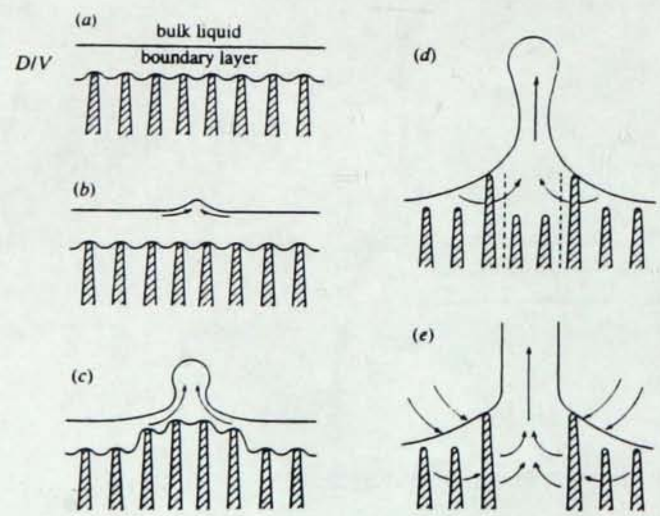


Figure 1.27: Schematic representation of boundary layer perturbation leading to plume and channel formation (Hellawell and others, 1993). Note that the entrainment of bulk liquid must first accelerate the dendritic growth front locally, while the release of solute-rich liquid from below causes subsequent melting, in the order of (a to e).

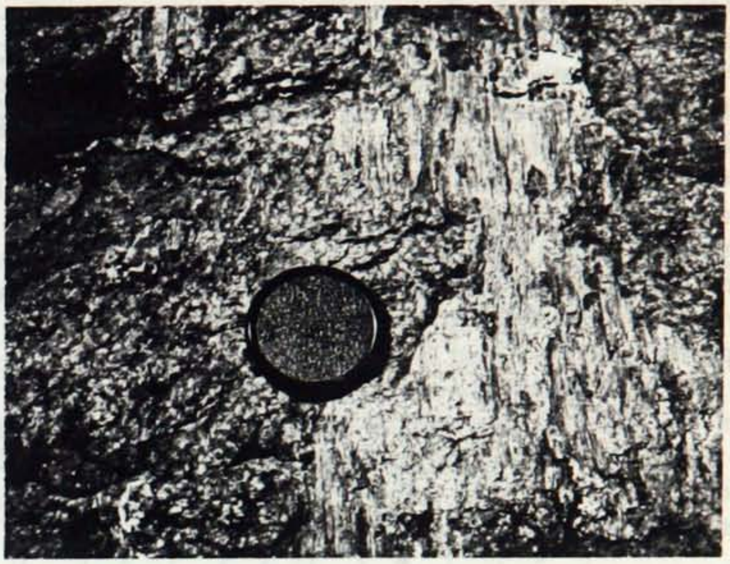


Figure 1.28: Slickenside surfaces in an ophitic flow on Raspberry Island.

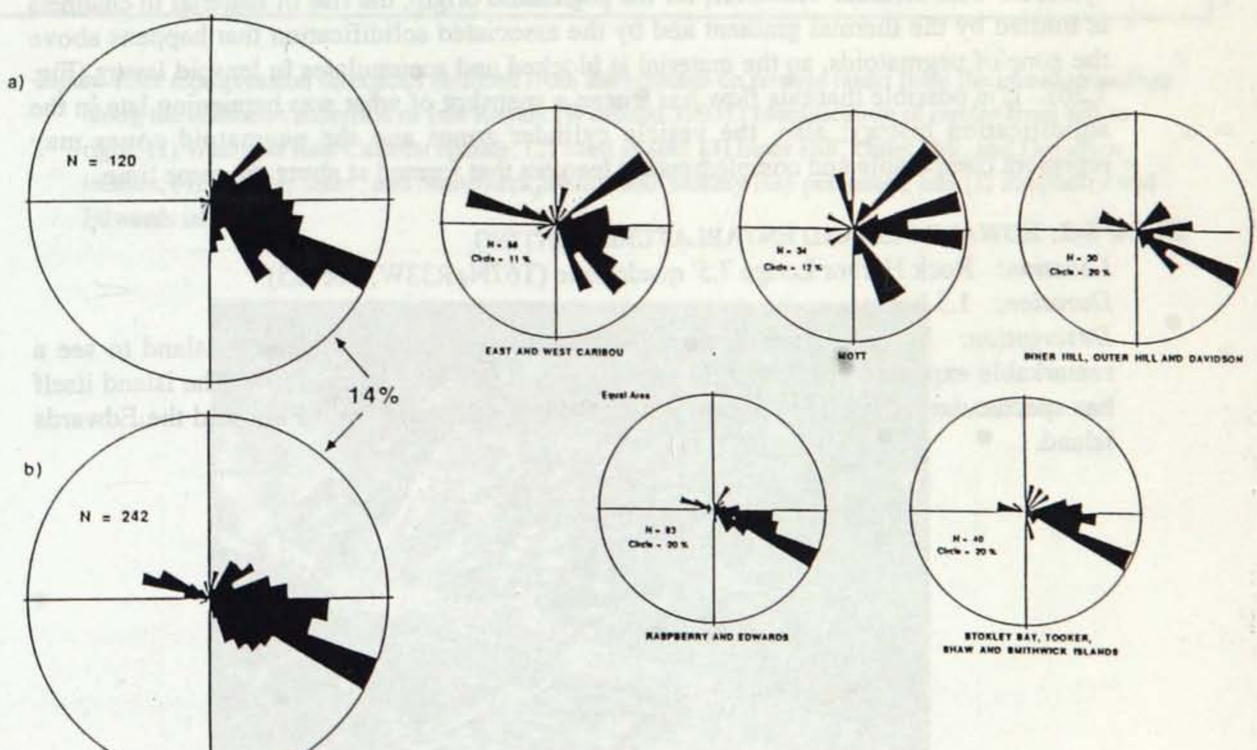


Figure 1.29: *Left:* Equal area rose diagrams of the trend of slickensides on (a) normal and (b) reverse faults on Isle Royale (Witthuhn 1993). Notice the similar trends that define the resolved shear stress on the faults. *Right:* Rose diagrams of the trends of slickensides on reverse faults measured on islands along the southeast shoreline of Isle Royale (Witthuhn 1993).

The East Part of the Southeast Shore: Pegmatoid Zones

About two-thirds of the way along the shore of Raspberry Island, the exposures that occur are stratigraphically higher in the flow. Here the flow has a laminar structure that consists of fractures that are parallel to the bedding and spaced about 0.5-3 cm (0.2-1.2 in) apart. Within this part of the flow, vesicle cylinders are not seen, but small pegmatoid lenses occur (Fig. 1.31). Paces (1988) describes what these lenses look like:

> Pegmatoid horizons are similar to vesicle cylinders in that they consist of gas-rich, coarsely crystalline, granophyric material. However, they occur as discontinuous lenses and layers, typically 10 cm (4 in) to several meters thick, and are usually located between the flow top and most massive portion of the flow interior. Pegmatoids are best developed in thicker flows that have cooled slowly enough to allow in situ differentiation (Cornwall 1951; Lindsley and others 1971). This material represents the last remaining volatile-rich liquid, which is injected into fractures oriented sub-parallel to the upper flow surface. Both vesicle cylinders and pegmatoid layers contain significant void space in the form of vesicles and gas pockets and contribute to the permeability of the lava flows.
>
> --Paces 1988

The origin of the pegmatoids is possibly related to the process by which the vesicle cylinders were formed. However, for the pegmatoid origin, the rise of material in channels is limited by the thermal gradient and by the associated solidification that happens above the zone of pegmatoids, so the material is blocked and accumulates in lensoid layers (Fig. 1.26). It is possible that this flow has frozen a snapshot of what was happening late in the solidification history; also, the vesicle cylinder zones and the pegmatoid zones may represent comparable and complementary features that formed at about the same time.

STOP 5-2: EDWARDS ISLAND ENTABLATURE JOINTING

Location: Rock Harbor Lodge 7.5' quadrangle (T67N, R33W, Sec. 25)

Duration: 1.5 hours

Description: Weather permitting, we will motor and land on Edwards Island to see a remarkable exposure of entablature jointing in the Edwards Island Flow. The island itself has spectacular morphology of two tilted flows, the Scoville Point Flow and the Edwards Island.

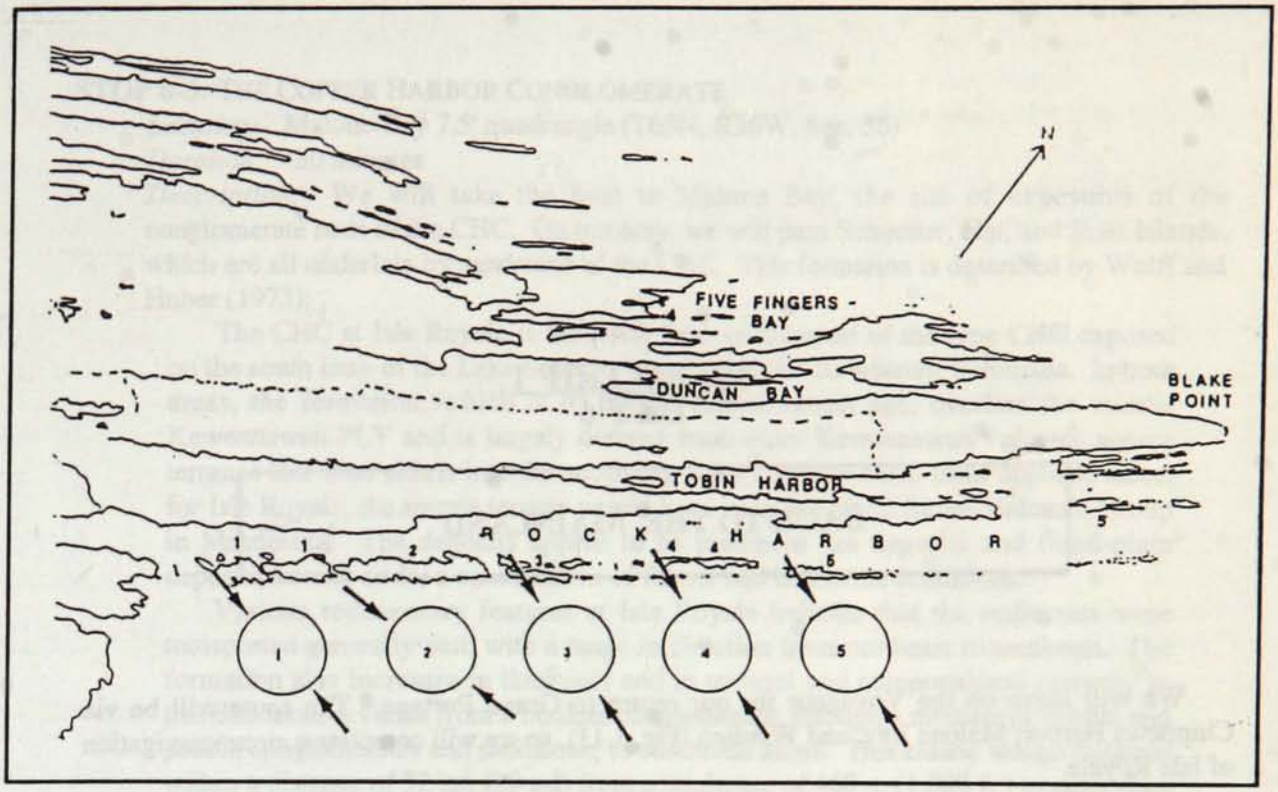


Figure 1.30: Compression directions deduced from slickenlines on reverse faults from the island groupings along the southeast shoreline of Isle Royale (Witthuhn 1993). Identification of circles from left to right: (1) West and East Caribou islands; (2) Mott island; (3) Inner Hill, Outer Hill, and Davidson islands; (4) Tooker, Shaw, and Smithwick islands and Stokley Bay peninsula; and (5) Raspberry and Edwards islands.

Figure 1.31: Thin (10 cm) planar lensoid pegmatite zone in the ophitic flow on Raspberry Island.

FIELD TRIP 1
DAY 6

BACK TO THE MAINLAND

We will leave on the Voyageur for our return to Grand Portage. The route will be via Chippewa Harbor, Malone Bay, and Windigo (Fig. 1.11), so we will complete a circumnavigation of Isle Royale.

STOP 6-1: CONGLOMERATE BAY

Location: Lake Ritchie 7.5' quadrangle (T66N, R34W, Sec. 26, 35)

Duration: 10 minutes

Description: The first leg of the trip is southwest along Rock Harbor to Daisy Farm Campground, and then past the Isle Royale Lighthouse and Tonkin Bay to Conglomerate Bay. Conglomerate Bay is named for one of the interflow sediments that crops out halfway along the inlet. On the edge of this bay was the site of the Saginaw Mine, which was a very brief mining effort with two shafts and a winze from 1877 to 1879.

The regular 15° dip of the PLV strata is disrupted in this part of Isle Royale with a system of steep and curved faults. This causes the more southeast parts to be offset progressively upward, as though the regular subsidence of the rift basin was disrupted in this place.

STOP 6-2: CHIPPEWA HARBOR

Location: Lake Ritchie 7.5' quadrangle (T65N, R34W, Sec. 17)

Duration: 30 minutes

Description: The entire region of Middle Island Passage to Siskiwit Lake consists mainly of the ophitic flows and their intervening sedimentary units (Fig. 1.18). At Chippewa Harbor, we will see exposures of these flows and the sedimentary units, and near the dock is an outcrop of the interflow conglomerate.

The harbor is the site of a fishing camp and is now a favored campsite. The beaches south of Chippewa Harbor are some of the best for finding Isle Royale greenstones (see Huber 1983, pp 58-9).

STOP 6-3: THE COPPER HARBOR CONGLOMERATE

Location: Malone Bay 7.5' quadrangle (T65N, R36W, Sec. 36)

Duration: 30 minutes

Description: We will take the boat to Malone Bay, the site of exposures of the conglomerate beds of the CHC. On our way, we will pass Schooner, Hat, and Ross Islands, which are all underlain by sandstone of the CHC. This formation is described by Wolff and Huber (1973):

> The CHC at Isle Royale is the north limb counterpart of the type CHC exposed on the south limb of the Lake Superior syncline on the Keweenaw Peninsula. In both areas, the formation, which is of middle Keweenawan age, overlies the middle Keweenawan PLV and is largely derived from older Keweenawan volcanic source terranes that shed debris into the subsiding Lake Superior basin from opposite sides; for Isle Royale, the source terrane would have been the North Shore Volcanic Group in Minnesota. The deposits appear to be piedmont fan deposits and flood-plain deposits formed under a combination of fluvial and lacustrine conditions.
>
> Various sedimentary features at Isle Royale indicate that the sediments were transported generally east, with a range in direction from northeast to southeast. The formation also increases in thickness and in textural and compositional maturity in this direction; it varies from a boulder conglomerate, through a mixture of cobble and pebble conglomerates and sandstone, to sandstone alone. This clastic wedge thickens within a distance of 32 km (20 mi) from a minimum of 460 m (1,500 ft.) to more than 1830 m (6,000 ft.) between stratigraphic marker horizons; the top of the formation is nowhere exposed, however, and the total thickness probably is considerably greater.
>
> The clastic materials in the formation are predominantly derived from older Keweenawan volcanic rocks. Felsic varieties are slightly dominant over mafic varieties. Metamorphic rocks are minor components, and clasts of intrusive igneous rocks appear to be absent. The overall local composition of the formation, chiefly reflecting degree of sediment maturity, is closely related to grain size; rocks of similar grain size have the same composition throughout the formation. Calcite cement is ubiquitous; in the conglomerate and coarse-grained sandstone, it commonly amounts to about 15 percent. As the textures become finer, rock fragments and calcite cement decrease in abundance, whereas quartz, feldspar, and opaque mineral percentages increase.
>
> The felsic and many of the mafic Keweenawan rock types that occur as clasts in the CHC are not known to occur as flows within the PLV on the island. Most of the rock types, however, are known to occur as flows in the North Shore Volcanic Group, which is stratigraphically lower than the PLV and is present in the direction from which the clastic debris was transported. Several lines of evidence suggest the presence of an unconformity between the North Shore Volcanic Group and the PLV. Such an unconformity would facilitate the erosion of the North Shore Volcanic Group and the deposition of the erosional debris basinward on top of the PLV.
>
> The situation would be analogous to that on the south side of the Lake Superior syncline. There, many clasts in the CHC appear to have been derived from the lowermost Keweenawan volcanic terrane, the volcanic rocks of the South Trap Range which underlie the PLV.
>
> --Wolff and Huber (1973)

We will move along shorelines underlain by the CHC all the way to Grace Harbor, passing Menangerie Island, Long Island, Point Houghton, McCormick Reef, Long Point, the Head, Rainbow Point, and Cumberland Reef (Fig. 1.11). Figure 1.32 shows the variation of sediment lithology across the section and Figure 1.33 shows sediment disposal patterns measured by Wolff and Huber (1973).

Composition of very fine grained sandstones, very coarse grained sandstones, and pebble conglomerates taken from the measured section on Houghton Ridge
[Specimen numbers indicate relative position with lowest number lowest in section. See plate 1 for locations]

	Very fine grained sandstones						Very coarse grained sandstones						Pebble conglomerates				
Specimen No.	4	6	7	11	13	28	12	15	18	22	24	27	5	16	30	31	32
MV	0.3	0.6	0.9	Trace	0.9	0.0	30.7	36.9	35.5	35.6	23.9	21.7	22.7	34.3	28.2	22.9	29.4
FV	.9	.6	.9	0.9	.6	.0	31.8	38.8	31.7	33.6	24.2	39.6	43.7	39.9	37.2	21.9	34.6
SF	4.2	.3	.0	Trace	.0	.0	.0	.9	.6	.9	1.6	Trace	.9	.3	1.9	3.5	.3
MF	.0	.0	.3	.0	Trace	.0	.6	Trace	2.3	Trace	.6	.3	.9	.3	1.3	1.3	Trace
UQ	46.2	39.7	42.3	36.5	32.4	28.1	4.8	3.5	3.9	4.5	4.5	1.6	1.9	2.9	4.2	6.1	4.2
UnQ	21.6	16.2	18.8	24.2	23.2	11.9	4.8	1.6	1.9	1.3	2.9	1.9	.9	.6	1.3	3.5	1.3
PQ	.6	.6	.6	.0	.9	.0	.3	.6	.9	.6	1.6	1.3	1.3	1.9	.6	1.3	.9
PF	4.8	15.5	12.9	23.3	18.5	22.9	1.6	1.6	.9	.9	3.9	4.5	7.1	.6	1.6	2.9	1.9
PIF	5.5	8.4	5.2	5.5	6.1	20.6	9.0	11.6	5.4	6.8	10.0	2.3	2.9	5.9	1.9	2.9	4.5
O	6.6	6.8	8.1	5.2	5.8	6.6	6.1	2.3	1.3	5.2	7.1	1.6	1.3	1.6	.9	8.1	7.1
Ca	6.1	6.8	5.5	1.9	8.4	4.5	9.4	1.3	15.3	9.7	17.1	22.3	14.8	9.8	19.0	23.3	14.2
Other	3.2	4.5	4.5	3.5	3.2	5.4	.9	.9	.3	.9	2.6	2.9	1.6	1.9	1.9	2.3	1.6

NOTE.—Abbreviations:
MV. mafic volcanic rock fragments
FV. felsic volcanic rock fragments
SF. sedimentary rock fragments (shale and sandstones)
MF. metamorphic rock fragments
UQ. unstrained quartz grains
UnQ. undulatory quartz grains
PQ. polycrystalline quartz grains
PF. potassium feldspar grains
PIF. plagioclase grains
O. opaque grains
Ca. Calcite cement
Others epidote, pyroxene, zeolites

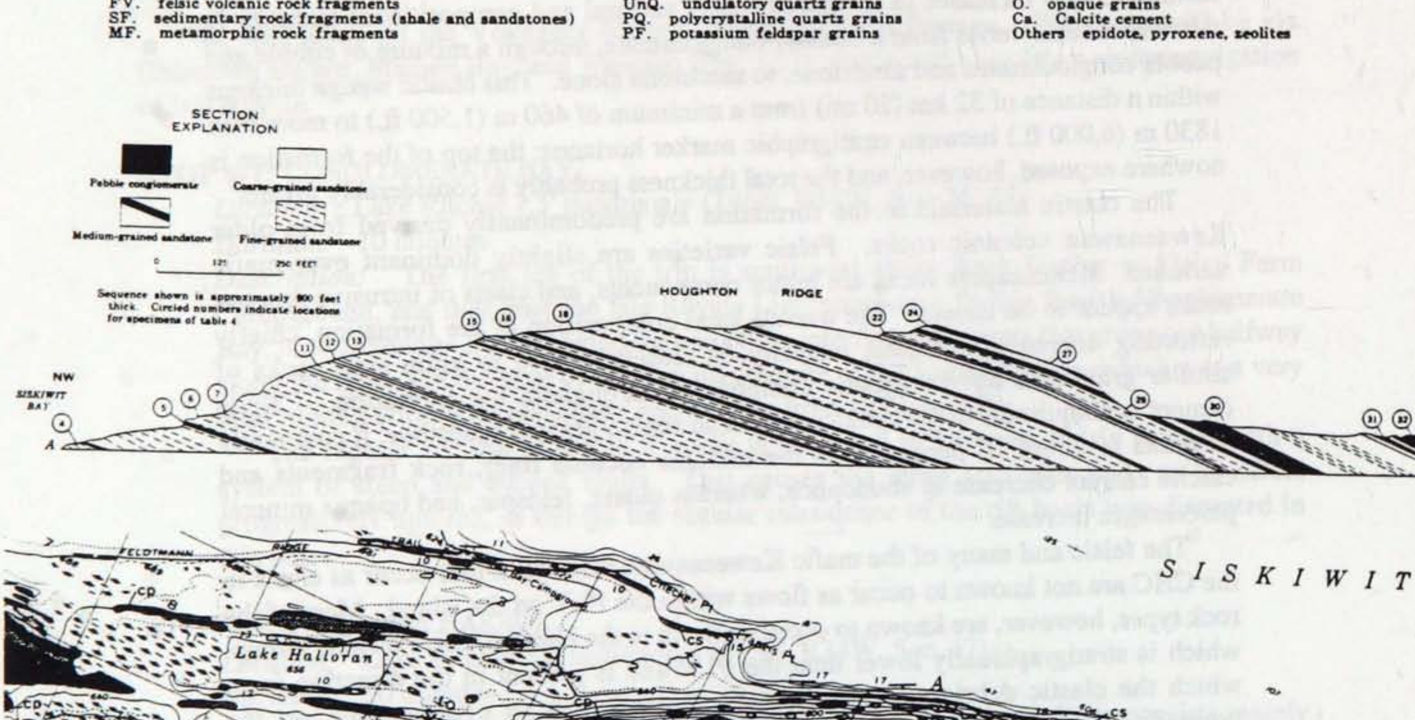


Figure 1.32: The variation of sediment lithology across Houghton ridge (Wolff and Huber 1973).

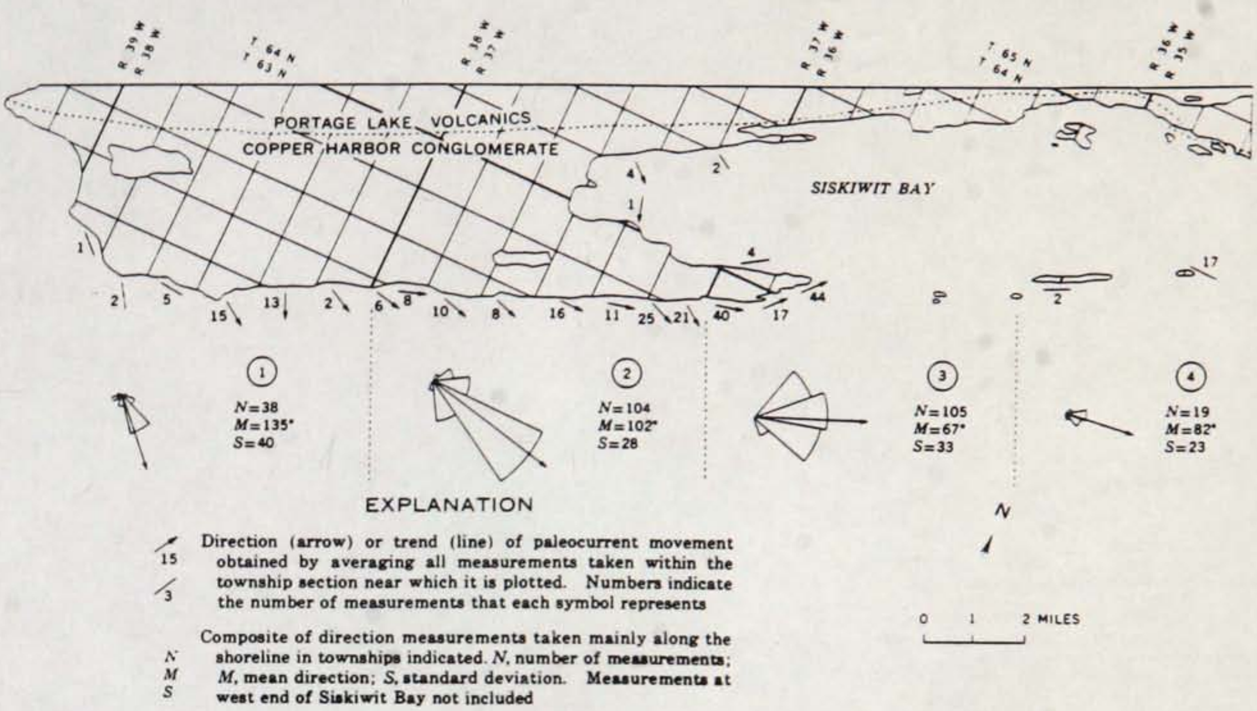


Figure 1.33: Sediment dispersal patterns in the Copper Harbor Conglomerate on Isle Royale (Wolff and Huber 1973).

STOP 6-4: THE GLACIAL DEPOSITS

Location: Most of southwest Isle Royale

Duration: 2 hours

Description: It is in this southwest section of Isle Royale (Fig. 1.34) that the thickness of glacial deposits is the greatest. The glacial setting is summarized by Huber (1973b):

> Isle Royale was overridden by glacial ice during each of the four major glaciations of the Pleistocene Epoch, and each successive glaciation essentially obliterated all direct evidence of preceding glaciations on the island. In the waning phase of the last major glaciation, the Wisconsin Glaciation, the frontal ice margin retreated north from at least the greater part of the Lake Superior basin, then readvanced into the basin during Valders time, about 11,000 years ago. We can attribute to the Valders ice the final aspect of glaciation at Isle Royale, including both erosional and depositional features.
>
> It is impossible to estimate the quantity of glacial debris or other surficial materials that might have been present at Isle Royale prior to the Valders readvance, but the readvancing ice appears to have removed most of what might have been present, as judged by the thin surficial cover on the east two-thirds of the island today. During the Valders retreat, a series of lakes formed in the Lake Superior basin in front of the retreating ice margin.

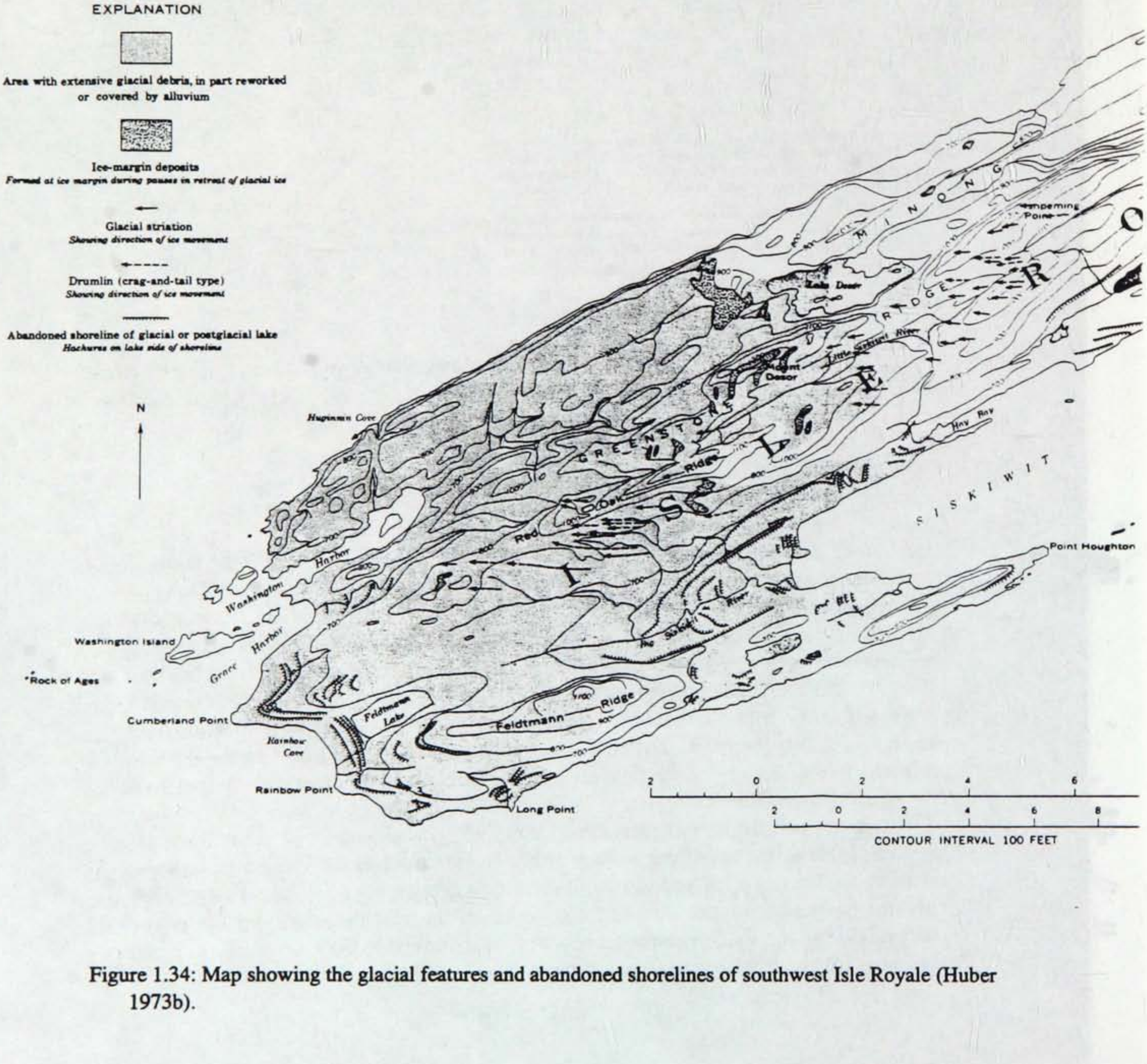


Figure 1.34: Map showing the glacial features and abandoned shorelines of southwest Isle Royale (Huber 1973b).

The retreating ice opened successively lower outlets, and thus the general trend of lake elevations is downward. Distinct lake stages reflect periods of relative stability during which well-defined shoreline features developed. The ice front forming the north margin of the earlier lakes probably remained south of Isle Royale until about the time of glacial Lake Beaver Bay, when it retreated to a position straddling Isle Royale west of Lake Desor. Abundant deposits of glacial debris were left upon the newly deglaciated west end of the island, and the ice front remained stable long enough to build a complex of ice-margin deposits across the island. Shorelines formed by the glacial lake associated with this ice front are found on the west part of the island about 61 m (200 ft.) above present Lake Superior.

Subsequent renewed and complete retreat of the ice margin from Isle Royale was rapid enough that only a minor amount of glacial debris was deposited on the central and east parts of the island. When the ice margin reached the north edge of the Lake Superior basin, Lake Minong was formed, and the entire basin was filled for the first time since the Valders advance. Lake Minong marked a relatively stable episode in the history of the basin, and its beaches are among the best developed of the abandoned shoreline features on Isle Royale. Lake Minong beaches and later lower beaches are best developed on the southwest end of Isle Royale, where abundant glacial debris provided easily worked materials for beach construction.

--Huber (1973b)

The discussion about changing shoreline levels in Huber (1983, see pp 41-55) is important to understand the geomorphology we will be seeing, and if you are a field trip participant, you should particularly read the discussion of shorelines at Rainbow Cove, which includes clear exposures of both Nipissing and Minong shorelines (Huber 1983, see pp 52-3).

STOP 6-5: THE END OF OUR TRIP

Location: Washington Harbor, Windigo 7.5' quadrangle (T64N, R38W, Sec. 29)

Duration: 30 minutes

Description: Rounding the Cumberland Reef we will arrive once again at Washington Harbor, where we began on the first day. You have now seen Isle Royale from all sides.

Acknowledgments

The opportunity to write a detailed guide to Isle Royale and to lead a field trip comes from the cooperation of many people. I would like to thank Jack Oelfke of Isle Royale National Park and Lynne Olson at UMD for helping with logistics and formalities. King Huber provided us with a complete set of his many publications about Isle Royale and also with lots of cheerful encouragement. Jim Paces, Tony Longo, and Rick Wunderman provided me with a lot of insight on the volcanic geology of Isle Royale. Kate Witthuhn supplied some unpublished data. Discussions with Bruce Marsh and Angus Hellawell about solidification helped me to understand a little better what may have been going on inside Isle Royale's lava flows. Dave Schneider planned the food and logistics for the trip. Finally, Libby Titus did all the work of putting this field guide together: revisions, copy-editing--everything but the fun parts!

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FIELD TRIP 2

STRATIGRAPHY, STRUCTURE, AND ORE DEPOSITS OF THE SOUTHERN LIMB OF THE MIDCONTINENT RIFT SYSTEM

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Overview

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The Midcontinent rift of North America (MCR) extends more than 2000 km northeasterly from Kansas through the Lake Superior region and then southeasterly through lower Michigan (Fig. 1.1). The rift has been postulated to be the result of a mantle plume, based on the enormous volume of dominantly tholeiitic composition igneous rocks and the short interval of subaerial eruption (Paces and Bell, 1989; Hutchinson and others, 1990; Nicholson and Shirey, 1990). This field trip is a traverse from west to east through northern Wisconsin and upper Michigan, providing a complete temporal and stratigraphic cross-section of the MCR, including early and late rift-filling volcanic rocks, sedimentary rocks in both the central rift graben and flanking basins, and the major reverse faults (originally normal faults) bounding the central graben (Figs. 2.1 and 2.2). In addition, rift rocks in Michigan host the actively mined world-class White Pine copper deposit, the dormant but historically important Keweenawan native copper district, and a zone of chalcocite mineralization in basalt for which production is planned in 1995 (Fig. 2.3). This summary of the stratigraphy, structure, and mineralization of the MCR provides an introduction to the rift rocks in northern Wisconsin and upper Michigan. More detailed descriptions can be found in the overviews and specific stop descriptions for each day of the field trip.

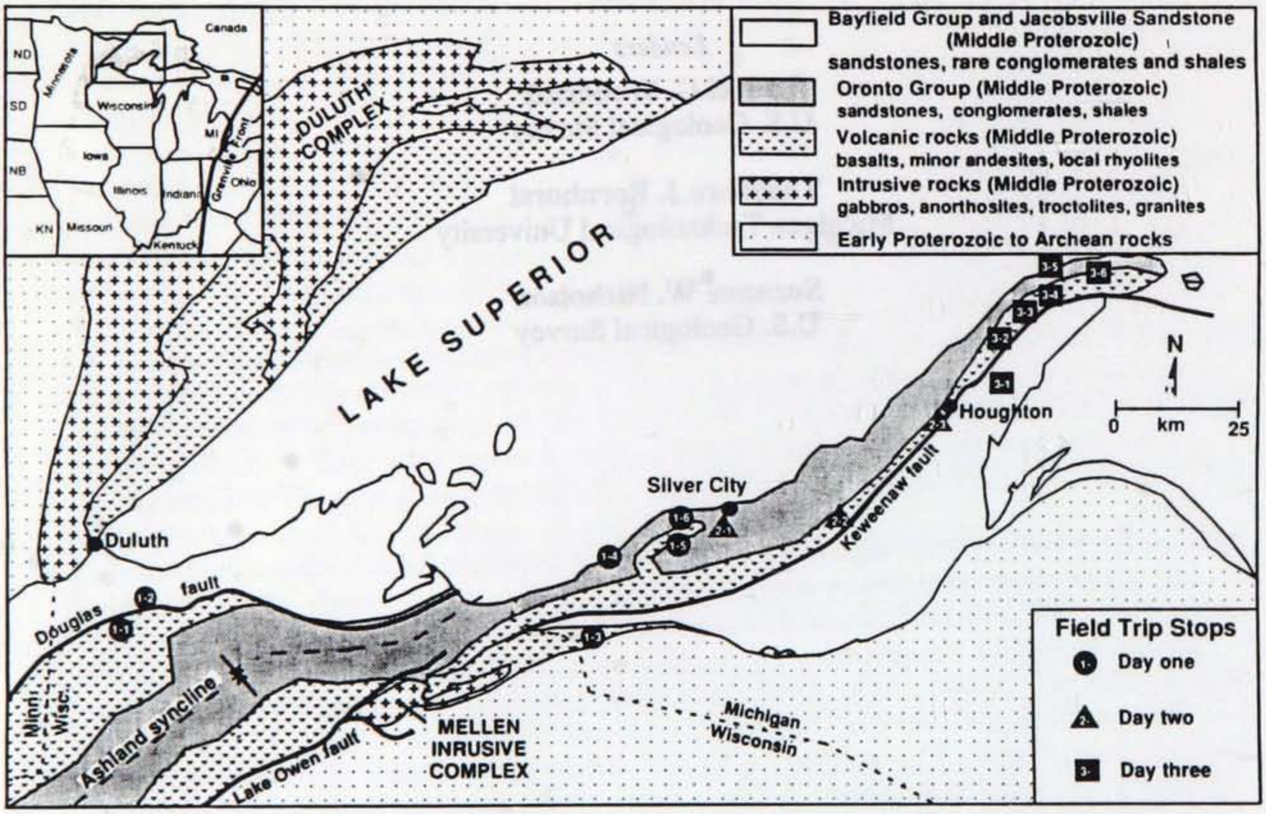

Figure 2.1: Generalized geological map of the western Lake Superior region showing location of field trip stops. Inset shows the location of the Midcontinent rift and the Grenville Front.

Stratigraphy and Structure

The oldest unequivocal rift rocks exposed in northern Wisconsin and upper Michigan are quartz-rich fluvial and lacustrine sandstones of the Bessemer Quartzite. These sandstones were deposited in a broad basin in response to initial thinning of the crust on the site of the future rift (Ojakangas and Morey, 1982) (Fig. 2.2). In this area the beginning of active volcanism, estimated to be about 1109 Ma (Davis and Sutcliffe, 1985), is marked by eruption of high-alumina basalts of the Siemens Creek Volcanics of the Powder Mill Group, which spilled out directly from fissures onto wet sands of the Bessemer Quartzite. The first 50 to 100 m of the Siemens Creek Volcanics are pillowed, but nearly all following volcanic eruptions associated with the rift were subaerial. Above the Siemens Creek Volcanics is a thick section of mafic to felsic rocks, the Kallander Creek Volcanics, stretching more than 100 km along strike from central Wisconsin eastward into Michigan. This section is composed of basalts, andesites, and rhyolites, postulated to be the remnants of a central volcano that erupted between 1108 and 1099 Ma (Cannon and others, 1993a; Zartman, unpublished date).

Emplaced into the Siemens Creek Volcanics and the Kallander Creek Volcanics in north-central Wisconsin were several layered to massive intrusions. These include the Mellen Intrusive Complex (Fig. 2.1), a composite intrusion made up of several layered gabbro bodies (gabbro at Mineral Lake and gabbro at Potato River) and other related intrusions, including a cross-cutting granite body (granite at Mellen), a series of gabbro and granophyre sills, and a small peridotite body (gabbro at Rearing Pond). U-Pb zircon dates from a granophyre in the gabbro at Mineral Lake and the granite at Mellen are indistinguishable from one another at 1102±2 Ma (Cannon and others, 1993a).

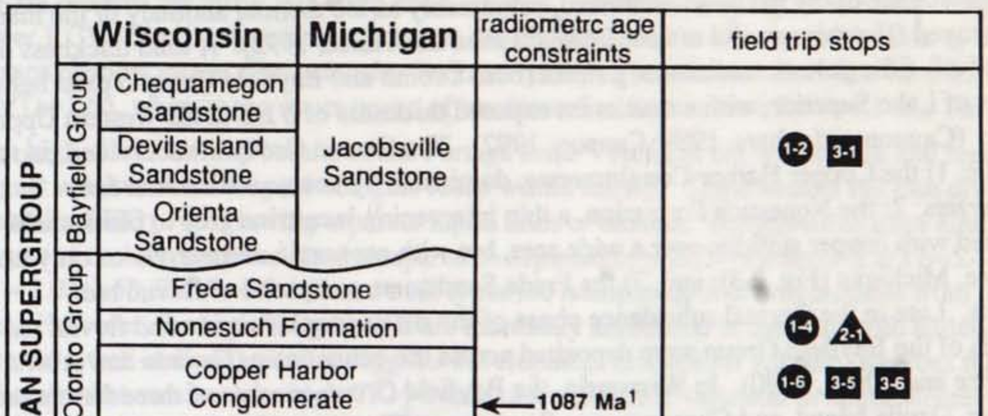

Figure 2.2: Generalized stratigraphic correlation for Keweenawan Supergroup units in Michigan and Wisconsin with field trip stops and radiometric age constraints. Age date references: 1) Davis and Paces (1990), 2) Zartman (unpublished data), 3) Cannon and others (1993a), 4) Davis and Sutcliffe (1985).

The Kallander Creek Volcanics are overlain by a voluminous section of olivine tholeiite flood basalts that erupted within the rapidly subsiding rift basin, now beneath Lake Superior. These basalts are the Chengwatana Volcanics in western Wisconsin and Minnesota and the Portage Lake Volcanics of the Bergland Group to the east in Wisconsin and Michigan. Where exposed on the Keweenaw Peninsula of Michigan, the Portage Lake Volcanics have a thickness of about 5 km, although the bottom of the section is cut off by the Keweenaw fault. Near the rift axis the Portage Lake Volcanics and older basalts are estimated to be more than 20 km thick based on seismic reflection profiles (Cannon and others, 1989). High precision U-Pb dates on thick flows near the top and bottom of the Portage Lake Volcanics section indicate that most of the section was erupted between about 1096 and 1094 Ma (Davis and Paces, 1990). Interflow sediments of coarse immature conglomerate and sand deposited by streams flowing towards the rift center record hiatuses in volcanic activity (Merk and Jirsa, 1982). Interspersed among the Portage Lake basalt flows, for example, there are 22 prominent interflow sedimentary layers, several of which are marker beds that are traced more than 100 km along strike (Butler and Burbank, 1929).

Overlying the Portage Lake Volcanics of the Bergland Group in western upper Michigan are the Porcupine Volcanics, a 5 km thick sequence of basalt, andesite, felsite, and rhyolite, erupted from a large shield volcano that was active late in the volcanic history of the rift (Nicholson and others, 1991a; Cannon and others, 1992). This eruptive activity marked the end of major volcanism in the region; later events in the rift were dominated by fluvial and lacustrine sedimentation.

A thick succession of rift-filling clastic sedimentary rocks, the Oronto Group and Bayfield Group, overlies the rift-filling volcanic rocks (Figs. 1.2, 2.1, and 2.2). When magmatic activity waned, subsidence of the rift basin continued, presumably as the thermal anomaly of the mantle plume decayed (Hutchinson and others, 1990; Cannon and Hinze, 1992). A total thickness of up to 8 km of rift-filling clastic sedimentary rocks (both Oronto and Bayfield Groups) exist beneath the center of Lake Superior, with a maximum exposed thickness of 6 km in the western Upper Michigan (Cannon and others, 1989; Cannon, 1992). The Oronto Group is subdivided into three formations: 1) the Copper Harbor Conglomerate, dominated by coarse red alluvial fan conglomerates; 2) the Nonesuch Formation, a thin intervening lacustrine gray to black shale mineralized with copper sulfides over a wide area, but with economic concentrations only near White Pine, Michigan (Fig. 2.3); and, 3) the Freda Sandstone, composed of fluvial red sandstones. Late in the thermal subsidence phase of the rift mature lacustrine and fluvial red sandstones of the Bayfield Group were deposited across the entire basin (Cannon and others, 1989; Hinze and others, 1990). In Wisconsin, the Bayfield Group consists of three formations, the Orienta, Devils Island, and Chequamegon Sandstones (Thwaites, 1912; Morey and Ojakangas, 1982). In Michigan the Jacobsville Sandstone is the approximate stratigraphic equivalent of the Bayfield Group.

The last phase of the MCR was characterized by a transformation of original graben-bounding normal faults into reverse faults (Figs. 1.2, 1.5, and 2.1). The Douglas fault of northern Wisconsin and the Keweenaw fault of Upper Michigan are now low- to high-angle reverse faults, but originally were major graben bounding growth faults (Cannon and others, 1989). Reverse displacement along the Keweenaw fault resulted in steepening of volcanic and sedimentary units. Faults, fractures, and broad open folds within rift-filling strata developed in response to tectonic adjustment of the subsiding volcaniclastic section and to a compressional event (Butler and Burbank, 1929; White, 1968). Reset Rb-Sr biotite ages within older Precambrian basement rocks near the Michigan-Wisconsin border suggest that the high-angle reverse faulting occurred about 1060 ± 20 Ma (Cannon and others, 1993b), possibly starting as early as 1080 Ma, and was likely completed by about 1040 Ma (Cannon, 1994). The probable cause of this compressional event was continental collision within the Grenville province (Hoffman, 1989; Cannon and Hinze, 1992; Cannon, 1994) (inset, Fig. 2.1).

Geochemistry

On the south limb of the MCR, the poorly exposed lowermost flows of the Siemens Creek Volcanics consist of slightly alkaline to transitional tholeiites that fall into two compositional groups (Table 2.1). One group has clinopyroxene phenocrysts, a rarity in younger high-alumina flows, and is characterized by low Al_2O_3 content, steep REE patterns (ave. La/Yb ≈ 23), Th/Ta = 1.45, and Zr/Y = 11. In contrast, the second group is either aphyric or contains olivine and plagioclase phenocrysts rather than clinopyroxene phenocrysts. Additionally, this group has higher Al_2O_3 content, lower LREE enrichment (ave. La/Yb ≈ 11.2), Th/Ta = 2.6, and Zr/Y = 6.1 (Nicholson and others, 1991b).

The most striking geochemical characteristic of the basalts in the overlying Kallander Creek Volcanics is the high average TiO_2 content (about 3.5 wt %) compared to basalts of the Siemens Creek Volcanics (about 1.5-2.5 wt %), Portage Lake Volcanics (about 1.7-2.7 wt %), and Porcupine Volcanics (about 2.2 wt %). Distinctive trace element characteristics make it possible to recognize two groups of volcanic rocks within the Kallander Creek Volcanics (Nicholson and others, 1994). Both groups range from basalt to intermediate rocks and rhyolite and are very similar in major element composition. The lowermost 1.5 km of the Kallander Creek Volcanics (KCV) consist of basalt with minor andesite and rhyolite, all of which have steep straight REE patterns (for the basalts, La/Yb = 21.1) without significant Eu*/Eu anomalies. The upper part of the Kallander Creek contains all three rock types as well, but andesites dominate. The slopes of these REE patterns are much less steep and L-shaped (for the basalts, La/Yb = 6.35), compared to the lower KCV. Similar distinctions between the upper and lower suites can be made by using other trace element ratios. For example, the lower group of basalts and andesites have Zr/Y= 9.4 and Th/Ta=1.65, whereas the upper group of basalts and andesites have Zr/Y=6.1 and Th/Ta=2.8.

The high-alumina flood basalts of the Portage Lake Volcanics are widespread and represent the largest volume of the erupted volcanic rocks within the rift. These basalts fall into distinctly high- and low-TiO_2 types having separate liquid lines of descent. Incompatible trace elements in both types of tholeiites are enriched compared to depleted or primitive mantle (La/Yb=5.4-5.6; Th/Ta = 2.03-2.53; Zr/Y= 4.5-4.8) and both types are isotopically indistinguishable from one another. However, the high-TiO_2 basalts are intimately associated in the field with andesites and rhyolites in a well-defined unit that suggests development in a central volcano, whereas the low-TiO_2 basalts that surround the high-TiO_2 basalt unit show little variation in major element composition, and represent the fissure-fed flood basalts (Nicholson, 1990).

The Porcupine Volcanics consists of basalt, abundant andesite, and rhyolite flows, and is believed to represent a central volcano complex that overlies the Portage Lake Volcanics. In comparison with other Midcontinent rift basalts, incompatible trace element concentrations of the Porcupine Volcanics basalts are most similar to those of basalts in the upper Kallander Creek Volcanics and the high-TiO_2 basalts of the Portage Lake Volcanics (ie., Nb, Zr, Y, Hf and Ta). Like these other units, the Porcupine Volcanics show varying degrees of crustal contamination and has been postulated to be related to a central volcano complex.

The youngest rift basalts are exposed in the North Shore Volcanic Group in Minnesota but occur as dikes cutting older rift basalts in the Powder Mill Group in upper Michigan and Wisconsin. Compared to the mainstage basalts, the youngest basalts are more depleted in incompatible trace elements (La/Yb = 4.6, Th/Ta = 1.8, and Zr/Y= 4.4).

TABLE 2.1: COMPARISON OF AVERAGE BASALT COMPOSITIONS OF FORMATIONS RELATED TO THE MIDCONTINENT RIFT IN WESTERN LAKE SUPERIOR

	SIEMENS CREEK VOLCANICS		*KALLANDER CREEK VOLCANICS*		*PORTAGE LAKE VOLCANICS*		*PORCUPINE VOLCANICS*	*LATE BASALTS*
	Lower Unit (Cpx-bearing Basalts)	Upper Unit (Non-cpx-bearing Basalts)	Lower Unit Basalt	Upper Unit Basalt	Low-TiO2 Basalt	High-TiO2 Basalt	Basalt	Basalt
Element	N=10	N=15	N=10	N=6	N=11	N=8	N=27	N=6
SiO2	49.78	52.01	50.92	49.54	48.29	49.37	50.41	49.14
TiO2	2.18	1.63	3.62	2.76	1.70	2.68	2.14	1.08
Al2O3	9.41	14.94	14.28	14.26	16.99	15.21	15.12	17.39
Fe2O3	2.01	1.69	2.05	2.17	1.65	1.84	2.02	1.40
FeO	11.41	9.55	11.64	12.32	10.90	12.13	11.43	7.92
MnO	0.22	0.18	0.18	0.24	0.22	0.24	0.20	0.16
MgO	9.66	6.37	4.08	6.17	8.16	6.05	6.05	8.58
CaO	11.85	9.46	7.11	8.03	6.50	7.32	7.00	11.69
Na2O	2.11	2.82	4.18	2.71	2.86	3.01	2.90	2.35
K2O	1.06	1.14	1.40	1.41	2.53	1.81	2.11	0.17
P2O5	0.30	0.22	0.54	0.40	0.20	0.36	0.63	0.12
SUM	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
Nb	32	18	50	18	9	16	15	6
Zr	229	131	290	238	126	200	237	74
Y	21	23	31	39	26	44	43	17
Ba	374	362	471	474	559	379	1140	86
Cr	705	178	32	86	159	71	90	272
Ni	190	91	49	101	172	75	63	191
La	39.70	21.21	53.11	24.78	11.91	20.94	43.16	6.41
Sm	8.64	4.95	11.16	8.10	4.03	6.88	9.04	2.49
Yb	1.71	1.89	2.52	3.90	2.12	3.88	4.42	1.38
Hf	5.46	3.25	7.08	5.83	2.65	5.00	5.70	1.72
Ta	2.15	1.15	3.56	1.15	0.59	1.01	1.02	0.37
Th	3.12	2.65	5.86	3.18	1.20	2.56	5.38	0.66
La/Yb	23.2	11.2	21.1	6.4	5.6	5.4	9.7	4.6
Zr/Y	10.9	5.7	9.4	6.1	4.8	4.5	5.5	4.4
Th/Ta	1.5	2.3	1.7	2.8	2.0	2.5	5.3	1.8

Mineralization

Copper mineralization in rift-filling rocks began during diagenesis and continued through burial metamorphism with associated hydrothermal activity that altered primary minerals and filled open spaces with secondary minerals. Mauk and others (1992a, 1992b) presented evidence for two distinct stages of copper mineralization in the White Pine deposit (Fig. 2.3). Main-stage copper sulfides and subordinate native copper, which make up the bulk of the mineralization, formed during diagenesis of the Nonesuch Formation. A second stage of native copper and subordinate copper sulfide mineralization was synchronous with thrust faulting and interpreted as contemporaneous with late compression of the rift.

In the Portage Lake Volcanics, native copper and native silver mineralization occurs with secondary minerals (typically K-feldspar, chlorite, calcite, prehnite, epidote, quartz, and calcite) in amygdules, interstices, and replacements in basalt, and as open-space filling in interflow sedimentary layers. The timing of native copper mineralization, estimated to be about 1060 Ma (Bornhorst and others, 1988), corresponds to faulting related to the late compression of the rift. It is probable that the initiation of compression resulted in faults and fractures that created pathways along which hot, copper-mineralizing fluids generated in hotter and deeper parts of the rift were able to move buoyantly upwards and be focused into dipping flow tops and sedimentary layers (Bornhorst, 1992). More than 5 billion kilograms of native copper and 0.5 million kilograms of native silver were produced from the Keweenaw Peninsula mining district (Fig. 2.3) from the 1840's until the 1960's.

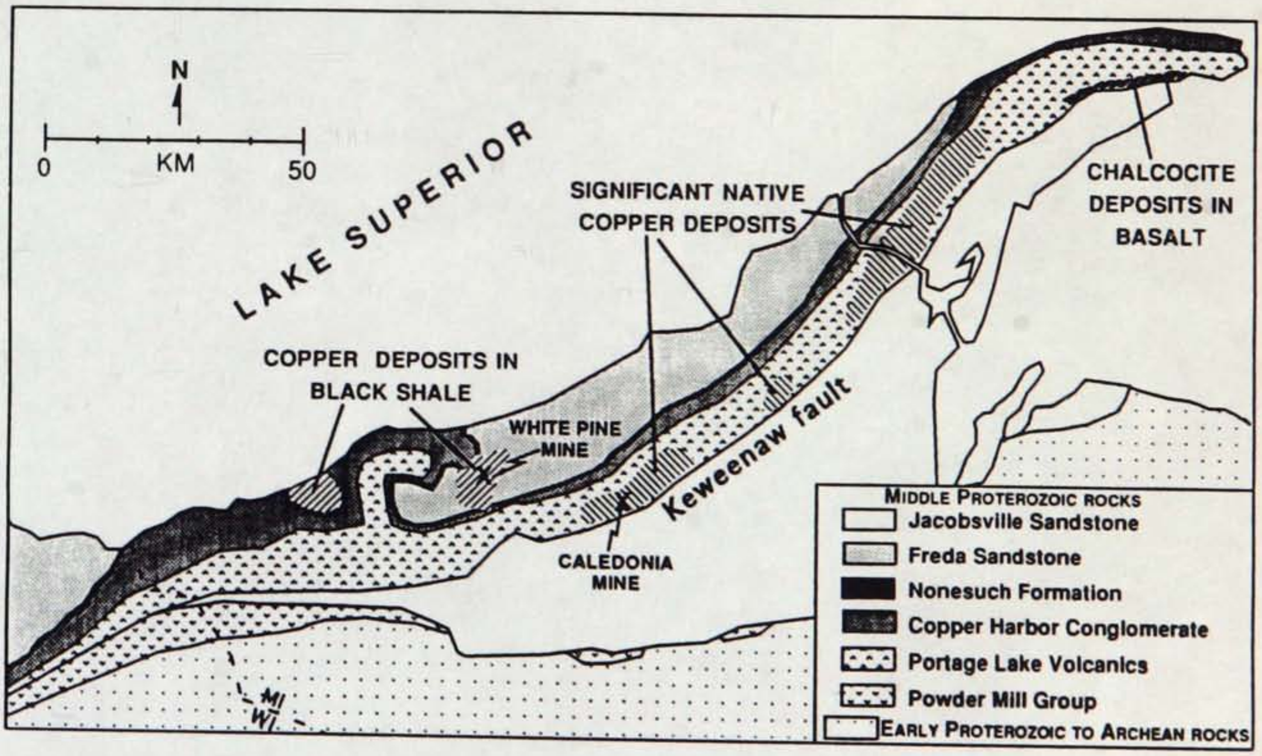

Figure 2.3: Generalized geological map of western Upper Michigan and location of principal mineral deposits.

Small, high-grade, basalt-hosted chalcocite deposits are confined to an area about 13 km long and 2 km wide within and near the base of the Portage Lake Volcanics at the eastern end of the Keweenaw Peninsula (Broderick and others, 1946) (Fig. 2.3). Mineralization occurs as roughly stratabound chalcocite in brecciated and amygdular basalt flow tops, and as chalcocite and very rarely native copper in veins and tension fractures in basalt and andesite dikes. The deposits contain a total of about 7 million tons of ore with an average grade of 2.3% Cu. The largest of the deposits, 543-S, is estimated to have minable reserves of 1.1 million tons grading 4% copper and is scheduled for development in 1995 (Northern Miner, v. 80, no. 41).

FIELD TRIP 2
DAY 1

STRATIGRAPHY AND STRUCTURE OF THE MIDCONTINENT RIFT BETWEEN DULUTH AND THE PORCUPINE MOUNTAINS

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The first day of this field trip will travel across northern Wisconsin and into western Upper Michigan, crossing from the north limb of the Midcontinent Rift to the south limb, and emphasizing the stratigraphy and structure of the rift. Because crustal-scale thrusting shortly after formation of the rift created thick monoclinal successions of vertical to steeply dipping strata along from the Wisconsin-Michigan shore of western Lake Superior, it is possible to visit a nearly complete cross-section of the rift in this area. On the first day of the field trip we will be able to see only a few parts of the rift, but several of the major stratigraphic features will be highlighted. These include in approximate sequence from oldest to youngest: 1) exposures of clastic sediments deposited in prerift sag basins and the first basalt flows that mark the onset of more than 15 m.y. of volcanism in the rift (Stop 1-3); 2) a recent discovery of significant copper sulfide mineralization in basalts of northern Wisconsin (Stop 1-2); 3) the eroded interior of a large central volcano that developed late in the volcanic stage of the rift (Stop 1-5); 4) sediments deposited in syn-rift extensional basins and associated volcanics erupted during the waning stages of rifting (Stop 1-4, Stop 1-6); and 5) quartz-rich sediments deposited in post-rift thermal subsidence basins (Stop 1-1). In addition, Stop 1-1 will be at an exposure of a high-angle reverse rift-bounding faults.

FIELD TRIP 2—DAY 1

Field Stop Descriptions

STOP 1-1: DOUGLAS FAULT, CHENGWATANA VOLCANICS, AND ORIENTA SANDSTONE OF THE BAYFIELD GROUP

Location: Amnicon Falls State Park. From Duluth, proceed on Route 2 to near Wentworth, Wisconsin, and follow signs to State Park and then to parking area for falls. **Note: rock collecting and hammering are not permitted in the park.** South Range, Wisconsin 7 1/2" quadrangle (T48N R12W, Sect. 29, NE of SE).

Duration: 30 minutes

Description: The first stop displays one of the last tectonic events related to the Midcontinent Rift. In this area of northern Wisconsin, the central part of the rift consists of fault-bounded blocks of volcanic rocks, chiefly basalts with minor rhyolites, flanked by half-graben basins filled with younger clastic sediments (Chandler and others, 1989). The bounding faults, including the Douglas fault on the northern limb of the rift, and the Lake Owen fault along the southern limb, are steeply dipping, reverse faults with basalts on the upthrown side. The Douglas fault, which brings the older Chengwatana Volcanics over the younger Orienta Sandstone of the Bayfield Group, is exposed in the Amnicon River. The Douglas fault initially formed as a normal fault during the extensional phase of the rift, which lasted from about 1100 Ma to about 1094 Ma. A compressional event dated at approximately 1060 Ma (Cannon and others, 1993b) resulted in reversing the sense of motion along the Douglas fault, and the development of the central horst.

The Orienta Sandstone is the arkosic lowest member of the Bayfield Group. The Bayfield Group also contains the quartzose Devils Island Sandstone, and the feldspathic Chequamegon Sandstone. The three units of the Bayfield Group represent the waning sedimentation during the last stages of tectonism in the rift. All three formations were deposited in fluvial or lacustrine environments and each preceding unit may have provided a source of material for the successive unit, resulting in nearly pure quartz sandstones. The Chengwatana Volcanics are high alumina tholeiitic flood basalts probably equivalent in age to the Portage Lake Volcanics of Michigan and Wisconsin, which erupted between about 1096 Ma and 1094 Ma (Davis and Paces, 1990). The basalts erupted from fissures probably near the center of the rift, which is to the south of this area, and spread laterally towards the margins of the central graben.

The Douglas fault is exposed at the base of the waterfall in the Amnicon River, where older Chengwatana Volcanics, dipping between 30° and 40° southeast, overlies the younger Orienta Sandstone. The fault is defined by a zone of fault gouge and breccia. The Orienta Sandstone is nearly vertical near the fault, but the dip decreases away from the fault, becoming nearly horizontal within 30 m. This contact is the northwestern margin of the central graben of the Midcontinent Rift, called the St. Croix horst in northern Wisconsin. The reverse displacement on the Douglas fault is the result of the regional compressional event that reactivated normal, listric faults developed during the extensional phase of the rift.

Upstream from falls is an exposure of reddish gabbro that was intruded into the basalt. The gabbro that is thought to be coeval with the Duluth Complex in Minnesota and the Mellen Intrusive Complex in Wisconsin (Paces and Miller, 1993; Zartman and others, 1995).

STOP 1-2: SULFIDE MINERALIZATION IN CHENGWATANA VOLCANICS

Location: South Range quarry. Just before the split near Amnicon Falls State Park between Rt. 2 heading east and Rt. 53 heading south, turn south and then west on County Road E. Proceed west about 1.7 miles and turn south onto Sam Anderson Road (crossing a railroad grade within 0.25 miles). From turnoff, proceed about a mile to bend in road (around outcrop). Turn west onto well-graded gravel road and proceed for 0.2 miles. Park and walk south along dirt road to quarries. South Range, Wisconsin 7 1/2" quadrangle (T47N R13W, Sect. 1, NE of NE).

Duration: 30 minutes

Description: Copper mineralization in the Midcontinent Rift is dominated by the occurrence of native copper in basalts and interflow sediments in Michigan (Day 2B and 3) and by copper in reduced facies in the Nonesuch Formation at the White Pine deposit (Day 2A). A third type of mineralization, copper sulfides, recently has been discovered in several basalt quarries in the north limb of the Chengwatana Volcanics in northern Wisconsin, and may have important implications for the regional metallogeny of the rift. The style of mineralization in these basalt quarries is analogous to native copper mineralization in Michigan in that it is associated with alteration of permeable regions in the flow tops, but is mineralogically distinct. The occurrence of copper-iron sulfides rather than native copper or chalcocite requires high sulfur activity in the mineralizing fluid; an anomalous feature for the typically sulfur-poor basalts of the MCR. Preliminary sulfur isotope measurements are near 0‰, suggesting an igneous source for sulfur in this sulfide mineralization.

A quarry cut into the south face of a basalt ridge exposes a south-dipping sequence of typical massive flows and amygdular and brecciated flow tops. At least two flow tops are exposed, both are highly altered to epidote, imparting a distinctly greenish color in contrast to the black unaltered basalt in flow interiors. The sulfides are dominantly chalcopyrite and bornite with minor pyrrhotite, and occur in amygdules and open space filling with a variety of minerals, including epidote, quartz, chlorite, and chalcedony. Chalcopyrite appears to be most abundant in the lower parts of altered flow tops where epidote alteration is less severe, rather than in the more thoroughly altered rock in the upper part of the flow tops. This relationship can been seen especially well near the east face of the quarry where a mass of strongly epidotized rock contains little or no sulfide minerals, but is surrounded by a halo of mineralization about one meter wide. This relationship of copper sulfides adjacent to strongly epidotized rock seems fairly common in the region, although characterization of this recently discovered mineralization is still in its early phase.

STOP 1-3: BESSEMER QUARTZITE AND SIEMENS CREEK VOLCANICS, POWDER MILL GROUP.

Location: North of Ironwood. From the town of Ironwood, Michigan, proceed east along Rt. 2. Turn north on Section 12 Rd. (about half a mile east of the Ottawa National Forest ranger station). Proceed about 0.6 mile. Outcrop to east of road. North Ironwood 7 1/2" quadrangle (T46N R47W Sect. 12 SW of SE).

Duration: 30 minutes

Description: In the Lake Superior region, the earliest deposits of the MCR consist of several thin, prevolcanic quartz sandstone units. These fluvial to lacustrine sediments were deposited on an apparently peneplaned Archean to Lower Proterozoic terrane in an incipient sag basin developed during initial subsidence of the rift (Ojakangas and Morey, 1982). In northern Wisconsin and western Upper Michigan, this sandstone unit, the Bessemer Quartzite, unconformably overlies the Early Proterozoic Tyler Formation graywacke. Excellent cross-bedding is preserved in the quartzite that suggests deposition in a body of standing water affected either by tidal processes or by opposing longshore currents (Ojakangas and Morey, 1982). Flood basalts of the Siemens Creek Volcanics of the Powder Mill Group, the oldest volcanic rocks of the MCR in the western Lake Superior area, were erupted directly onto the unconsolidated sand of the Bessemer. .

A low rock knob in a pasture east of the county road shows the contact between the upper part of the Bessemer Quartzite and the overlying basal basalt flow of the Siemens Creek, a contact that marks the beginning of more than 20 m.y. of volcanism in the rift (Fig. I.3). Local deformation in the quartzite, incorporation of sand into the overlying basalt, and the development of pillowed structures in the lowermost part of the basalt section indicate that basalts poured out subaqueously onto unconsolidated sands during the first stages of eruption. With continued eruption, the developing volcanic pile quickly emerged from the shallow water; nearly all of the remaining volcanic eruptions in the Keweenawan Supergroup were subaerial. The Siemens Creek is composed predominantly of thin flows of basalt and minor andesite. Average flow thickness is about 3 m, but the basal flow is about 50 m thick at this locality. Unlike some of the basal flows along strike to the west, this flow does not contain clinopyroxene phenocrysts. Chemical analyses of the basal flows are broadly similar to the younger high-alumina main-stage basalts of the Portage Lake Volcanics (Table 2.1).

STOP 1-4: NONESUCH FORMATION AND FREDA SANDSTONE, ORONTO GROUP.

Location: Presque Isle State Park. Follow Rt. 2 east to Wakefield. At the stoplight in Wakefield, turn north onto M28. About 1.2 miles later turn north on County road 519 and proceed for about 16 miles to the entrance to Presque Isle State Park. Follow signs to parking area for falls and picnic area. Park vehicle and then follow signs for trail to falls. Stay on the west side of the river (don't cross footbridge across Presque Isle River). Go to beach to see contact between the Nonesuch Formation and the overlying Freda Sandstone. **Note: rock collecting and hammering are not permitted in park.** Tiebel Creek quadrangle (T50N R45W, Sect. 30, NE of NE).

Duration: 1 hour, 15 minutes; lunch stop

Description: The upper portion of the Nonesuch Formation and the base of the Freda Sandstone of the Oronto Group are well exposed in the gorge of the Presque Isle River near the mouth of the river and along the shore of Lake Superior to the west (Fig. 2.4). The rocks exposed here are on the northeast limb of the Presque Isle syncline, a gentle northwest-plunging fold, that probably developed in response to the same compression that created the reversed movement along the rift-bounding faults. Dips range from nearly flat to about 10° SW. The Nonesuch Formation is distinguished from other sedimentary units of the Keweenawan Supergroup by the predominance of gray, green or black, fine-grained siltstones, fine-grained sandstones, mudstones, and more rarely true shales. Sedimentation in the Nonesuch is dominated by numerous graded beds and fining-upward sequences. On a small scale, normally graded sequences are common in units from a few centimeters to a few meters thick. Numerous thin micaceous and chloritic graded sandstones and siltstone tend to separate along parting lineations, symmetrical and asymmetrical ripple marks, rib and furrow structures, or occasional mud-cracked surfaces.

The lower part of the Nonesuch, which is not exposed in this area, is strongly mineralized with copper and contains a fine-scale stratigraphy directly correlatable with the stratigraphy at the White Pine mine (Fig. 2.4, see Day 2A), indicating that sedimentary conditions during Nonesuch deposition were very uniform over the entire region surrounding the Porcupine Mountains. The upper part of the Nonesuch displays a zone in which dark gray laminated and small-scale cross-bedded siltstones and sandy mudstones are interbedded with medium- to coarse grained reddish brown sandstones, presenting a general coarsening-upward trend to the contact with the Freda Sandstone. The approximate upper contact of the Nonesuch is placed where the reddish-brown, oxidized sandstones becomes dominant.

The Freda Sandstone is well exposed just west of the mouth of the Presque Isle River and along the shore of Lake Superior to the west. The Freda Sandstone is a thick bedded, fine- to medium-grained reddish-brown sandstone. Locally, the Freda is laminated with abundant cross-beds and scours, and contains small pebble zones. The Freda marks a return to fluvial redbed deposition following the lacustrine deposition of the underlying Nonesuch Formation.

STOP 1-5: PORCUPINE VOLCANICS, BERGLAND GROUP

Location: Summit Peak, Porcupine Mountains Wilderness State Park. At the entrance to Presque Isle State Park, turn onto the South Boundary Road and proceed northeast for almost 13 miles. This road follows the lower boundary of the state park, which is maintained as a wilderness area with access only by hiking. The access road to the Summit Peak Overlook turns north from the South Boundary Road and is well marked. Proceed to the parking area at the end of the road. **Note: rock collecting and hammering are not permitted in park.** Tiebel Creek quadrangle (T50N R44W, Sect. 11, NE of NW).

Duration: 1 hour, 15 minutes

Description: The Porcupine Volcanics of the Bergland Group are a sequence of rhyolite, andesite, and basalt forming a thick unit centered near the Porcupine Mountains and thinning both east and west (Fig. 2.4). The overlying Copper Harbor Conglomerate of the Oronto Group thins in response in lateral variation in the thickness of the Porcupine Volcanics, suggesting that the two units are nearly contemporaneous. The restricted extent, paleogeomorphology, and diverse lithologies of the Porcupine Volcanics, suggest that the rocks represent a shield volcano that erupted from a central vent region, centered in the Porcupine Mountains. This is in sharp contrast to the more typical flood basalt eruption style of both the Siemens Creek Volcanics (Stop 1-3) and the Portage Lake Volcanics that directly underlie the Porcupine Volcanics.

The observation tower at the top of Summit Peak provides a panoramic view of the Porcupine Mountains and is the highest point in Porcupine Mountains Wilderness State Park and one of the highest points in the state of Michigan. To the south, the highlands are underlain by the Portage Lake Volcanics and Porcupine Volcanics along the main monocline in the Keweenawan Supergroup. The lowlands to the southeast are underlain by rocks of the Oronto Group in the east-plunging Iron River syncline. To the north, the interior of the park extends over rugged topography to the shore of Lake Superior. The smelter stack at the White Pine mine is visible to the east.

Excellent exposures of typical intermediate and felsic units of the Porcupine Volcanics can be seen along the Beaver Lake Creek Trail about 0.5 miles from the Summit Peak parking lot. The rocks exposed on a hill slope to the north of the trail include in

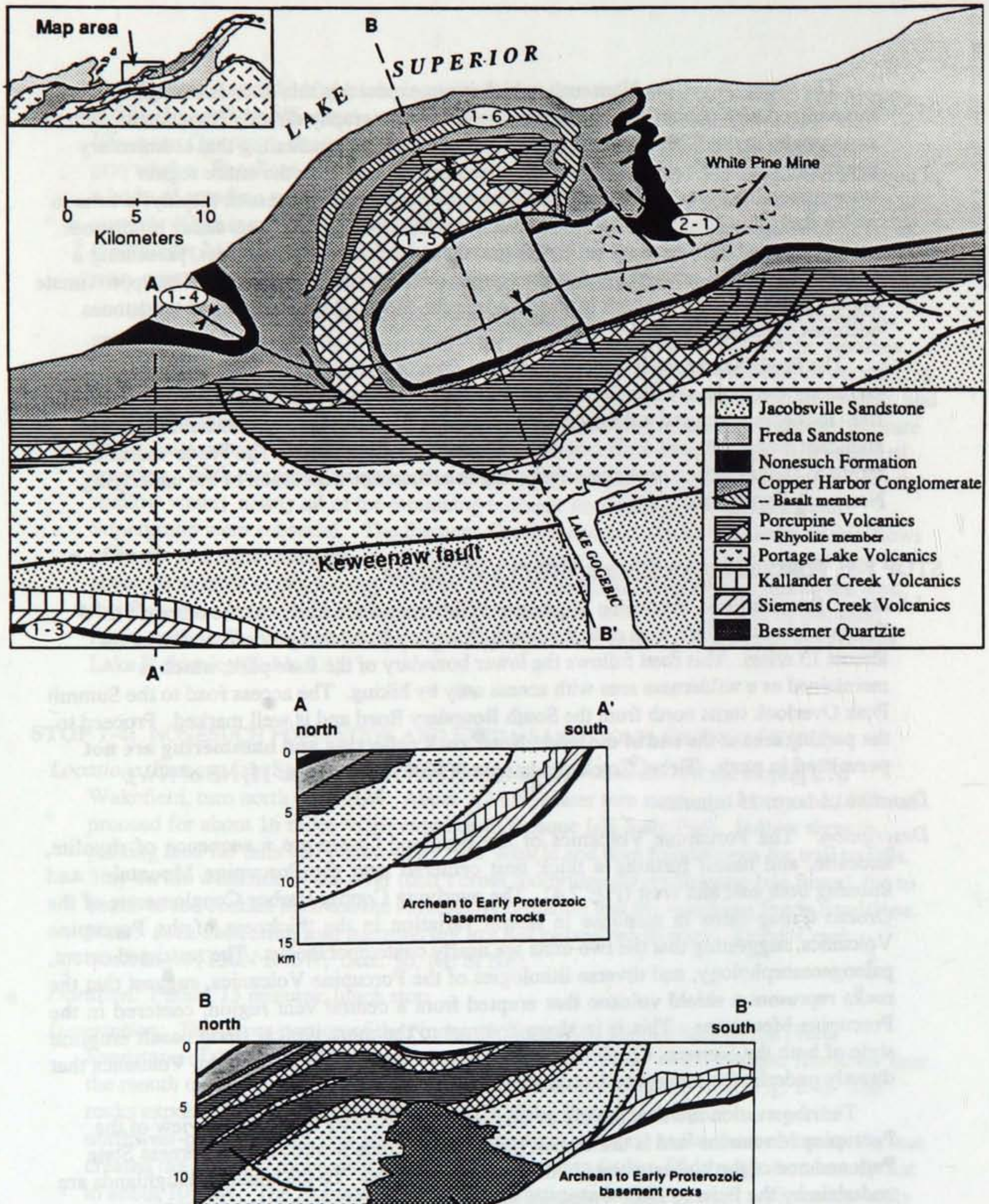

Figure 2.4: Generalized geologic map and cross-sections of the Porcupine Mountains area showing the location of field trip stops (modified from Cannon and others, 1992). All rocks are part of the Middle Proterozoic Keweenawan Supergroup.

stratigraphic order, sparse outcrops of intermediate to mafic rocks in the creek bed, overlain by a vesicular siliceous andesite, which in turn is overlain by a coarse rhyolite breccia or debris flow. The breccia is clast supported, with clasts ranging in size from nearly 1 m to less than 1 cm. Overlying the breccia is a medium-grained basalt flow, which in turn is overlain by a massive, aphanitic, microspherulitic rhyolite that caps the hill.

STOP 1-6: BASALT FLOWS IN COPPER HARBOR CONGLOMERATE, ORONTO GROUP.

Location: Lake of the Clouds, Porcupine Mountains Wilderness State Park. Drive to the dead-end of Michigan Highway 107 to parking area overlooking Lake of the Clouds. Along the road are several exposures of the Copper Harbor Conglomerate. Take XXmi. hiking trail to Lake of the Clouds overlook. **Note: rock collecting and hammering are not permitted in park.** Carp River 7 1/2" quadrangle, Michigan (T51N R43W, Sec. 21, SW of NE).

Duration: 30 minutes

Description: The overlook is along the south escarpment of a high ridge supported by a series of north-dipping basalt flows within the Copper Harbor Conglomerate of the Oronto Group. These basalts represent the very final stages of volcanism in the rift. From the overlook, the low area south of the ridge, including the Lake of the Clouds, is underlain by sandstone and siltstone, with the small ridges held up by a few small basalt flows within the lower section of the Copper Harbor Conglomerate. The higher, prominent peaks farther south are knobs of rhyolite of the Porcupine Volcanics of the Bergland Group (Stop 1-5).

Towards the east end of the overlook over the small retaining wall, a large glaciated surface beautifully exposes a series of thin basalt flows, that dip northwest into Lake Superior. Individual flows can be readily identified by chilled vesicular bases and by rubbly or vesicular flow tops. Replacement of basalt by epidote gives several of the flows a greenish color.

FIELD TRIP 2
DAY 2A (morning)

UNDERGROUND GUIDE TO THE WHITE PINE MINE, MICHIGAN

Jeff L. Mauk
University of Auckland, New Zealand

Safety

The personnel at the mine are very concerned about the safety of visitors, and ask you to be cautious while underground. You may sample while underground, but please be careful! The ribs (walls) and back (roof) of the mine workings tend to develop cracks over time, and there may be a significant amount of loose rock on the ribs or back. Sampling in an underground mine is far more dangerous than sampling a road cut.

Everyone must wear safety glasses at all times while underground. Federal law requires that everyone carry a self-rescuer in the mine; its use will be explained by the mine staff before you enter the mine.

Introduction

The White Pine mine is a room-and-pillar mining operation which exploits sediment-hosted stratiform copper mineralization. The mine operates on a single level which follows the contact of the Nonesuch Formation and the Copper Harbor Formation (Fig. 2.5). The ore lies within the basal Nonesuch Formation and locally in the top one to three meters of the Copper Harbor Formation (Fig. 2.6). The mine extends eight km (5 mi) in a north-south direction and eight km (5 mi) in an east-west direction (Fig. 1). Mining began in the area around the portal in 1953 (Fig. 2.5), near the subcrop of the basal Nonesuch Formation, and progressed to depth from there. All of the ore that is removed from the mine is carried out on conveyor belts.

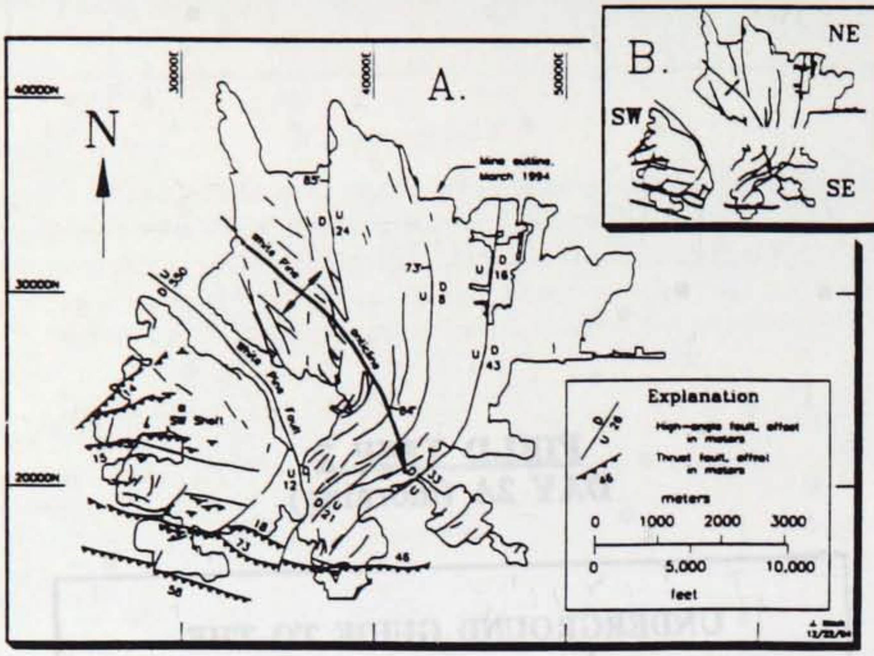

Figure 2.5. Generalized structure map of the White Pine mine. Modified from Mauk and others (1992b). Tick marks represent mine coordinates (in feet).

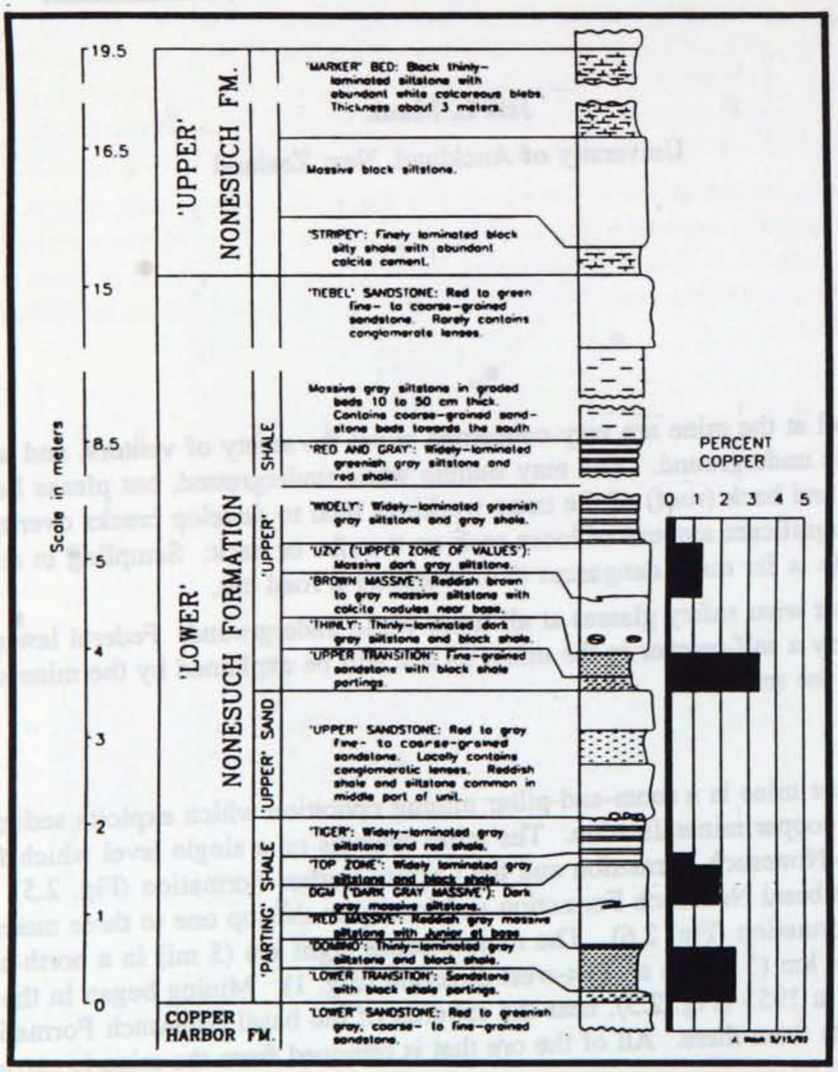

Figure 2.6. Stratigraphic column for the 'lower' Nonesuch Formation and uppermost Copper Harbor Formation at the White Pine mine. Modified from Ensign and others (1968).

Stratigraphy

The Proterozoic Nonesuch Formation consists of alternating siltstones and subordinate shales with local to rare sandstones. It is informally subdivided into two subunits: the 'upper' Nonesuch Formation, which lies between the Freda Sandstone and the base of the 'stripey' shale, and the 'lower' Nonesuch Formation, which lies between the base of the 'stripey' shale and the Copper Harbor Formation (Fig. 2.6). The 'stripey' shale is a distinct, laminated to thinly-bedded shale with mm-scale layers of calcite-cemented siltstone. Between Houghton, Michigan and the Michigan/Wisconsin border (Fig. 2.3), the entire Nonesuch Formation averages approximately 200 m thick, whereas the 'lower' Nonesuch Formation averages approximately 20 m thick.

The 'lower' Nonesuch Formation contains alternating sequences of chemically reduced and chemically oxidized strata; reduced strata are green to gray to black due to pigmentation from chlorite and/or organic matter, whereas oxidized strata are reddish due to pigmentation from diagenetic hematite and lack of organic matter. The 'lower' Nonesuch Formation has been further subdivided into three informal stratigraphic intervals: the 'parting' shale, the 'upper' sandstone, and the 'upper' shale (Fig. 2.6). The 'parting' shale and the 'upper' shale have also been informally subdivided into a sequence of named strata (Fig. 2.6).

Structure

The mine is divided into three structural domains. The northeastern domain contains dominantly extensional, north-south-striking high angle faults, the southwestern domain is characterized by compressional faults, and the southeastern domain is transitional between extensional and compressional regimes. The region is believed to have evolved from an early extensional regime to a later, rift-closing compressional regime (Mauk and others, 1992b).

Mineralization

White Pine is a noted example of sediment-hosted stratiform copper mineralization, but the geologic and mineralization history of the mine area is moderately complex. Mauk and others (1992b) described three episodes of faulting in the mine area and two stages of copper mineralization. The first, or "main-stage" of copper mineralization is the classic sediment-hosted, stratiform copper mineralization described by many workers in the past (e.g. White and Wright, 1954, 1966; Ensign et. al, 1968; Brown, 1971). Second-stage copper mineralization, which is well-exposed in many parts of the southwestern domain of the mine, is spatially and temporally associated with compressional faulting.

Main-stage mineralization This stage accounts for 80-90% of the copper that is recovered from the mine. This predominantly sulfide mineralization consists of disseminated fine-grained chalcocite and native copper within rocks of the basal Nonesuch Formation Copper contents commonly reach 1-3% Cu in the dark, thinly-laminated to massive, carbonaceous sedimentary rocks of the 'lower transition', 'domino' shale, and DGM of the 'parting' shale, and the 'upper transition', 'thinly' shale, and UZV of the 'upper' shale (Fig. 2.6). Intervening strata generally contain less than 0.5% Cu.

At the mine scale, main-stage mineralization extends upward through all basal Nonesuch strata to an abrupt, 1 to 5 cm-thick, peneconcordant "fringe" surface which separates the cupriferous zone from pyritic rocks occupying overlying portions of the Nonesuch Formation. The "fringe" contains, from bottom to top, chalcocite, djurleite ($Cu_{1.96}S$), digenite, bornite, chalcopyrite, and pyrite (Brown, 1971). The abrupt transition from the cupriferous to pyritic zones has been interpreted as the final position of an early diagenetic mineralization front representing the invasion of copper into initially iron sulfide-rich beds at the base of poorly consolidated Nonesuch beds (White, 1960b; White and Wright, 1966; Brown, 1971).

Second-stage mineralization- Exposures in the southwestern domain of the mine have led to recognition of a second-stage of copper mineralization at the White Pine mine. This second-stage mineralization post-dates main-stage mineralization, and apparently records additional copper and sulfur brought into the mine area during compressional faulting (Mauk and others, 1992b; Mauk and Hieshima, 1992). Two characteristics are used to identify second-stage mineralization: (1) strata that are unmineralized elsewhere, such as the 'stripey' shale, locally contain copper mineralization adjacent to compressional faults, and (2) ore-grade strata are locally 20% to 30% higher-grade near thrust and tear faults than in unfaulted regions.

Mauk and others (1992b) recognized four main types of second-stage mineralization (I) sheet copper, (II) bedding-plane-parallel chalcocite veins, (III) stockworks, and (IV) disseminated native copper in the uppermost Copper Harbor Formation Because second-stage copper mineralization commonly occurs as vein-filling material, it is commonly coarser-grained than main-stage copper mineralization. However, in many samples from the southwestern domain of the mine, it is extremely difficult to distinguish whether some disseminated native copper and/or chalcocite is a product of main- or second-stage processes.

Many of the larger thrust and tear faults are not intimately associated with second-stage mineralization; instead second-stage mineralization is most commonly developed around smaller faults, suggesting that fractured envelopes which surround the major structures provided the main conduits for second-stage fluids, not the major structures themselves.

Petroleum

White Pine is a well-known locality for Precambrian petroleum (e.g. Eglinton and others, 1964; Barghoorn and others, 1965; Kelly and Nishioka, 1985; Hieshima and Pratt, 1991), and Precambrian microfossils (e.g. Meinschein and others, 1964; Moore and others, 1969). In the mine area, all seeps emanate from the back of the mine, suggesting that they are derived from source or reservoir units above the ore stratigraphy. In addition, seeps are found in all domains of the mine, regardless of the local structural setting, suggesting that they post-date all recognized faulting events. Mauk and Hieshima (1992) suggest that the most likely source for the seep petroleum was Nonesuch Formation sediments from deeper within the Midcontinent rift; this petroleum may have been generated by burial or hydrothermal maturation, and it migrated into the White Pine area during thrusting. In this scenario, they hypothesize that the petroleum was stored in veins and vugs in fractures, then released during subsequent deformation to form active seeps. This deformation may be either an unrecognized structural event, or mining-induced fracturing of the rocks.

Mining Configurations

There are four basic mining configurations at White Pine which are named for the stratigraphic units that they extract: (1) "full column", (2) "parting shale", (3) "upper shale" and (4) "modified upper shale". This section describes three of these configurations.

"Full column" configuration - This configuration includes all the strata between the top of the Copper Harbor Formation and the "dimple back" in the 'brown massive' siltstone (Fig.

2.7). The heights of "full column" workings are moderately variable, depending primarily on the thickness of the 'upper' sandstone. Where the 'upper' sandstone is about 2 m (6 feet) thick, the mining height is about 4 m (12 feet); but in the southern portion of the mine, the 'upper' sandstone thins to approximately 0.1 m (4 in), and the mining height is only 3.3 m (10 feet) high. Most of the equipment operating underground requires a clearance of about 3.3 m (10 feet), so that is the minimum height of the mine workings.

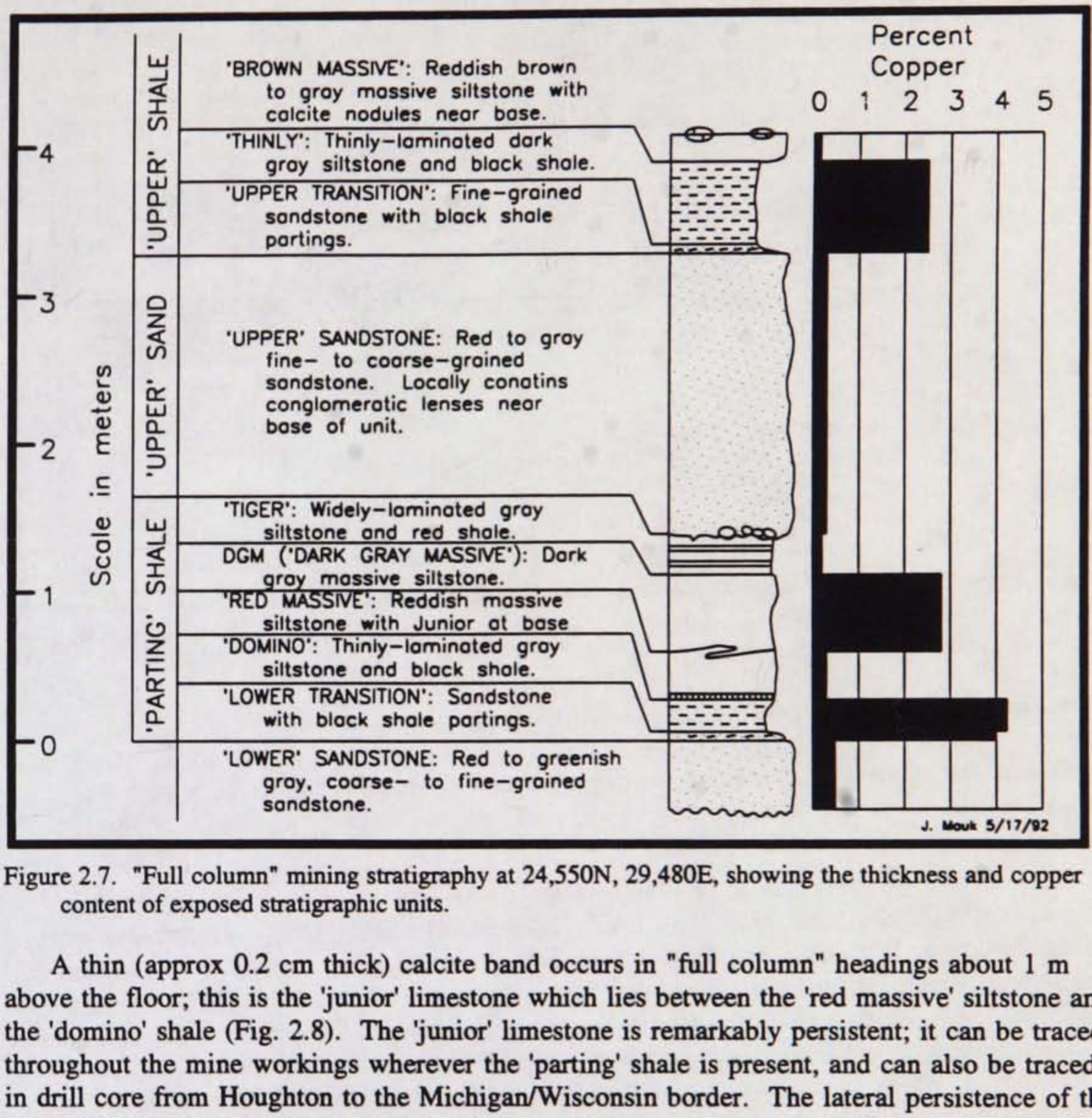

Figure 2.7. "Full column" mining stratigraphy at 24,550N, 29,480E, showing the thickness and copper content of exposed stratigraphic units.

A thin (approx 0.2 cm thick) calcite band occurs in "full column" headings about 1 m above the floor; this is the 'junior' limestone which lies between the 'red massive' siltstone and the 'domino' shale (Fig. 2.8). The 'junior' limestone is remarkably persistent; it can be traced throughout the mine workings wherever the 'parting' shale is present, and can also be traced in drill core from Houghton to the Michigan/Wisconsin border. The lateral persistence of the 'junior' limestone strongly suggests that it is a synsedimentary calcite band.

Some "full column" headings have a well-developed "dimple back". The dimpling reflects calcite nodules that occur along a parting in the otherwise relatively homogeneous 'brown massive' siltstone. Calcite nodules that form the "dimple back" are typically 8-10 cm in diameter, lenticular in cross-section, and 3-4 cm thick at their thickest portion. The homogeneity of the 'brown massive' siltstone makes it a good back; it is far more resistant to cracking and peeling than the laminated to thin-bedded siltstones and shales in the rest of the 'lower' Nonesuch Formation

"Parting shale" mining configuration-This configuration provides a good opportunity to examine the 'parting' shale and 'lower' sandstone (Fig. 2.8). The exposures of the 'lower' transition are particularly good, because the entire unit is typically exposed. The back in "parting shale" mining configurations consists of the 'upper' sandstone, which, because of its extreme competency, is the strongest back in the mine.

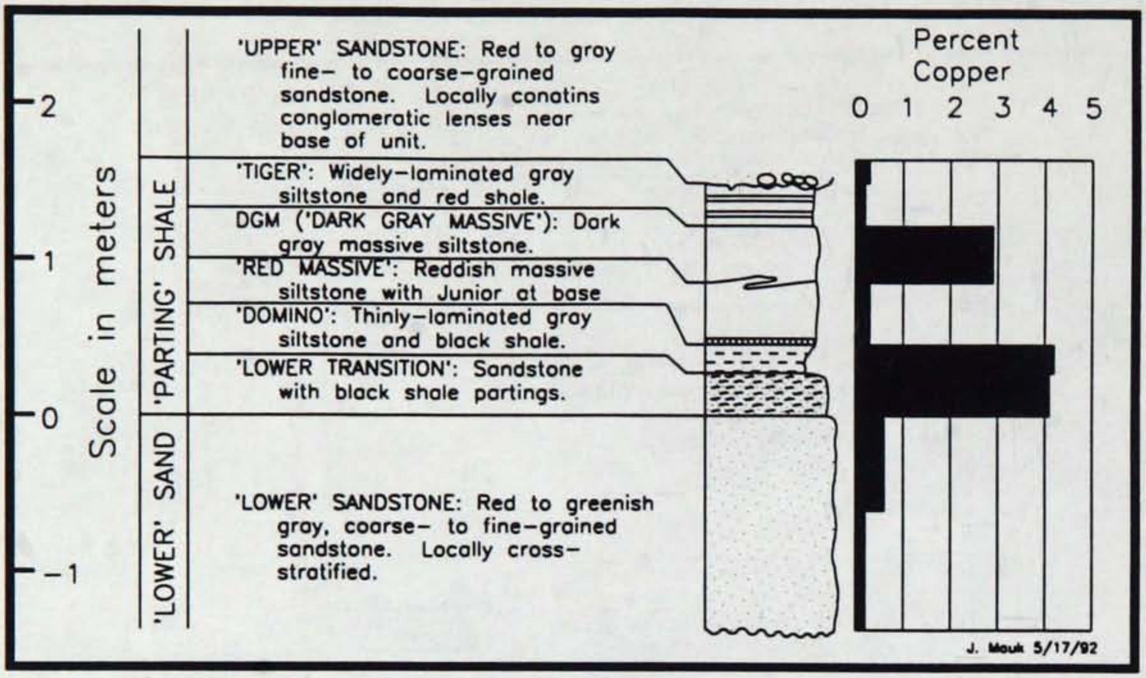

Figure 2.8. "Parting shale" mining stratigraphy at 24,500N, 29,720E, showing the thickness and copper content of exposed stratigraphic units.

"Modified upper shale" configuration- "Modified upper shale" configurations occur in the southern portion of the mine, where the 'upper' shale rests disconformably on the 'lower' sandstone; the 'upper' sandstone and 'parting' shale are absent due to nondeposition (Fig. 2.9). This configuration contains the 'lower' sandstone, the 'upper' transition, the 'thinly' shale, and the 'brown massive' siltstone.

In some places, the 'lower' sandstone contains dense disseminations of native copper (Nishio, 1919; Brady, 1960; Hamilton, 1967; Kelly and Nishioka, 1985). Associated with the native copper is chlorite and pyrobitumen (altered petroleum) cement; the pyrobitumen may reflect petroleum that was altered during the formation of native copper (Kelly and Nishioka, 1985; Mauk and Hieshima, 1992). Brady (1960) and Hamilton (1967) observed a positive correlation between volume percent native copper and volume percent organic matter, based on point counts of mineralized 'lower' sandstone samples from the northeast domain of the mine, near the White Pine fault. All known occurrences of disseminated native copper in the 'lower' sandstone are spatially associated with thrust faults or presumed tear faults, suggesting that this type of occurrence is related to second-stage mineralization.

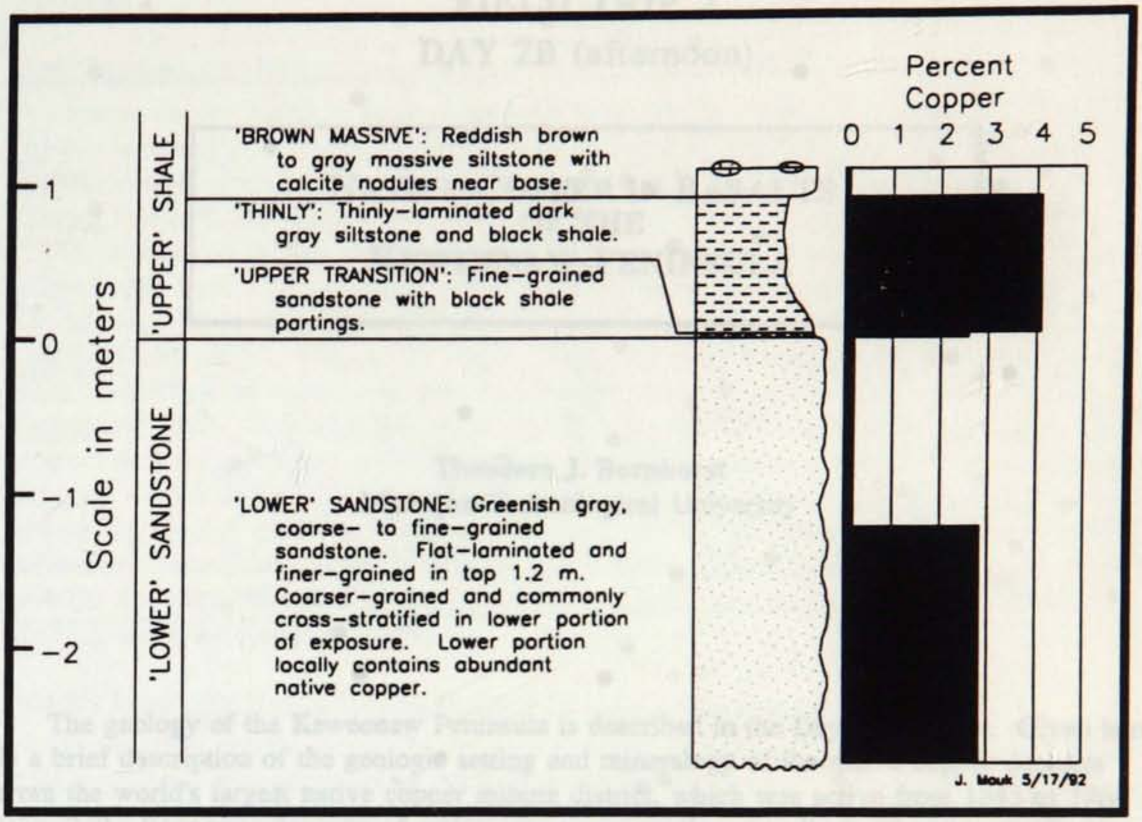

Figure 2.9. "Modified upper shale" mining stratigraphy at 19,010N, 28,755E

Acknowledgements

I thank Dick Andrews, Alex Brown, C. Stewart Eldridge, Glen Hieshima, Eileen Ho, Bill Kelly, Bill Nelson, and Bob Seasor for their contributions to my understanding of the geology of White Pine.

FIELD TRIP 2

DAY 2B (afternoon)

NATIVE COPPER IN BASALTS OF THE KEWEENAW PENINSULA

Theodore J. Bornhorst
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The geology of the Keweenaw Peninsula is described in the Day 3 overview. Given here is a brief description of the geologic setting and mineralogy of the native copper deposits from the world's largest native copper mining district, which was active from 1845 to 1968 (Fig. 2.3). Economic deposits of native copper are stratigraphically restricted to the Portage Lake Volcanics. Stops 2-2 and 2-3 will provide the opportunity to have underground and surface views of dipping basalt lava flows and altered and mineralized flow tops (Fig. 2.10). Deposits of chalcocite also occur within the Portage Lake Volcanics near the tip of the Keweenaw Peninsula. These economically minor occurrences are under investigation. The description provided below is abstracted from Bornhorst (1992 and in press).

Introduction to Ore Bodies

Native copper occurs in brecciated and amygdaloidal flow tops (58.5% of production), interflow conglomerate beds (39.5% of production), and cross vein systems (about 2% of production). The four largest deposits in the district produced 85% of the 5 billion kg total district production at an average grade of about 2%.

Brecciated amygdaloidal flow tops are much more common hosts for native copper deposits than unbrecciated amygdaloidal flow tops. Flow top deposits are stratigraphically sandwiched between a footwall of barren basalt in the massive interior of the same flow and a hanging wall of barren basalt in the massive interior of the succeeding flow. Although, the distribution of contained native copper is irregular, usually native copper is more abundant near the top and bottom of the brecciated flow top.

Interflow conglomerate beds make up only a small volume of the Portage Lake volcanics (<5%) but they host comparatively large amount of native copper (~40% of district production). Deposits occur as lenticular beds with hanging wall of massive basalt and footwall of lava flow top which commonly contains sand and silt. Usually native copper is concentrated along stratigraphic bands (0.5 to 5 m thick) and rich bands tend to jump from one stratigraphic position to another within the same conglomerate (Weege and others, 1972).

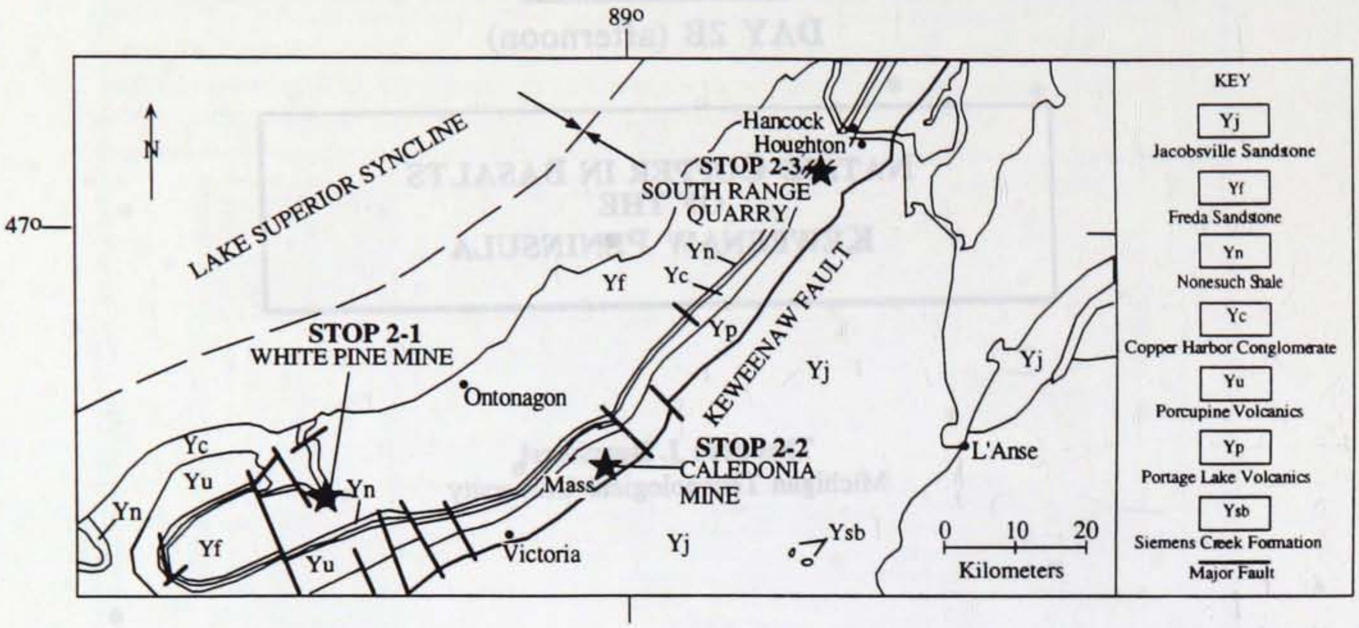

Figure 2.10: Geologic map showing the location of Day 2 field trip stops.

The first mines in the district were developed on veins which tend to cut bedding at high angles, vein deposits and are of slight economic importance in the district. Distribution of native copper in veins is more erratic than either flow top or conglomerate lodes and can contain masses weighing many tons. Veins hosting native copper also exist within stratabound lode deposits.

Ore and Gangue Minerals of Native Copper Deposits

Native copper represents over 99% of the metallic minerals in the mined orebodies of the Keweenaw Peninsula. Persistent small quantities of native silver (less than 0.1%; White, 1968) accompany the native copper. Most of the native copper carries a small amount of arsenic in solid solution (typically less than 0.2% arsenic in total copper+silver+ arsenic; Broderick, 1929). Copper-nickel arsenides occur in veins that are paragenetically late (Butler and Burbank, 1929; Moore, 1971). Within the native copper deposits chalcocite, also paragenetically late, occurs as small veins cutting flow top deposits and as coatings on joints with calcite in conglomerate deposits (White, 1968).

Flow tops and interflow sedimentary rocks were altered pervasively by hydrothermal fluids producing low-temperature metamorphic mineral assemblages. The minerals occur as amygdule and vein fillings and as whole rock replacements in the most permeable units. While flow tops are intensively altered, massive interiors of lava flows are much less altered and hydrothermal alteration is limited to the vicinity of faults and fractures. The interiors of lava flows acted as aquicludes in the paleohydrothermal system.

A close relationship in time exists between native copper mineralization and secondary minerals formed during burial metamorphism/alteration of the Portage Lake volcanics, although individual deposits may not exactly follow the district-wide paragenesis. Metamorphic zoning based on distribution of amygdule-filling minerals, equivalent to zeolite and prehnite-pumpellyite facies, varies vertically and laterally within the Portage Lake volcanics.

Age Constraints of Native Copper Deposits

Native copper found in both stratabound lodes and in veins is accompanied by the same gangue minerals (Butler and Burbank, 1929; Broderick, 1931; White, 1968) indicating contemporaneous epigenetic deposition. Native copper mineralization is younger than the Copper Harbor Conglomerate which hosts rare veins of calcite and native copper. White (1968) interpreted the age of native copper mineralization as after deposition of much or all of the Freda Sandstone and undetermined relationship with respect to the Jacobsville Sandstone. Broderick (1929, 1931), Butler and Burbank (1929), Broderick and others (1946), White (1968), Weege and others (1972) and other geologists have pointed out the close connection between native copper mineralization and deformation related to the Keweenaw Fault. Bornhorst and others (1988) used the Rb-Sr method on amygdule-filling microcline, calcite, epidote, and chlorite to determine the absolute age of mineralization as between 1060 and 1047 Ma (+/- ~20 Ma), consistent with an approximate age of Keweenaw reverse faulting of 1060 Ma (Cannon and others, 1990). Thus, both field relationships and radiometric dating are consistent with an age of native copper mineralization of about 1060 to 1050 Ma, some 30 Ma after rift-filling volcanism but contemporaneous with reverse faulting along the Keweenaw fault.

Genetic Model for Native Copper Deposits

The favored genetic model for the native copper deposits of the Keweenaw Peninsula calls upon epigenetic ore-bearing fluids related to burial metamorphic processes (Stoiber and Davidson, 1959; White, 1968; Jolly, 1974). Widespread distribution of native copper in Keweenawan basalts throughout the rift is consistent with regional ore fluids related to burial metamorphic processes. The age of mineralization, some 30 Ma after most Keweenawan magmatic activity, suggests direct magmatic processes are not important for ore genesis. Stable isotope data are consistent with burial metamorphism but cannot rule out magmatic hypotheses (Livnat, 1983). Although evidence is not conclusive, cumulative arguments favor formation of ore fluids during burial metamorphism of the rift rocks.

The source of native copper is likely the rift-filling basalts. Dissolution of only a few ppm copper from basalts within the edge of the rift yields more than suficient adequate amounts of copper (Bornhorst, in press). Burial metamorphism of rift-filling basalts at temperatures of 300°C to 500°C could result in the generation of a Cu-rich ore fluid. Degassing of sulfur from subaerial erupted lava flows would result in low residual sulfur contents; burial metamorphism of this low sulfur basaltic source rock would yield a low sulfur ore fluid. These low sulfur fluids mineralized low-sulfur basalts so that native copper would be the favored copper species.

The fundamental control on the movement of ore fluids to sites of deposition is permeability. Primary permeability includes brecciated and vesicular lava flow tops and interflow sedimentary rocks separated by variably impermeable massive flow interiors. Thin, relatively isolated aquifers did not favor simple up dip movement of ore fluids. Secondary permeability provided by the network of faults/fractures produced during late compression (reverse movement along the Keweenaw fault) integrated the plumbing system allowing for upward movement and focusing of large volumes of ore fluids. The Keweenaw fault is relatively mineralized whereas subsidary perpendicular structures are closely connected with mineralization (Bornhorst, 1992). NW-SE directed compression could have caused the Keweenaw Fault to be under tight compression as compared to perpendicular structures.

The Midcontinent rift is not particularly unusual in either igneous activity or in geothermal gradient (Hutchinson and others, 1990). Low grade burial metamorphism/ hydrothermal alteration of mafic volcanics is observed throughout the world, yet native copper deposits of the Keweenaw Peninsula are unique. Significant late-stage reverse faulting distinguishes the Midcontinent rift system from other flood basalt provinces. The superposition of this deformation event on temporally available burial metamorphic fluids is likely the critical component in the genetic model of the native copper deposits.

Field Stop Descriptions

STOP 2-2: CALEDONIA NATIVE COPPER MINE, PORTAGE LAKE VOLCANICS

Location: Caledonia Mine, approximately 2 miles south of Greenland, Michigan. Rockland 7.5' quadrangle (T50N R39W, Sec 12).

Duration: 2.5 hrs.

Description: The Caledonia Mine, owned by Red Metal Explorations, is located near Mass, Michigan within an area of native copper deposits southwest of the major deposits of the Keweenaw Peninsula (Fig. 2.3). The Caledonia Mine produced about 3 million kg of native copper from 1863-1881 and 1951-1958. Since 1985, Red Metal Explorations has tested the mine and recovered native copper and other minerals which are sold as specimens. The description provided below is based on Bornhorst and Whiteman (1992) and on-going studies.

The Caledonia Mine intersects the Butler and Knowlton flow tops. These flows are part of the Evergreen Series which consists of tholeiitic basalt lava flows up to 210 m in total thickness. The Caledonia Mine includes a crosscut that connects with the Nebraska adit exposing flows of the upper part of the Evergreen Series. The Evergreen flows are glomeroporphyritic as compared to aphyric and ophitic underlying and overlying flows. The flows dip about 45° northwest towards Lake Superior. The flow tops trend northeast-southwest within the Caledonia Mine at the southwest end of the Mass area (Fig. 2.11). Between the Mass and Adventure Mines, to the Northeast, the strike of the bed makes a sharp change of about 35° from N70°E to N35°E. The native copper deposits are situated about this anticline. Many veins and faults with small displacement are found in the mines, but only a few faults have displacements of more than 1 m so the structure is simple. Many veins are in tension fractures related to folding. Some veins are parallel to strike but dip in opposite directions.

For most of its length, the adit at the Caledonia Mine follows the brecciated amygdaloidal top of the Knowlton flow. Stopes raised from the adit are on the Knowlton lode. The average thickness of the Knowlton lode (flow top) is about 2.5 m, but it locally thickens to more than 6 m. The most abundant minerals filling amygdules and spaces between fragments are subequal amounts of quartz, calcite, epidote and red K-feldspar with lesser amounts of prehnite, pumpellyite, and chlorite. Native copper is present in small amounts and native silver in much lesser amounts. Distribution of secondary minerals is highly variable; secondary minerals may be absent or present in various parts of the lode. In some areas the occurrence of secondary minerals is banded. Paragenetically, K-feldspar is an early formed mineral followed by epidote and then calcite+quartz. Native copper tends to be more associated with epidote, calcite, and quartz. Veins are of two types. Some veins, within the lode and extending into underlying massive basalt, contain the same basic minerals of the lode including native copper and are considered synchronous with lode formation. Once such vein has recently yielded outstanding specimens of native silver and specimens of copper in calcite crystals. Other veins are late-stage and crosscut the native copper mineralized lode. Late-stage veins containing calcite and laumontite are

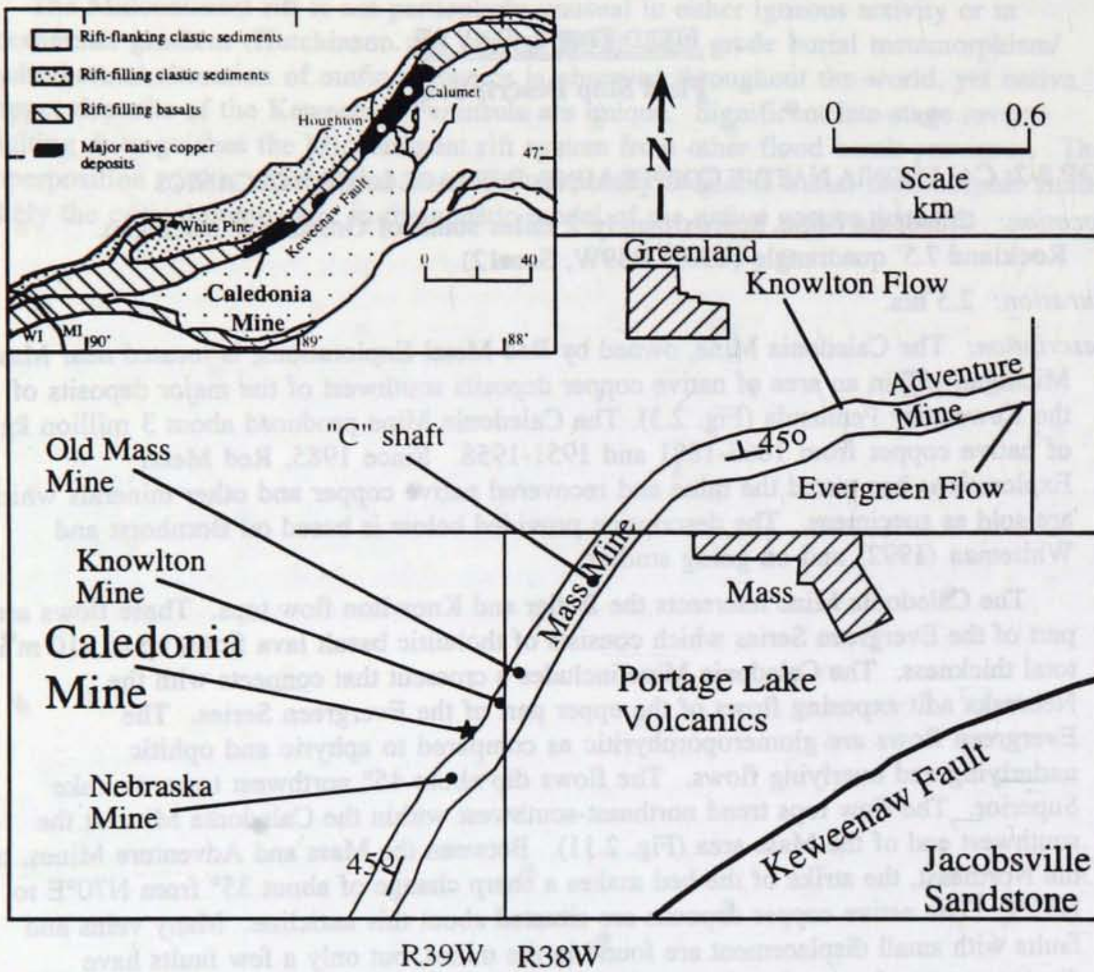

Figure 2.11: Geologic sketch map showing the location of the Caledonia Mine in reference to Michigan township and range (modified from Whiteman and Bornhorst, 1992).

readily visible in the adit. Datolite and adularia are additional late-stage secondary minerals that occur in veins and in the lode. Alteration due to the movement of meteoric groundwaters is noticeable in some part of the lode. Native copper occurs coated with tenorite in some cavities.

STOP 2-3: SOUTH RANGE QUARRY, PORTAGE LAKE VOLCANICS

Location: **Short distance off M-26 just north of South Range village limit. South Range 7.5' Quadrangle (T54N, R34W, Sec. 17, SE of NE)**

Duration: **30 minutes**

Description: **Volcanic textures and structures typical of moderate-to-thick subaerial lava flows within the PLV are well exposed in this old quarry. As one traverses up-section into and through the quarry (Fig. 2.12), one crosses over a 4 m thick interflow conglomerate bed exposed in the path. This is overlain by an 18 m thick lava flow (the amygdaloidal flow top is exposed just as you enter the quarry itself), followed by a complete section through a 42 m thick ophitic basalt flow (the bulk of the quarry** walls). Finally one crosses the lowermost 17 m of the overlying ophitic lava flow (at the northwest end of the quarry). The quarry is positioned approximately in the middle of the PLVs' stratigraphic section. Locally, lava flows strike approximately N45°E (subparallel to the northwestern shoreline of the Keweenaw Peninsula) and dip toward the center of the rift (Lake Superior) at about 60°.

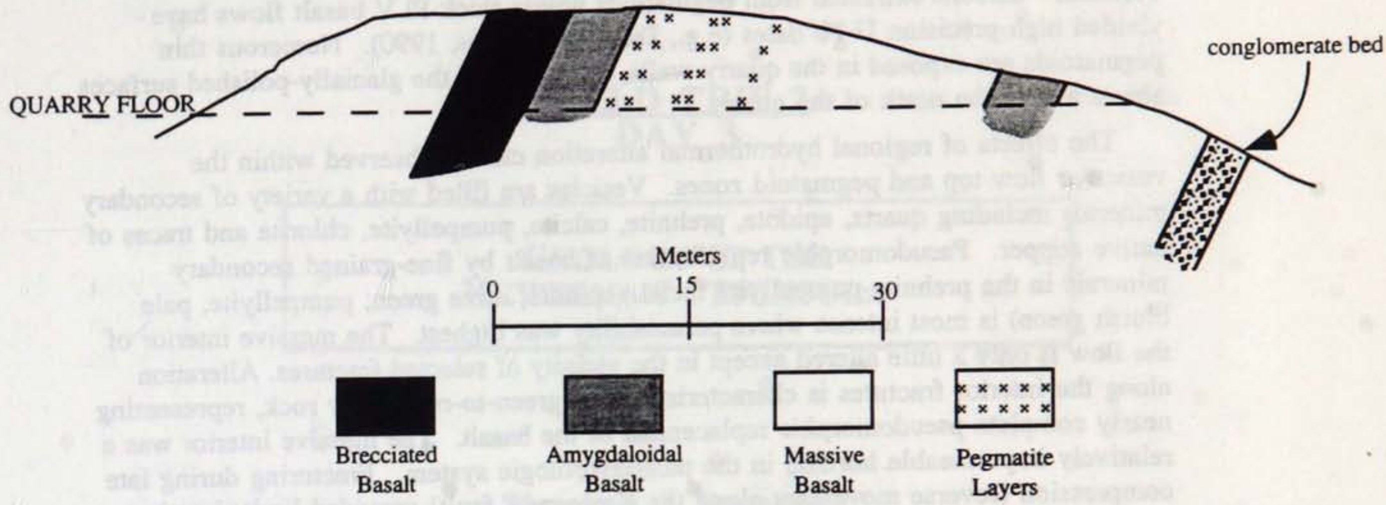

Figure 2.12: Geologic section of the northeast wall of South Range quarry (modified from White, 1971).

Laterally continuous interflow sedimentary beds provide critical stratigraphic markers within an otherwise uniform volcanic pile (PLV). The unit exposed below the quarry has been correlated with the National Sandstone, a marker bed in the Mass-Rockland area. This marker is a massively bedded, pebble-cobble framework conglomerate, composed of silicic-with-subordinate mafic, volcanic, subangular-to-subrounded clasts, within a matrix of poorly-sorted medium-to-coarse sand of similar composition.

The basalts in this portion of the PLV are mainly olivine tholeiites erupted as thick, ponded subaerial lava sheets. The top and bottom of the principle lava flow are exposed at the two ends of the quarry and consist of aphanitic chilled basalt. It was deposited directly on top of the underlying lava flow, so its base occurs where amygdules in the top of the underlying flow disappear abruptly. The upper surface of the main flow was brecciated slightly by the movement of lava after the formation of an upper crust, but it rapidly grades downward to an unbrecciated, highly vesicular flow top. Note the variation in vesicle size and downward distribution of vesicles in the flow. The flow top breccia (locally called fragmental amygdaloid) is laterally discontinuous for this flow. Slow cooling of the lava flow caused solidification toward the flow interior at a rate which allowed development of subophitic to ophitic textures (large oikocrysts of clinopyroxene enclosing a felted framework of An-rich plagioclase and intergranular olivine). The resulting massive, non-vesicular flow interior constitutes about two-thirds of the flow.

Before final solidification, small amounts of volatile-rich, differentiated residual liquid were concentrated in thin discontinuous zones and lenses. Many of these are subparallel to the bottom and top surfaces of the flow. A typical pegmatoid zone consists of a 4 cm to 1.3 m core of vesicular basalt surrounded by a 4 to 9 cm border zone at the top and bottom. The vesicular core of the pegmatoid zones contains coarse laths of Ab-rich plagioclase, prisms of Fe-rich clinopyroxene and abundant Fe-Ti oxides, as well as accessory minerals such as apatite and zircon. The border zone is composed of a medium-to-coarse grained aggregate of albite/oligoclase, augite, ilmenite, and magnetite. Pegmatoid zones toward the top of the flow are more vesicular. Zircons extracted from pegmatoids within thick PLV basalt flows have yielded high-precision U-Pb dates (e.g., Davis and Paces, 1990). Numerous thin pegmatoids are exposed in the quarry walls, as well as in the glacially-polished surfaces above and to the north of the quarry.

The effects of regional hydrothermal alteration can be observed within the vesicular flow top and pegmatoid zones. Vesicles are filled with a variety of secondary minerals including quartz, epidote, prehnite, calcite, pumpellyite, chlorite and traces of native copper. Pseudomorphic replacement of basalt by fine-grained secondary minerals in the prehnite-pumpellyite facies (epidote, olive green; pumpellyite, pale bluish green) is most intense where permeability was highest. The massive interior of the flow is only a little altered except in the vicinity of selected fractures. Alteration along the interior fractures is characterized by a green-to-red cherty rock, representing nearly complete pseudomorphic replacement of the basalt. The massive interior was a relatively impermeable horizon in the paleohydrologic system. Fracturing during late compression (reverse movement along the Keweenaw fault) provided limited pathways for upward movement of ore fluids.

FIELD TRIP 2
DAY 3

GEOLOGY OF THE KEWEENAW PENINSULA

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The geology of the Keweenaw Peninsula is described in the Day 3 overview. Given world's largest native copper mining district. About 96% of the 5 billion kg of refined copper production came from a 45 km long belt in vicinity of Houghton and Calumet. Lenticular, blanket-like orebodies are found along brecciated and amygdaloidal tops of lava flows and interflow conglomerate units within the Portage Lake Volcanics of the Keweenaw Peninsula (see Day 2B and Stop 3-2). Minor copper production was from vein orebodies of the Keweenaw Peninsula (see Stop 3-4). Chalcocite occurs in basalt within the Portage Lake Volcanics near the tip of the Keweenaw Peninsula (Fig. 2.3). Mineralization and alteration of rocks of the Keweenaw Peninsula are discussed in the overview of Day 2 afternoon.

On the northwest side of the Keweenaw Peninsula, rift-filling Middle Proterozoic volcanic and sedimentary rocks dip moderately toward Lake Superior (Figs. 2.2 and 2.13). In stratigraphic succession, these rocks are the Portage Lake Volcanics (Stop 2-2, 2-3, 3-1, 3-2, 3-3, and 3-4), Copper Harbor Conglomerate (Stops 3-5 and 3-6), Nonesuch Shale (Stops 1-4 and 2-1), and Freda Sandstone (Stop 1-4). The Jacobsville Sandstone (Stop 3-1) is slightly younger than the Freda Sandstone and fills a rift-flanking basin on the southeast side of most of the Keweenaw Peninsula. It is in fault contact with the Portage Lake Volcanics along the reverse Keweenaw fault. The Precambrian bedrock of the Keweenaw Peninsula is unconformably capped by a variety of unconsolidated Pleistocene glacial deposits. General descriptions of these units are given below.

Portage Lake Volcanics

The Portage Lake Volcanics (PLV) are composed of a succession of subaerial tholeiitic flood basalt lava flows erupted during a 2.2 +/-1.2 Ma span of time at about 1095 Ma (Davis and Paces, 1990). A total thickness of 5 km, over 200 individual lava flows, is exposed in the Keweenaw Peninsula with the base truncated by the Keweenaw Fault (Fig. 2.13). The basalt section, including the Portage Lake Volcanics, is more than 18 km thick near the center of the rift beneath Lake Superior (Cannon and others, 1989). Rhyolitic volcanic and subvolcanic rocks comprise less than 1 volume percent of the PLV (Nicholson, 1992). Dikes of mafic and intermediate composition cut the exposed Portage Lake Volcanics but are as a whole uncommon. A diorite stock intrudes the base of the Portage Lake Volcanis at Mt. Bohemia. Interflow reddish colored conglomerate and sandstone units total less than 5 volume percent of the Portage Lake Volcanics (Merk and Jirsa, 1982), but increase in abundance toward the top of the formation. Lavas flowed away from feeders along the axis of the rift zone; during intervals of volcanic quiescence sediments were transported from the edges toward the center of the rift. Interflow sedimentary beds were laid down in alluvial fan complexes on essentially flat-lying lava flows by streams flowing towards the center of the rift basin now under Lake Superior. The climate at that time was temperate or tropical (Kalliokoski and Welch, 1985).

Basalts of the Portage Lake Volcanics are relatively primitive, magnesia-rich, high-alumina olivine tholeiites and are typically aphyric (Paces, 1988). Olivine tholeiites are the most abundant, followed by primitive olivine tholeiites with lesser amounts of quartz tholeiites and iron-rich olivine tholeiites (Table 2.2). Minor amounts of basaltic andesite, andesite, dacite, and rhyolite interfinger with the basalts (Table 2.2). Geochemical stratigraphy within the basalts is cyclical with minor and major cycles superimposed on an overall trend toward more primitive compositions in younger basalts within the Portage Lake Volcanics or stratigraphically upward. Basaltic magmas were apparently derived by partial melting of relatively shallow, sub-continental upper mantle with younger basalts being more primitive with little or no contamination by crustal material (Paces and Bell, 1989). The overall compositional trend toward younger, less contaminated primitive magmas can be explained by repeated dike injection and magma eruption at the rift axis that gradually modifies the continental crust through which magmas must pass, and by progressive crustal thinning which provides more efficient transport of magmas to the surface without residence in intracrustal chambers. Major geochemical cycles superimposed on the overall trend are due to fractional crystallization and replenishment in large magma chambers near the crust/mantle interface (Paces, 1988). Minor cycles and silicic rocks result from closed system fractional crystallization in small magma chambers within the crust. Isotopic evidence suggests that the large volumes tholeiitic basaltic magmas which erupted in less than 5 million years as the Portage Lake Volcanics were generated by partial melting of enriched mantle upwelling in a plume (Nicholson and Shirey, 1990; Hutchinson and others, 1990).

Flow thickness ranges from 1 to 450 meters with typical thickness of about 10 to 20 m (White, 1960a; Paces, 1988). A few of the thicker flows can be traced laterally along strike for up to about 90 km but most flows have much less continuity in strike direction. A typical lava flow consists of a thin vesicular base overlain by a massive (vesicle-free) interior capped by a vesicular and sometimes brecciated flow top. The uppermost 5 to 20 % of most individual lava flows is vesicular with between 5 and 50% vesicles. Because vesicles are commonly filled with secondary minerals, flow tops in local terminology are amygdaloids and brecciated flow tops are fragmental amygdaloids. White (1968) estimated that 21% of the lava flows in the Portage Lake Volcanics are fragmental amygdaloids (brecciated).

Table 2.2: Composition of least altered lavas of the Portage Lake Volcanics (table from Bornhorst, 1992; data from Paces, 1988).

	POT	QT1	QT2	IOT	FOT	AND	DAC	RHY
Ni(ppm)	400-300	300-250	250-200	200-100	100-15			
	n=5	n=9	n=14	n=8	n=6	n=1	n=1	n=1
Wt.%								
SiO_2	47.82	47.34	48.03	48.55	49.94	56.39	68.44	77.89
Al_2O_3	15.89	15.27	15.32	15.12	13.28	13.78	15.17	12.77
FeO_t	9.77	11.82	12.32	12.86	14.91	9.87	4.46	1.11
MgO	12.44	11.69	9.85	9.06	7.78	5.52	1.14	0.17
CaO	10.58	10.24	10.16	9.65	6.64	5.10	1.40	0.01
Na_2O	2.04	2.10	2.25	2.31	2.91	3.94	4.74	3.67
K_2O	0.19	0.22	0.33	0.42	1.43	2.27	3.86	4.28
TiO_2	0.98	1.13	1.35	1.60	2.34	1.83	0.51	0.08
P_2O_5	0.16	0.19	0.22	0.25	0.36	1.00	0.19	0.01
MnO	0.14	0.16	0.16	0.18	0.24	0.30	0.08	0.01
PPM								
Ni	326	279	231	172	54	10	7	5
Cu	37	51	73	86	126	5	13	61
Zr	78	85	101	126	212	430	573	145

FeO_t total Fe as FeO

POT Primitive olivine tholeiite
OT1 Olivine tholeiite
OT2 Olivine tholeiite
IOT Intermediate olivine tholeiite
FOT Iron-rich olivine and quartz tholeiites
AND Andesite
DAC Dacite
RHY Rhyolite

Interflow sedimentary rocks, with thickness from a few cm up to about 40 m, are important stratigraphic markers in an otherwise monotonous succession of basalt lava flows. In drill core, sedimentary material in an underlying basalt flow top allows a sedimentary marker horizon to be recognized even where the bed itself is missing (White, 1968). Interflow sedimentary rocks are dominated by well-lithified conglomerate with lesser amounts of interbedded sandstone and occasional significant thicknesses of siltstone and shale. Conglomerates are characterized by sub-rounded to angular, pebbles to boulders (pebbles typical) in a sandy matrix. Clast lithologies are predominantly felsic. In detail, considerable variation exists within and between specific beds, reflecting diversity in the source terrane.

Sedimentary-dominated Rock Units

A thick succession of mostly red-bed sedimentary rocks overlie the Portage Lake Volcanics (Fig. 2.1 and 2.2) and represent a change from a period dominated by volcanism to a period of sedimentation. While the basalts of the Portage Lake Volcanics are interpreted as due to passive rifting, the sedimentary rocks were deposited during a period of post-rift thermal subsidence (Hutchinson and others, 1990). The sedimentary-dominated section filling the rift beneath Lake Superior is as much as 8 km thick (Cannon and others, 1989).

The Copper Harbor Conglomerate, a red-brown basinward-thickening wedge of conglomerate-dominated sediments ranging from 100 to 1800 m thick, conformably overlies and locally interfingers with the Portage Lake Volcanics. Conglomerates are composed predominantly of felsic volcanic clasts interbedded with lithic sandstones deposited as coalesced alluvial fans and sand flats. The middle portion of the Copper Harbor Conglomerate (northeast of Calumet) includes a succession of 31 lava flows with a maximum thickness of 600 m termed the Lake Shore Traps (Lane, 1911). The composition of the Lake Shore Traps is more variable than the Portage Lake Volcanics ranging from Fe-rich olivine tholeiites at the base to Fe-rich, olivine-bearing tholeiitic basaltic andesites to tholeiitic andesites at the top (Paces and Bornhorst, 1985). Geochemical stratigraphic relationships can be explained by a combination of fractional crystallization, parental magma replenishment, and wall rock assimilation. Davis and Paces (1990) report a U-Pb age on zircon of 1094 +/- 3.6 Ma for the Lake Shore Traps (Figs. I.3 and 2.2).

The Nonesuch Shale, maximum thickness 215 m, is a succession of siltstones, shales, carbonate laminates and minor sandstones with low to moderate amounts of total C that overlies and interfingers with the Copper Harbor Conglomerate. The Nonesuch Shale was deposited in a lacustrine environment.

The Freda Sandstone is a cyclic succession of red-brown ferruginous sandstone, siltstone, and mudstone overlying and in gradational contact with the Nonesuch Shale. Its thickness is greater than 3,700 m as the top is not exposed. Sediments of the Freda Sandstone were deposited in a fluvial braided stream environment.

The Jacobsville Sandstone is a red to bleached-white succession of coarse- to fine-grained feldspathic and quartzose sandstones with varying amounts of siltstone, shale, and conglomerate over 3,000 m thick. The Jacobsville Sandstone is in fault contact with the Portage Lake Volcanics along the Keweenaw Fault on the southeast side of the Keweenaw Peninsula. The Jacobsville strata are completely devoid of igneous rocks and were fluvially deposited in a rift flanking basin, at least in part during active reverse movement along the Keweenaw Fault (Kalliokoski, 1988; Hedgman, 1992).

Structure

The Keweenawan strata dip moderately north to northwest toward the center of the rift (Lake Superior) with dip angles steepening toward the Keweenaw Fault at the base of the section. About half of the dip of the Portage Lake Volcanics is due to syn-depositional downwarpage (White, 1968) while the remaining tilting of these beds occurred during reverse movement along the Keweenaw fault. Bedding in the Jacobsville Sandstone dips less than 50 in most areas except near the Keweenaw fault where dips steepen to vertical in response to drag along the fault.

The Keweenaw Fault is the major fault in the Keweenaw Peninsula. It is a low- to high-angle reverse fault marking the border of the main rift-filling volcanic and sedimentary rocks with rift-flanking sedimentary rocks. The Keweenaw fault strikes more or less parallel to the bedding of the Portage Lake Volcanics and dips generally sub-parallel to the Portage Lake Volcanics. The Keweenaw Fault was originally a graben-bounding normal fault that was transformed at 1060 +/- 20 Ma (Cannon and others, 1990) into a high-angle reverse fault late in the history of the rift (Cannon and others, 1989). The probable cause of the compression was continental collision along the Grenville front (Cannon, 1994; Cannon and Hinze, 1992; Hoffman, 1989).

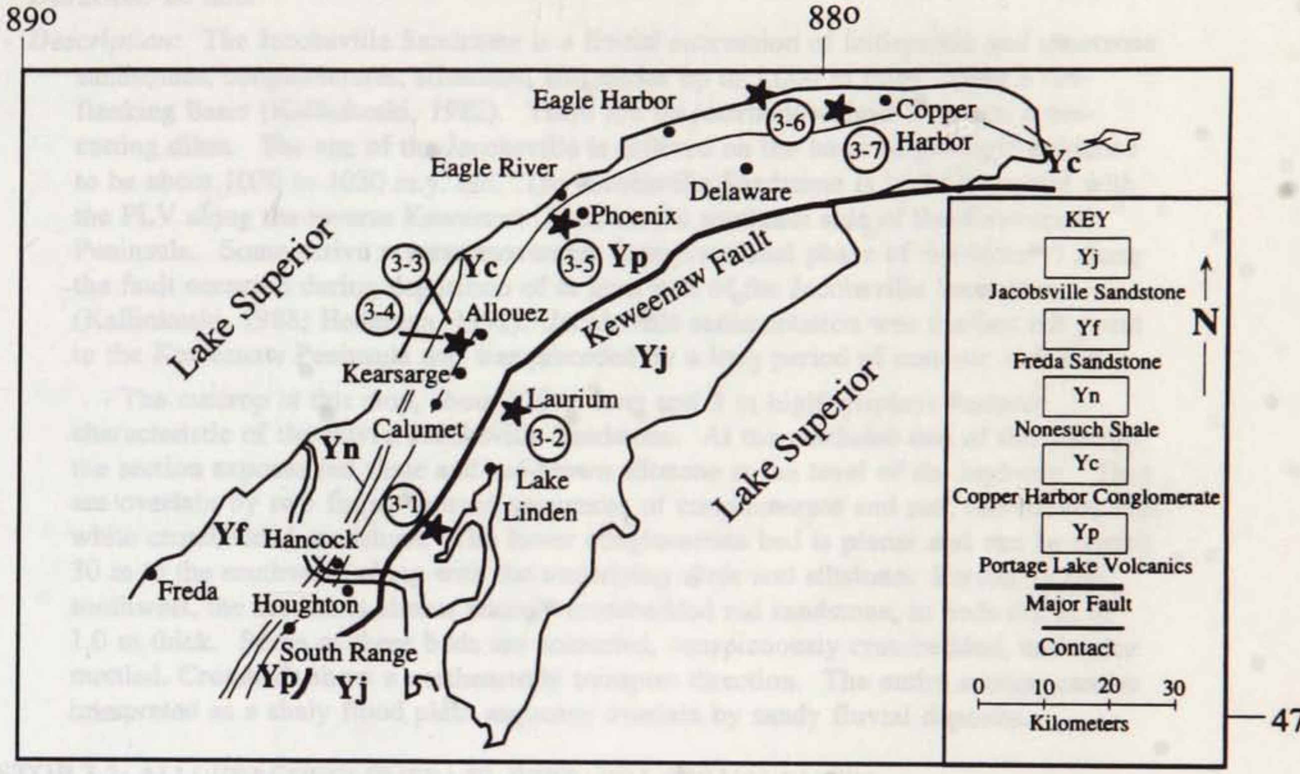

Figure 2.13: Geologic map showing the location of Day 3 field trip stops.

FIELD TRIP 2 DAY 3

Field Stop Descriptions

STOP 3-1: JACOBSVILLE SANDSTONE

Location: From Houghton, proceed across the Portage Lake Lift Bridge towards Hancock. Turn right onto M-26 just off the bridge. Pass through Dollar Bay and Mason. Stop 3-1 is 6.6 miles from the bridge. Laurium Quadrangle (T55N, R33W, Sec. 14, SE of SE).

Duration: 20 min.

Description: The Jacobsville Sandstone is a fluvial succession of feldspathic and quartzose sandstones, conglomerates, siltstones, and shales up to 1,000 m thick filling a rift-flanking basin (Kalliokoski, 1982). There are no interbedded lava flows or cross-cutting dikes. The age of the Jacobsville is inferred on the basis of geologic evidence to be about 1070 to 1030 m.y. old. The Jacobsville Sandstone is in fault contact with the PLV along the reverse Keweenaw Fault on the southeast side of the Keweenaw Peninsula. Some active reverse movement (compressional phase of rift-history) along the fault occurred during deposition of at least part of the Jacobsville Sandstone (Kalliokoski, 1988; Hedgman, 1992). Jacobsville sedimentation was the last rift event in the Keweenaw Peninsula and was preceded by a long period of cratonic stability.

The outcrop at this stop, about 120 m long and 3 m high, displays features characteristic of the fluvial Jacobsville Sandstone. At the northeast end of the outcrop, the section exposes red shale and red-brown siltstone at the level of the highway. They are overlain by two fining-upward sequences of conglomerate and red, red-brown, and white crossbedded sandstone. The lower conglomerate bed is planar and can be traced 30 m to the southwest, along with the underlying shale and siltstone. Farther to the southwest, the section is almost entirely crossbedded red sandstone, in beds 0.2 m to 1.0 m thick. Some of these beds are contorted, conspicuously crossbedded, and color mottled. Crossbeds show a northeasterly transport direction. The entire section can be interpreted as a shaly flood plain sequence overlain by sandy fluvial deposits.

STOP 3-2: ALLOUEZ CONGLOMERATE, PORTAGE LAKE VOLCANICS

Location: Continue on M-26 through Tamarack City, Hubbell, Lake Linden, and Laurium. Turn right onto US-41 in Calumet, and follow it through Centennial, Kearsarge, and Allouez. 0.2 miles into Allouez, turn left onto Bumbletown Road (located just before a gas station). Bear right on this road and park on the right at the dirt road. This is 18.7 miles from Stop 3-1, 0.55 miles from Allouez. Ahmeek Quadrangle (T57N, R32W, Sec. 31, NE of SW).

Duration: 30 min.

Description: The rock piles here are from the Allouez Conglomerate Mine, which operated from 1869 to 1892 and produced about 1.2 million kg of copper. The Allouez Conglomerate is one of a small number of interflow sedimentary horizons within the PLV These sedimentary horizons are important for stratigraphic correlations within the otherwise monotonous pile of basalt lava flows of the PLV. This bed can be traced along strike from the tip of the Keweenaw Peninsula west and south, to at least the Mass area. Its strike length is more than 120 km, making it one of the most continuous sedimentary horizons in the Keweenaw Peninsula. The Allouez Conglomerate is exposed in underground workings at the Quincy Mine Adit, Hancock, and in underground workings at the Delaware Mine. It is stratigraphically just below the

Greenstone Flow, the thickest flow in the PLV. The Greenstone Flow is prominent in the northern half of the Keweenaw Peninsula. Like other interflow conglomerate beds within the PLV, the Allouez Conglomerate consists of mostly conglomerate with lesser amounts of sandstone and siltstone. These red-colored clastic sedimentary rocks were deposited in a terrestrial alluvial fan environment. The dominant transport of sediment was from the margins of the rift toward the center (current center of Lake Superior) during a hiatus of volcanic activity.

The rock piles from the Allouez Conglomerate illustrate features of interflow sedimentary beds of the PLV. The largest boulders in this conglomerate are about 65 cm in diameter, and the median size is about 8 cm. A pebble count of boulders more than 20 cm across gave the following results: mafic rock, mostly amygdaloidal, 16%; quartz porphyry, 36%; feldspar porphyry, 11%; and granophyre, 37% (White, 1971b).

Little evidence of native copper mineralization is present in this rock pile. Occasionally, one can find a specimen with native copper filling the void space between clasts and grains. Calcite and chlorite are the dominant pore-filling secondary minerals. Calcite-rich cemented zones may represent vadose carbonate or paleocliche (Kallioski, 1986), indicating that not all calcite is secondary hydrothermal in origin. Thin black veinlets cutting the Allouez Conglomerate are calcite with inclusions of chalcocite. Supergene alteration resulting from the downward percolation of groundwater is rare in Keweenawan native copper deposits. The effects of supergene alteration are quite visible here as chrysocolla, malachite, and cuprite are present in numerous samples.

STOP 3-3: ALLOUEZ FAULT GAP AND PHYSIOGRAPHY OF THE KEWEENAW PENINSULA

Location: From Stop 3-2, continue on the main paved road to the top of Bumbletown Hill. Turn right on Cedar Street. The top of the hill is near the communication towers. Ahmeek Quadrangle (T57N, R32W, Sec. 31, NW of SE).

Duration: 15 minutes

Description: From this location on a clear day, Isle Royale may be seen 80 km to the northwest. The Huron Mountains may be seen beyond Keweenaw Bay, 60 km to the southeast. The southeast flank of the Keweenaw Peninsula has a steeper slope at the skyline which approximately follows the line of the Keweenaw Fault. The low-lying plain between the fault and Keweenaw Bay is underlain by flat-lying Jacobsville Sandstone. The Huron Mountains in the distance consist of Archean and early Proterozoic basement unconformably overlain by Jacobsville Sandstone. Bumbletown Hill is located on the southwest side of the Allouez Gap which follows the SE trending valley near the relatively new headframe. The Allouez Gap Fault is a zone of faults and fractures along which the PLV and Keweenaw Fault are offset. At this gap, the strike of the PLV swings from about N35°E to N50°E. Almost every permeable horizon near the Allouez Gap Fault contains above average amounts of native copper. In fact, nowhere else in the district has so many mineralized beds. About 60% of the district production can be linked to the fault as a primary pathway for ore fluids. The fault bisects the Kearsarge deposit, which is the second largest producer in the district.

Looking northeast along the strike of the PLV, one can see the cuesta form of the ridge upheld by the Greenstone Flow. At Bumbletown Hill, this flow is only 85 m thick, but it thickens abruptly to more than 400 m at the near end of the cuesta ridge. To the right of the Greenstone ridge, the more distant hills are formed by lava flows lower in the section.

STOP 3-4: PHOENIX MINE AND GREENSTONE FLOW, PORTAGE LAKE VOLCANICS

Location: From the intersection of US-41 and Bumbletown Road in Allouez, continue through Ahmeek and Phoenix. 1.1 miles outside of Phoenix, turn left on a dirt road just 0.1 miles before the US-41 and M-26 junction. The Phoenix Mine rock pile is about 100 m from the paved road. Phoenix Quadrangle (T58N, R31W, Sec. 30, NE of SE).

Duration: 45 min.

Description: The Phoenix Mine worked numerous veins below the Greenstone Flow. One of the earlier mines in the district, the mine operated off and on from 1849 to 1917. The vein was worked to a vertical depth of about 300 m, with varied grade (average grade around 1.5%). Since the vein cuts thin lava flow tops, no mining was done on adjacent mineralized flow tops. The Phoenix Mine produced a total of about 8 million kg of refined copper (Butler and Burbank, 1929). It also worked the Ashbed Amygdaloid where it is mineralized, in the vicinity of vein copper occurrences. The Phoenix Mine rock pile is notable for halfbreeds (native copper plus native silver). A spectacular example of this is crystallized analcite and chlorastrolitic pumpellyite (Michigan "greenstone"). Other minerals that have been reported in the Phoenix Mine area (Clarke, 1974) include pumpellyite, chlorite, natrolite, and apophyllite.

To look at the Greenstone Flow you must climb over the rock pile and then pass one of the fissure zones just above the shaft. Proceed ahead and climb to the base of the steep cliffs, which are composed of a portion of the massive flow interior of the Greenstone Flow. This flow is an enormous lava flow over 400 m thick, perhaps representing the greatest single outpouring of lava on Earth. A very coarse ophitic zone occurs near the base of the Greenstone Flow. By following the exposures along the cliff, the pegmatoid, subophitic, and ophitic zones can all be observed.

The Greenstone Flow has been identified for a distance of 90 km along the length of the Keweenaw Peninsula, and throughout the length of Isle Royale, 90 km northwest on the opposite limb of the Lake Superior Syncline. The extent is 5000 km2 with volume of 800 to 1500 km3 (Longo, 1983; White, 1960a). Very slow solidification of this great mass of magma allowed extensive in-situ magmatic differentiation. This resulted in a massive, ophitic zone at the base of the flow, an overlying zone of intercalated subophitic and pegmatoidal layers, an upper ophitic zone, and a fine-grained, vesicular flow top (Cornwall, 1951a and b). The lower ophitic zone experienced rates of undercooling low enough to allow growth of clinopyroxene oikocrysts up to 5 cm in diameter.

The composition of the Greenstone Flow magma was more evolved than typical olivine tholeiites, which constitute the greatest volume of the PLV. Primitive olivine tholeiite and quartz tholeiite occur between the Greenstone Flow and the top of the PLV. Generally, magmas become more primitive and less crustally contaminated with time during the Midcontinent rift development, reflecting changes in magmatic and tectonic processes. A model of magmatic and rift evolution based on PLV data involves primitive parental magma modification by complex, open-system fractional crystallization in large reservoirs at the base of the thinned crust (Paces, 1988). The resulting olivine tholeiite magma is either erupted at the surface or supplied to smaller chambers at higher levels, where further crystallization produces evolved tholeiites and silicic rocks.

STOP 3-5: LAKE SHORE TRAPS, COPPER HARBOR CONGLOMERATE

Location: Turn left onto M-26 towards Eagle River. Stay on M-26 for 14.25 miles (from Stop 3-4). Delaware Quadrangle (T59N, R30W, Sec. 36, SE of NW).

Duration: 30 min.

Description: The basalts cropping out at Esrey Park within the Copper Harbor Conglomerate are an informal member called the Lake Shore Traps, some 800-1000 m above the PLV. They are a part of a succession of Fe-rich olivine tholeiite, basaltic andesite, and andesite lava flows known collectively as the Lake Shore Traps (an informal member within the Copper Harbor Conglomerate). This succession is thickest at the tip of the Keweenaw Peninsula (approximately 600 m). It thins toward the east (on Manitou Island) and the west-southwest, where the lava flows pinch out near Calumet (a total strike length of about 90 km). Individual lava flows vary in thickness from about 4 to 40 m and exhibit volcanological features similar to flood basalts of the PLV. The lowermost mafic flows were deposited as ponded sheets while upper andesite flows may have formed a low, positive topographic feature such as a shield volcano. Magmatic variation within lava flows at the tip of the Keweenaw Peninsula imply that fractional crystallization of plagioclase, clinopyroxene, olivine, Fe-Ti oxide, apatite, and zircon played an important role in the petrogenesis of the Lake Shore Trap magmas (Paces and Bornhorst, 1985). Additional processes of parental magma replenishment and possible wall rock assimilation, however, are required to explain geochemical-stratigraphic relationships.

The large outcrop between the parking lot and the shore is a massive flow interior of fine-grained, Fe-rich olivine tholeiitic basalt. The flows strike parallel to the shoreline and dip 20-30° toward the lake. The upper portion of this flow is not exposed, but the top of the underlying basalt flow can be seen at the shoreline on either side of the large outcrop. Because of its higher stratigraphic level within the rift-fill sequence, the degree of metamorphism/alteration in the Lake Shore Traps is much lower than in the PLV. Zeolite facies metamorphism, as opposed to prehnite-pumpellyite, affected the flow top and deposited chalcedony, laumontite, analcite, calcite, and smectite in amygdules. From the east end of the large basalt ridge next to the parking lot, walk about 100 m east along the shore to see the flow top of the underlying flow and associated amygdule fillings. Massive flow interiors of the Lake Shore Traps often retain relict olivine and glassy, intersertal mesostatis in contrast to the PLV, where both olivine and intersertal glass are invariably replaced by Mg-Fe phyllosilicates.

STOP 3-6: COPPER HARBOR CONGLOMERATE AND BROCKWAY MOUNTAIN VIEWPOINT

Location: Turn around and head back toward the junction of Brockway Mountain Drive, 0.9 miles from Stop 3-5. Turn left onto this road and follow it to the top. Lake Medora Quadrangle (T59N, Sec. 34, SE of NW).

Duration: 15 min.

Descripton: This high conglomerate ridge reaches an elevation of over 400 m, with excellent views of the ridge and valley topography of the northern shore of the Keweenaw Peninsula. Underfoot, the Copper Harbor Conglomerate dips about 20o to the north. To the west, the Lake Shore Traps form island chains on a prominent ridge in the vicinity of Agate Harbor and Esrey Park. Copper Harbor Conglomerate is found in the drowned valleys and along the outer ridge jutting into Agate Harbor and projecting into a smaller island chain. The reefs of the Lake Shore Traps and Copper Harbor Conglomerate along the Keweenaw's north shore are the site of numerous shipwrecks. Lake Bailey (with the small island) and Lake Upsom occupy a topographically low valley on a finer-grained clastic horizon within the Copper Harbor Conglomerate. Just to the south of Lake Bailey is the conglomerate ridge of Mt. Lookout, marking the contact between the Copper Harbor Conglomerate and the PLV. The inland lake almost directly south is Lake Medora, and just before the lake is a prominent ridge which marks the stratigraphic position of the Greenstone Flow. Farther to the south across Lake Medora is Mount Bohemia, a dioritic stock-sized intrusion. To the east on the skyline, beyond Copper Harbor,lies East Ridge, a prominent conglomerate hill. To the north on the skyline 65 km away is Isle Royale, which is visible on a clear day. The skyline of Isle Royale is formed by the Greenstone Flow, as it is on the Keweenaw Peninsula.

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FIELD TRIP 3

GEOLOGY, PETROLOGY, AND METALLOGENY OF INTRUSIVE IGNEOUS ROCKS OF THE MIDCONTINENT RIFT SYSTEM

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INTRODUCTION

The Middle Proterozoic geology of northeastern Minnesota and northwestern Ontario is unique relative to other parts of the Midcontinent rift (MCR) because of the predominance of intrusive rocks over volcanic rocks. The Duluth Complex alone covers an area of over 5000 km^2 and locally extends to depths of 12 km (Allen, 1994). Such a concentrated volume of upper crustal intrusions also appears to be exceptional when compared to other subvolcanic intrusive complexes associated with continental flood basalt (CFB) and other large igneous provinces. Although abundant geochemical and geophysical evidence exists for lower crustal magma chambers providing differentiated magmas to flood basalts (Cox, 1980; White, 1992), upper crustal intrusions tend to occur as feeder dike swarms and isolated sills of limited (< 1 km) thickness that are volumetrically minor in comparison to their comagmatic volcanics. The abundance and variably differentiated character of the intrusions associated with the MCR provide an rare opportunity to study the interrelationship of the plutonic and volcanic components of intracontinental rift environments.

With an expand classification of Weiblen (1982), intrusive rocks of the MCR can be generally divided into five groups based on their compositional range, extent of internal differentiation, intrusive form, and nature of the country rock. They are:

1) Layered mafic intrusive complexes: As exemplified by the Duluth and Mellen complexes (Figs. I.2 and 3.1), these immense, multiple-intrusive complexes are composed of numerous, large (many >4 km thick), mafic (tholeiitic) layered intrusions, extensive accumulations of structurally complex anorthositic rocks, and large massive bodies of felsic rocks. The complexes were emplaced into the base of the volcanic edifice.

2) Subvolcanic intrusions: Other mafic to intermediate intrusions were emplaced into higher levels of the volcanic pile. With a few notable exceptions, these hypabysssal intrusions have noncumulate textures and typically occur as isolated dikes, sills, or stocks. The Beaver Bay Complex is an exceptionally high concentration of intrusions, including some mafic layered intrusions and felsic bodies, that were emplaced midway along Minnesota's North Shore (Figs. 3.1 and 3. 12).

3) Mafic sills hosted by sedimentary rocks: Thick (50-200 m), reversely polarized, gabbroic sills in subhorizontal pre-Keweenawan sedimentary rocks occur in two general settings. In a large area straddling the international boundary, sills intrude Early Proterozoic iron-formation, graywacke, and shale (Fig. 3.1). In the Lake Nipigon area, sills intrude the Middle Proterozoic epicontinental sedimentary rocks of the Sibley Group (Fig. I.2). In both areas, the intrusions are termed the Logan Sills.

4) Alkaline and carbonatite intrusions: Isolated intrusions of alkaline gabbroic to syenitic and carbonititic rocks and related breccia diatremes were emplaced into Archean granite-greenstone terrane in a vast area northeast of Lake Superior (Figs. I.2 and 3. 24). The largest and best known of these is the alkaline Coldwell Complex exposed along the North Shore at Marathon, which formed by multiple intrusion and some differentiation in place.

5) Dike Swarms: Mafic dike swarms of varied composition, age, orientation, and host rock occur throughout the Lake Superior region (Fig. I.2 ; Green and others, 1987).

This field trip introduces the geology, structure, and mineralization of the intrusive rock groups that are exposed along the northwestern and northern flank of the MCR in Minnesota and Ontario. Each of the five field days will focus on one of the major groups of Keweenawan intrusive rocks listed above that are exposed along the North Shore of Lake Superior from Duluth to Marathon. Each suite will be described in more detail at the beginning of each day's field log. The final day will emphasize the nature of the volcanic rocks which host these intrusions.

<u>FIELD TRIP 3</u>

DAY 1

THE DULUTH COMPLEX AT DULUTH

James D. Miller, Jr.
Minnesota Geological Survey

<u>Overview</u>

The Duluth Complex is a large composite mafic intrusive supersuite covering about 5,000 km^2 on the northwest flank of the main rift axis (Fig. I.2). Intrusions of the complex were emplaced generally in the vicinity of an uncomformity between a peneplained surface of Early Proterozoic (Animikian) graywacke, slate, and iron-formation and Archean granite-greenstone terrane and the overlying edifice of plateau lavas and minor clastic sedimentary rocks of the North Shore Volcanic Group (NSVG; Fig. 3.1). Strongly hornfelsed volcanic bodies occur locally within the complex and suggest that magma chambers were formed by effectively delaminating lava flows at the base of the volcanic pile. Internal structures and geophysical modeling (Chandler and Ferderer, 1989; Chandler, 1990) indicate that some intrusions were emplaced roughly concordant with overlying flows as south- to east-dipping sheets, whereas other intrusions have steeply dipping contacts that are more indicative of funnel-shaped or half-graben chambers. Overall, two strong (+50 and +70 mgal) Bouguer gravity anomalies indicate that the deepest parts of the complex are rooted to depths of more than 12 km (Allen, 1994). In addition to aiding in understanding the deeper structures of the complex, geophysical studies, especially aeromagnetic data, have been integral to interpreting the surface geology over poorly exposed areas of the southern and central complex (Fig. 3.1).

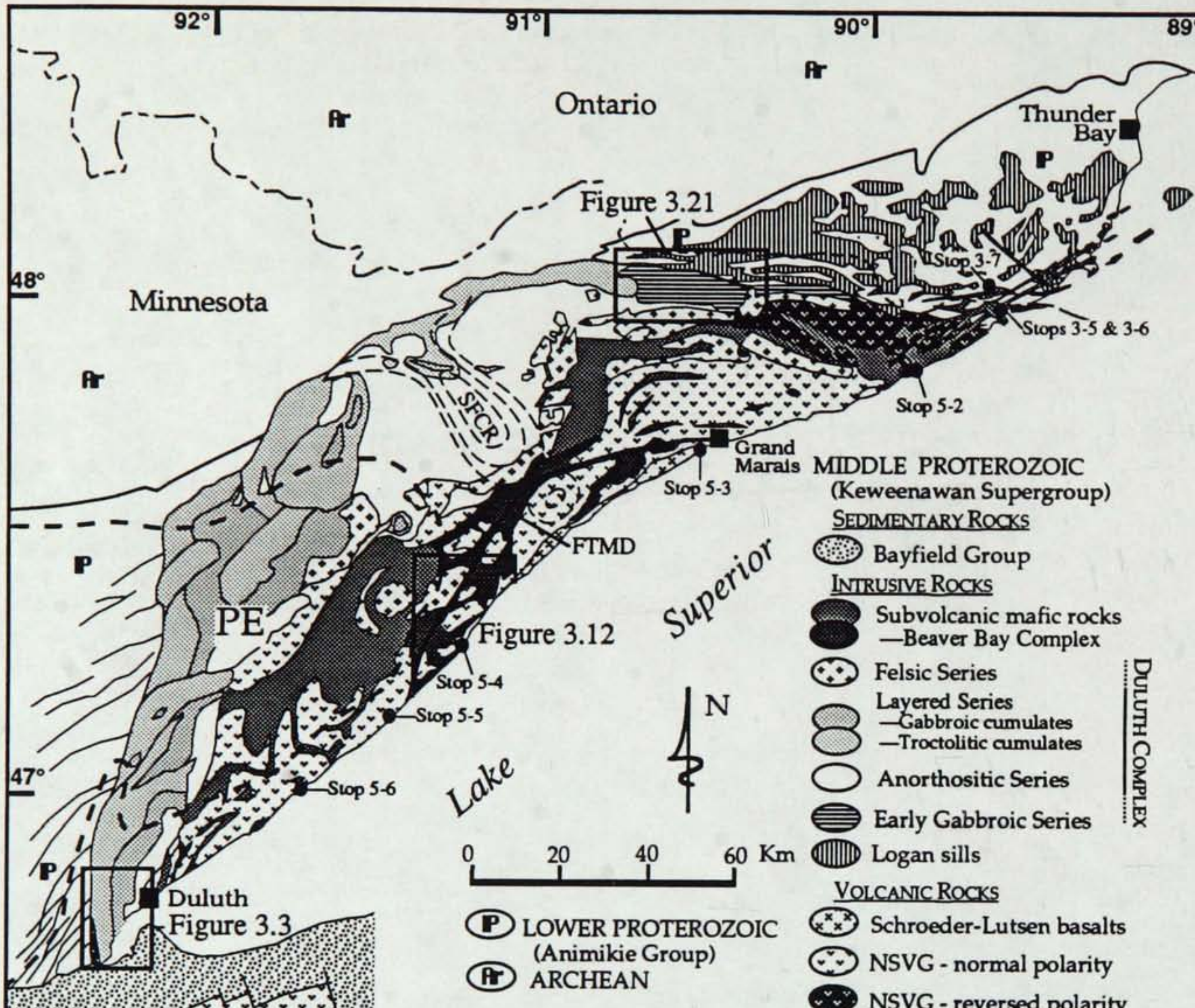


Figure 3.1: Middle Proterozoic geology of northeastern Minnesota and part of Ontario. Locations of Figures 3.3, 3.12 and 3.21, which show more detailed geology, and of Stops 3-5 to 3-7 and 5-2 to 5-6 are indicated. Abbreviated structural features are: FTMD - Finland tectono-magmatic discontinuity and SFCR - Schroeder-Forest Center crustal ridge (contours schematically portray position and depth of buried ridge). PE - area of poor exposure (heavy dashed line delimits area).

The Duluth Complex is typically subdivided into four major series based on lithology, internal structure, and intrusive relationships (Weiblen and Morey, 1980):

1) Early gabbro series (commonly referred to as Nathan's layered series)—occurs along the northern tier of the Duluth Complex as a layered sequence of oxide-rich gabbroic cumulates (Stops 3-1 to 3-4) that a recent U-Pb date shows to be about 8 m.y. older than the main mafic phases of the Duluth Complex (Paces and Miller, 1993).

2) Felsic series—refers to a collection of granitoid (commonly granophyre) intrusions strung along the roof zone of the Duluth Complex. Intrusive relationships with adjacent gabbroic rocks indicate a variety of ages, but radiometric dates have not been obtained.

3) Anorthositic series — typically occupies the structually highest levels of the complex as a multiple-intrusive, structurally complicated suite of laminated, but unlayered, plagioclase cumulates. Their erratic internal structure, felspathic composition, porphyritic character, and lack of signs of internal differentiation have been interpreted as resulting from emplacement and crystallization of plagioclase crystal mushes (Grout, 1918c; Taylor, 1964; Miller and Weiblen, 1990).

4) Layered (or troctolitic) series — composed of many, discrete layered mafic intrusions (some of the larger-scale bodies are delineated in Fig. 3.1). The stratiform internal structure and cryptic compositional layering of these intrusions are consistent with their formation by crystal fractionation accompanied by periodic recharge and eruption.

Layered series intrusions commonly intrude anorthositic series rocks as well as one another. However, nearly identical 1099 Ma U-Pb dates from samples of two layered series and two anorthositic series from different parts of the complex (Paces and Miller, 1993) disprove a long-standing interpretation that the anorthositic series is significantly older than the layered series. A reinterpretation of the contact relationships at Duluth (Miller, 1995b) opens the way for reinterpreting the petrogenetic relationship of the two main series of the Duluth Complex.

Several types of ore deposits occur in the Duluth Complex. The most extensive is low-grade, disseminated to locally massive Cu-Ni sulfide mineralization along the northwestern margin of the Duluth Complex (Bonnichsen, 1972; Weiblen and Morey, 1976; Morton and Hauck, 1987). Current ore reserves are estimated at 4.4 billion tons of 0.66% Cu and 0.2% Ni (Hauck and others, in press). A second ore type is Fe-Ti(-V) oxide mineralization, which occurs as stratiform layers enriched in cumulus ilmenite, as crosscutting, ilmenite-rich ultramafic bodies, and as oxide-rich metasedimentary hornfels inclusions (Hauck and others, in press). Platinum-group-element (PGE) mineralization has been found in association with Cu-Ni sulfide mineralization and with Cr-spinel-bearing troctolite (Morton and Hauck, 1987; Mogessie and Saini-Eidukat, 1992). The types of mineralization recognized in the Duluth area are Fe-Ti oxide mineralization in the form of crosscutting, coarse-grained, ilmenite dunite bodies (Stop 1-1) and PGE occurences associated with microgabbro layers in the medial part of the layered series (Miller, 1995a).

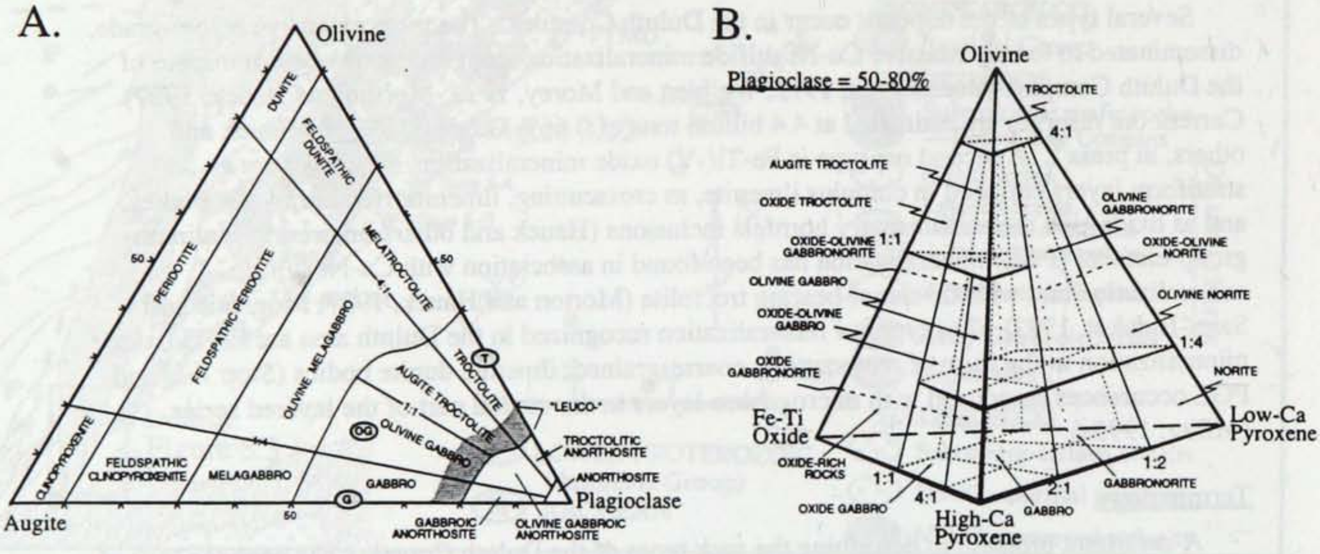
Terminology

A persistent problem in describing the rock types of the Duluth Complex has been the lack of a generally accepted rock classification scheme . Many studies of the Duluth Complex have used variations of the modal classification scheme developed by Phinney (1972a). Miller and Weiblen (1990) argued against such a classification scheme because it employs modal boundaries that ignore natural (experimental and empirical) cotectic proportions of mineral phases expected to crystallize from basaltic magmas at low pressures. The same can be said of the popular classification scheme of Streckeisen (1976). The modal rock terms defined in Figure 3.2A and B will be used to describe gabbroic and anorthositic rocks to be seen on this trip. However, a modal classification scheme alone gives no indication of how various gabbroic rocks may have crystallized and very often leads to much ambiguity. For example, the implied mineral paragenesis of an ophitic olivine gabbro is much different from that of an intergranular olivine gabbro or that of an intergranular poikilitic-olivine gabbro. Therefore in addition to the modal rock name, rock type descriptions in this guide commonly use 1) textural modifiers that note the grain size, lamination development, and the habit of mafic phases (e.g., ophitic, intergranular, poikilitic) and 2) additional mineral modifiers that note important accessory mineral phases (e.g., apatitic, biotitic) or essential minerals whose presence is important to note, though they may not be sufficiently abundant to affect the rock name (e.g., olivine-bearing gabbro, leucogabbro).

In addition to a standard modal mineralogy-based nomenclature, most studies of layered intrusions also employ a more interpretive cumulate nomenclature based on mineral habit and mode (e.g., McCallum and others, 1980; Zientek,and others, 1985). Most of the rocks comprising the intrusions of the Duluth Complex are cumulates as described by Irvine (1982, p. 131) - "an igneous rock characterized by a cumulus framework of touching mineral crystals or grains that were evidently formed and concentrated by processes of fractional crystallization. The fractionated crystals are called cumulus crystals. They are typically subhedral to euhedral, and generally they are cemented together by a texturally later generation of postcumulus material that appear to have crystallized from the intercumulus liquid. . . " Although cumulate classifications are used as an interpretive tool, they are empirically based on the mineral mode and texture of the rock. A cumulate classification scheme is therefore used here which is based on the following empirical criteria:

1) applies to rocks which show some igneous lamination or modal layering thereby indicating segregation of mineral phases from a liquid or from each other;

2) minerals comprising more than 2 vol.% are listed in decreasing order of abundance; and



C.

Cumulus/Intercumulus Mineral Codes			
PP/P/p -	plagioclase°	H/h -	hypersthene
O/o -	olivine	Hb/hb -	hornblende
A/a -	augite	B/b -	biotite
F/f -	Fe-Ti oxide	Ap/ap -	apatite
IP/ip -	inverted pigeonite	/gp -	granophyre

° Use PP for anorthositic compositions

Cumulate*	Conventional Rock Term
PPaof	ophitic olivine gabbroic anorthosite
OP	melatroctolite
PaOf	ophitic olivine gabbro
PAFO	intergranular olivine gabbro
PAFo	poiklitic olivine gabbro
PAIPf	intergranular oxide-poor gabbronorite
PAFbgp	intergranular granophyric biotitic gabbro

**List phases composing >2% mode in order of abundance*

Figure 3.2: Classification scheme used in this guide for mafic plutonic rocks. A) Modal rock names base on plagioclase, olivine, and augite. T, OG, G denote cotectic proportions of P+O, P+O+A, and P+A, respectively, based on low pressure experimental and empirical data (McCallum and other, 1980). B) Modal rock names of mafic rocks containing 50-80% plagioclase based on olivine, low-Ca pyroxene, high-Ca pyroxene, and Fe-Ti oxide. C) Abbreviations for cumulus (upper case) and intercumulus (lower case) mineral phases for a cumulate rock classification code (see text); some examples of the cumulate scheme are given in the box on the right.

3) granular (cumulus) mineral phases are denoted with upper case letter abbreviations, and interstitial (intercumulus) mineral phases with lower case abbreviations (Fig. 3-2C).

In effect, this cumulate classification scheme is a shorthand description of the modal/textural rock classification scheme (e.g., Fig. 3.2C). In many ways, it is more descriptive because it gives the general habit of all mineral phases.

The Duluth Complex at Duluth

The well-exposed gabbroic rocks forming the escarpment above Duluth have long been recognized as the type section of the Duluth Complex. The pioneering studies in the Duluth area by Frank Grout in the mid-1910s stand as the principal contribution to the petrology of the Duluth Complex. In a series of papers published in 1918 from his Ph.D. dissertation (Grout, 1918a-e), he recognized the complex to be a multiply intrusive body predominantly composed of early anorthositic gabbros, younger layered gabbros, and felsic differentiates. Recognizing that Grout's conclusions about the differentiation of the layered gabbros at Duluth resembled those of Wager and Deer in their classic 1939 treatise on the fractionational crystallization of the Skaergaard Intrusion of East Greenland, Richard Taylor set out in the mid-1950s to better characterize the petrology of Duluth Complex at Duluth for his Ph.D. dissertation. While Taylor found many similarities between the two magmatic systems, he observed little cryptic variation in feldspar and mafic mineral compositions in the Duluth layered series compared to the strong alkali and iron enrichment observed in the Skaergaard, and concluded that this probably reflected periodic magmatic replenishment of the Duluth system. Perhaps Taylor's greatest contribution to the Duluth Complex was the first large-scale (1:24,000) geologic map of the complex (Taylor, 1964) wherein he distinguished intrusive units of anorthositic gabbro, layered series gabbro, ferrogranodiorite, granophyre, and diabase. More recent mapping and petrologic studies of the Duluth Complex at Duluth confirm its status as a typical example of Duluth Complex geology (Miller and others, 1993; Miller,1995b).

Recent mapping of the Duluth Complex in the Duluth area (Miller and others, 1993) reconfirms the existence of the two fundamentally different and easily discernable rock series initially described by Grout (1918e) and established as mappable units by Taylor (1964). However, within the anorthositic series and especially in the layered series, many subunits have been delineated which more completely describe the cumulus and intrusive stratigraphy of the Duluth Complex (Fig. 3.3). General descriptions of these subunits are given below and are followed by descriptions of field trip locations that highlight the main stratigraphic intervals of the Duluth Complex at Duluth.

Anorthositic Series

As observed throughout the Duluth Complex (Fig. 3.1), the anorthositic series in the Duluth area is composed of a structurally complex suite of coarse-grained, plagioclase-rich cumulates that form a stratigraphic cap to the complex (Fig. 3.3). The average rock type of the anorthositic series is an altered, coarse-grained, moderately laminated (foliated), ophitic olivine leucogabbro composed of about 80% plagioclase. Moderate to nearly complete hydrothermal alteration of olivine, pyroxene and, to a lesser degree, plagioclase is evident in all anorthositic series rocks. With the exception of some occurrences of granular olivine, subhedral to euhedral plagioclase is the only cumulus phase in these rocks. As can be observed at Stops 1-5, 1-7, and 1-9, the anorthositic series contains a broad range of rock types resulting from variations in plagioclase mode from 70-99%; mafic mineral proportions; olivine texture from granular to poikilitic; grain size from medium to very coarse; and abundance of granophyric mesostasis up to 20%. Despite these variations in texture and modal mineralogy and complex zoning of individual crystals, the

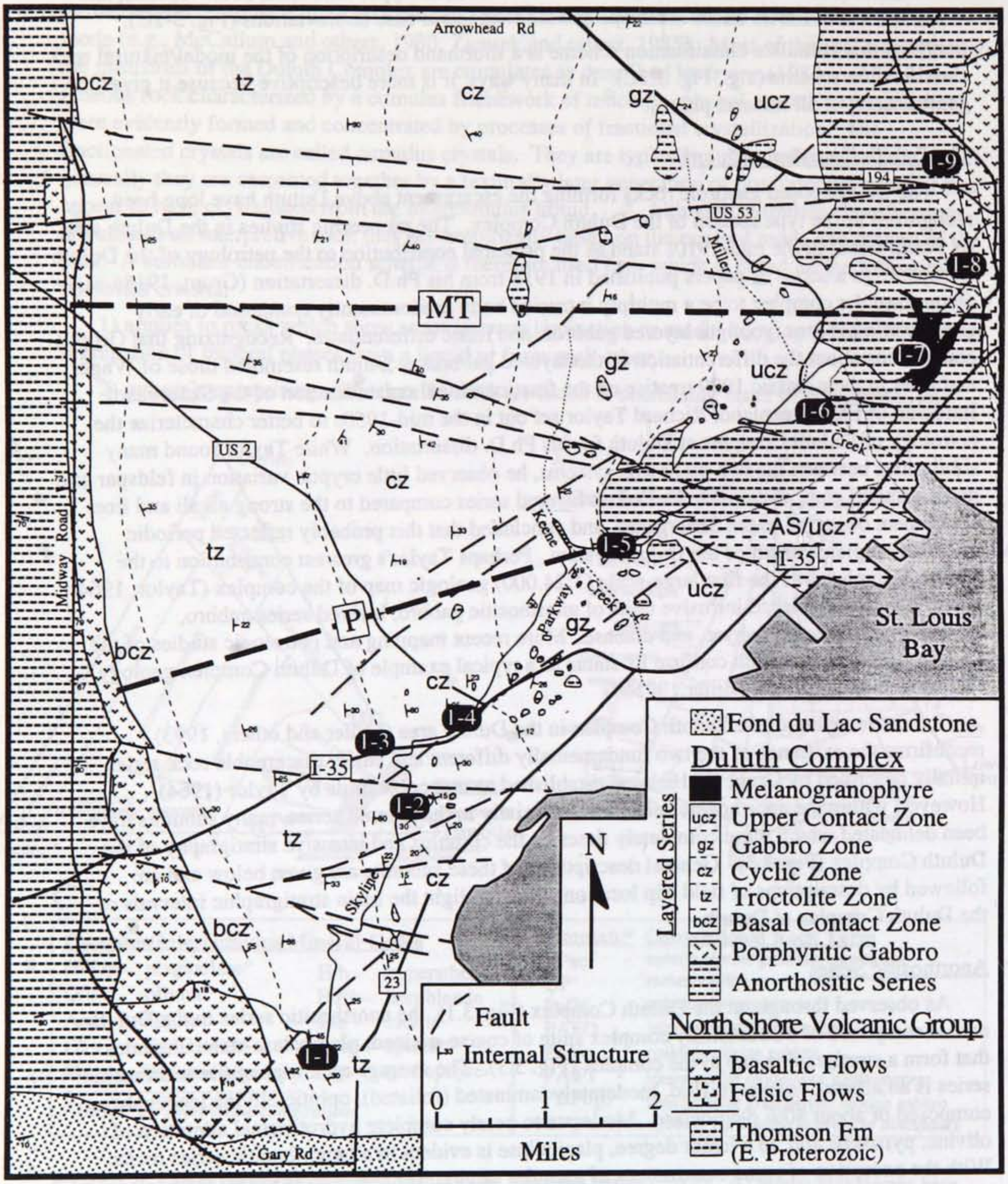


Figure 3.3: Geology of the Duluth Complex in the Duluth area showing field stop locations for Day 1. Dashed lines labelled P and MT are the Proctor and Morris Thomas profiles along which samples were analysed for their mineral chemistry and plotted in Figure 3.5. Area labelled AS/ucz represent an area of no outcrop which Taylor (1964) interpreted to be underlain by anorthositic series rocks, but which may contain layered series rocks (see text).

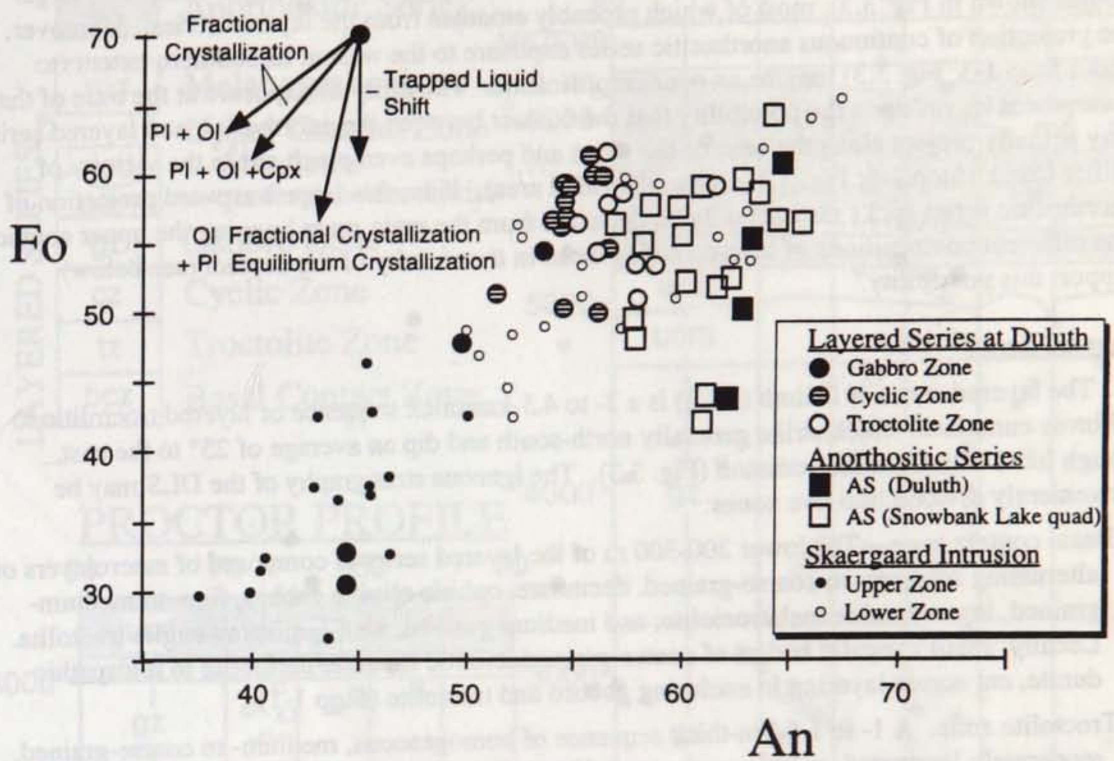


Figure 3.4: Covariation of Fo in olivine and An in plagioclase in the DLS, anorthositic series, and Skaergaard Intrusion rocks. Arrows indicate compositional effects of igneous processes that may give rise to the trends observed. DLS and anorthositic series data from Duluth are based on average of multiple microprobe analyses (Miller and others, 1991 and unpublished data). Anorthositic series data from the Snowbank Lake quadrangle in northwestern part of the Duluth Complex are from Miller (1986); Skaergaard data are from McBirney (1989).

average compositions of plagioclase remain remarkably constant at An_{65-62}. As observed among anorthositic rocks elsewhere in the complex (Miller and Weiblen, 1990; Saini-Eidukat, 1991), this constancy in An content contrasts markedly with considerable variability in the mg# of mafic silicates and defines a distinct differentiation trend of An-Fo variation compared to layered series rocks (Fig. 3.4).

A complex internal structure is perhaps the most cogent feature of the anorthositic series in understanding its formation. Although rarely layered, anorthositic rocks typically have some degree of plagioclase alignment. The attitude of this igneous lamination, however, is extremely variable on an outcrop-to-outcrop scale. Another structural complication is presented by changes in rock type that occur across sharp (but unchilled) to gradational boundaries. Discrete inclusions of one type of anorthositic rock (typically a more leucocratic composition) in another (e.g, Stop 1-5) are common. Because of these outcrop-scale structural complexities, it is difficult to divide the anorthositic series into lithologic subunits.

Map coverage of the anorthositic series in Figure 3.3 differs from that by Taylor (1964) in that many of the exposures that Taylor tied to the main mass are now interpreted to be inclusions in the layered series. Because anorthositic rocks are so much more resistant to erosion than the gabbroic and dioritic rocks which underlie and commonly include them, it seems likely that many isolated anorthositic outcrops are erosional remnants of xenoliths. The anorthositic cap itself is cut by many intrusions ranging from diabase to layered gabbro to granophyre (Stop 1-7; larger bodies shown in Fig. 3.3), most of which probably emanate from the layered series. Moreover, the projection of continuous anorthositic series exposure to the west at its southern extent (to about Stop 1-5, Fig. 3.3) may be an oversimplification. The rarity of exposure at the base of the escarpment leaves open the possibility that the contact between the anorthositic and layered series may actually project along the base of the slope and perhaps even pinch out in the vicinity of Miller Creek (Stop 1-6; Fig. 3.3, diagonally ruled area). If so, this large westward projection of anorthositic series rocks may actually be detached from the main mass hugging the upper contact. The different compositions of layered series rocks in the vicinity of the contact (see below) support this possibility.

Layered Series[1]

The layered series at Duluth (DLS) is a 3- to 4.5-km-thick sequence of layered troctolitic to gabbroic cumulates which strike generally north-south and dip an average of 25° to the east, though local variations are common (Fig. 3.3). The igneous stratigraphy of the DLS may be conveniently divided into five zones:

1) Basal contact zone—The lower 200-300 m of the layered series is composed of macrolayers of alternating medium- to coarse-grained, decussate, ophitic olivine gabbro, fine- to medium-grained, layered oxide melatroctolite; and medium-grained, well-laminated augite troctolite. Locally, small irregular bodies of coarse-grained, biotitic ilmenite peridotite to feldspathic dunite, cut across layering in enclosing gabbro and troctolite (Stop 1-1)

2) Troctolite zone. A 1- to 1.5-km-thick sequence of homogeneous, medium- to coarse-grained, moderately laminated, ophitic augite troctolite and troctolite (PO cumulates). Modal layering of cumulus olivine and plagioclase abundance are locally developed but variable in frequency, scale, modal extremes, uniformity, lateral continuity, and orientation. Modal enrichment in augite oikocrysts (1-5 cm) and Fe-Ti oxide clots (<2 cm) to as much as 15 % is very nonsystematic through the zone, and suggests periodic replenishment (and eruption?) during crystallization. This nonsystematic variation in mode is mimiced by irregular cryptic variation of olivine and pyroxene compositions (Fig. 3.5).

3) Cyclic zone. A 1-km-thick medial zone wherein cyclical variations in rock type represent a transition from dominantly two-phase (PO) to three- or four-phase (PAF±O) cumulates (Fig. 3.6). At least five major cycles characterized by gradational to abrupt PO-PAFO transitions are recognized. All boundaries between the cycles are narrowly gradational to sharp and are commonly marked by gabbroic anorthosite inclusions and microgabbro layers (Stops 1-3 and 1-4). Fundamentally, these boundaries mark a regression in the cumulus paragenesis of the DLS from PAFO back to PO. As discussed by Miller (1995b) and descriptions of Stops 1-2, 1-3, and 1-4, the progression of rock types within each macrocycle and the cumulus regressions between them is thought to be the result of low-pressure crystallization differentiation causing volatile buildup in the roof zone leading to eruption and decompression, which is then commonly followed by recharge of a more primitive magma.

1) Although kindred rocks in other parts of the complex are commonly referred to as belonging to the troctolitic series (Weiblen and Morey, 1980), we retain the term layered series given by Taylor (1964) since troctolitic rocks comprise less than half of the sequence.

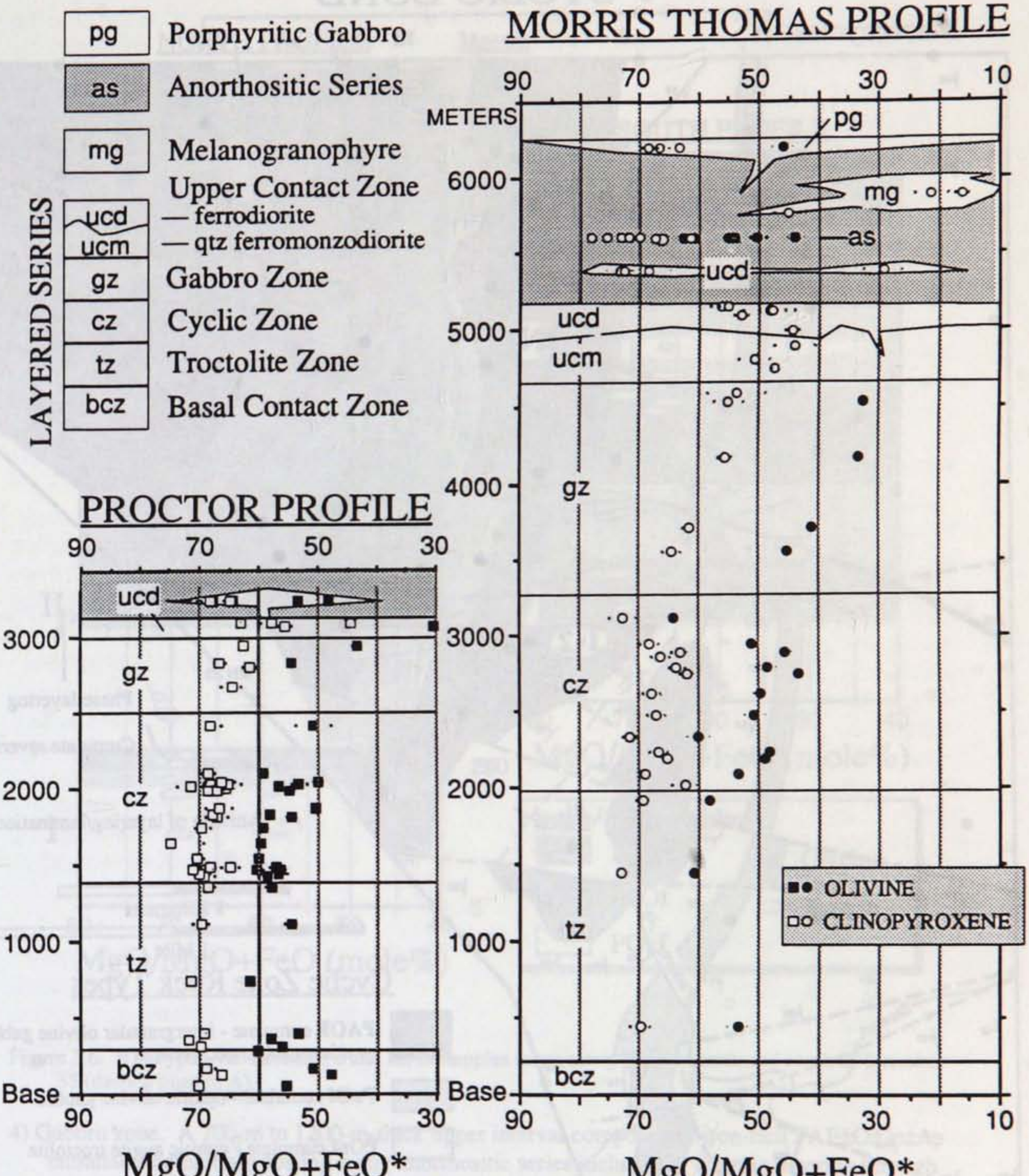


Figure 3.5: Cryptic variation of MgO/MgO+FeO (mole %) in olivine and pyroxene through the DLS. Data collected along profiles labelled P and MT in Figure 3.3. Small dots on either side of data points indicate standard deviation of multiple analyses. Plots of ferrodiorite (ucd) and quartz ferromonzodiorite (ucm) compositions represent their relative stratigraphic positions in the upper contact zone. Mineral compositions of all anorthositic series rocks (as) are grouped at a common horizon for comparison to DLS data.

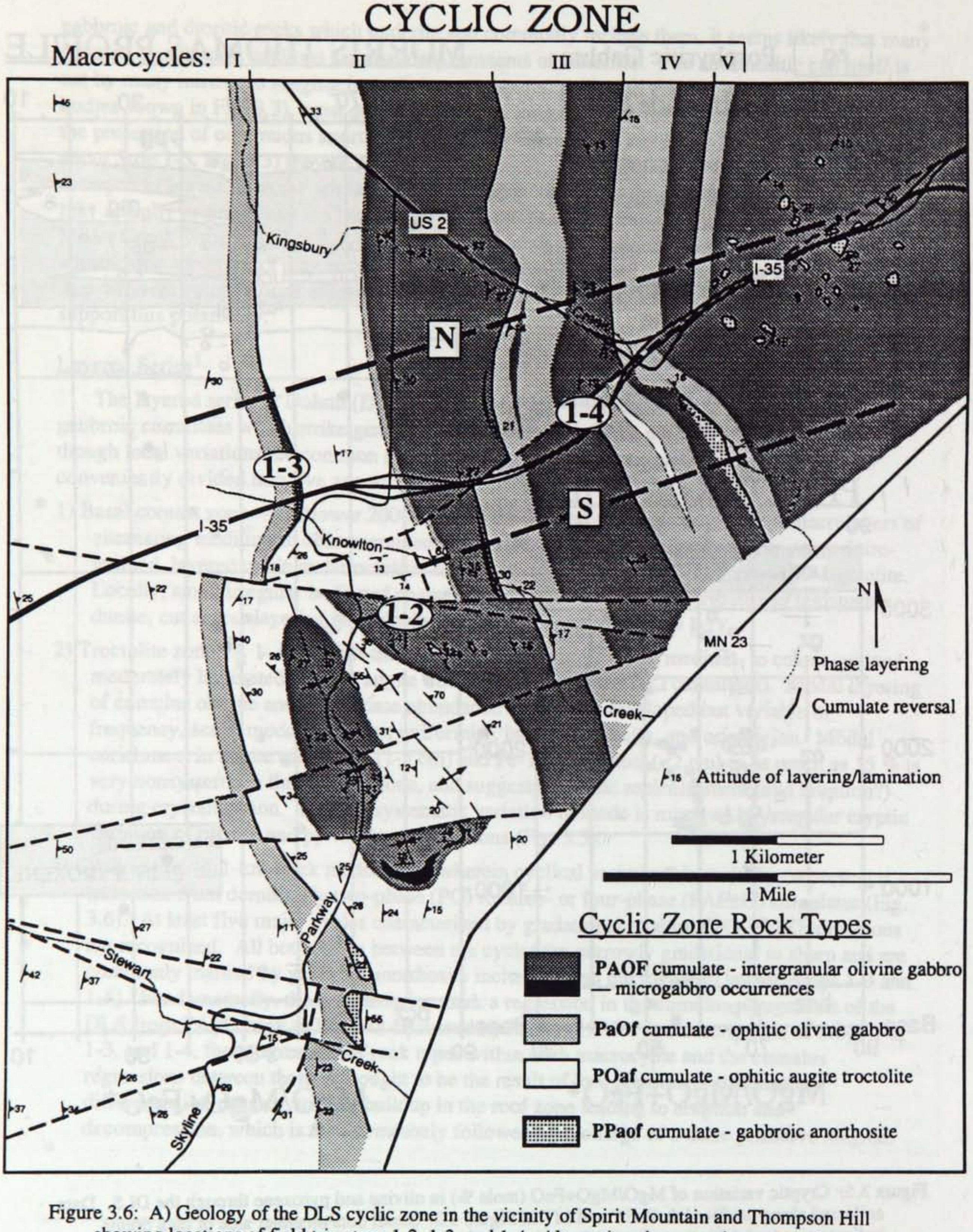


Figure 3.6: A) Geology of the DLS cyclic zone in the vicinity of Spirit Mountain and Thompson Hill showing locations of field trip stops 1-2, 1-3, and 1-4. Alternations between three general rock types define five macrocycles (I-V) as discussed in the text. Cumulate rock abbreviations as described in Figure 3.2.

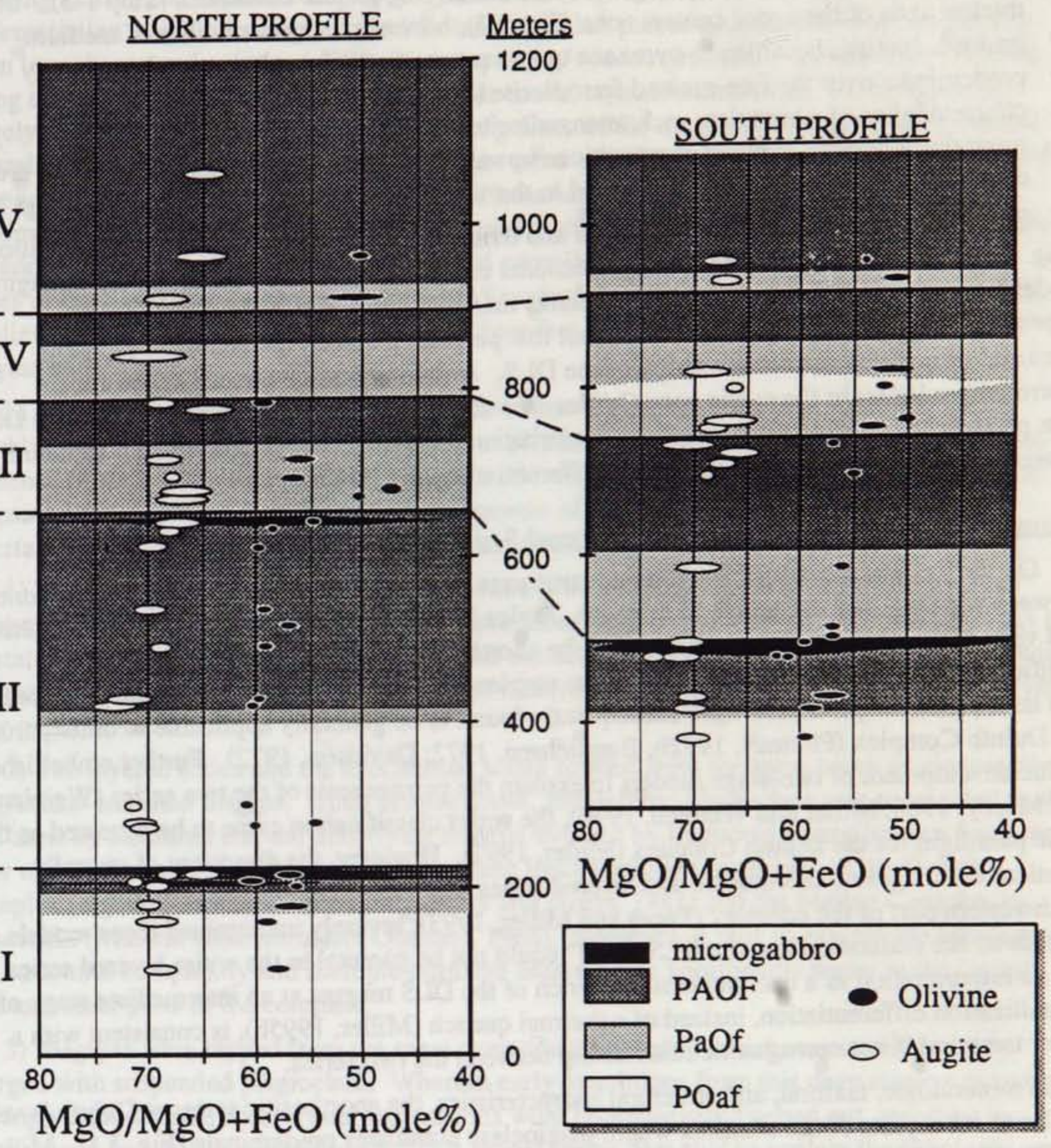


Figure 3.6: B) Cryptic variation of Fo and En in samples taken along profiles north and south of Interstate 35 (dashed lines in A).

4) Gabbro zone. A 700-m to 1200-m-thick upper interval composed of iron-rich PAF±O±Ip±Ap cumulates commonly hosting many anorthositic series inclusions. Cryptic layering through this zone varies by up to 20% in Fo and En content of olivine and augite (Fig. 3.5).

5) Upper contact zone. This uppermost zone, which hugs the irregular upper contact of DLS against the anorthositic series, contains hybridized noncumulate mixtures of fine-grained biotitic ilmenite ferrodiorite and medium-grained, apatitic quartz ferromonzodiorite. The fine-grained ferrodiorite is everywhere adjacent to the main DLS-AS contact and has been incorrectly interpreted to be a chill of the DLS parent magma against the anorthositic cap (see Stop 1-6 discussion and Miller, 1995b). At the westward extension of the anorthositic mass between Miller and Keene Creeks (Fig. 3.3), the ferrodiorite alone forms the narrow (< 50 m)

upper contact zone and abruptly grades into underlying gabbro cumulates (Stop 1-5). In the thicker parts of the upper contact zone (Fig. 3.3), however, irregular bodies of medium-grained, apatitic, hornblende-pyroxene quartz monzodiorite intrude into and eventually predominate over the fine-grained ferrodiorite (Stop 1-6). Although lacking textural characteristics of a cumulate rock, monzodiorite in the Duluth Heights area does have progressively more evolved compositions upward through the upper contact zone that are continuous with compositions observed in the underlying gabbro zone (DH profile, Fig. 3.5).

This succession of cumulate rock types and their cryptic variations in mineral composition (Fig. 3.5) are consistent with bottom-up fractional crystallization of a tholeiitic basaltic magma under low pressure. Local reversals in cumulus mineral paragenesis and cryptic layering, especially evident in the cyclic zone, suggest that periodic magma replenishment probably occurred throughout the accumulation of the DLS. A decompression quench origin for microgabbro layers in the cyclic zone (Miller, this abstract volume) further implies that the DLS was periodically open to eruption. Such perturbations to the DLS magmatic system apparently caused only minor setbacks to its overall differentiation.

Petrogenesis of Anorthositic Series and Layered Series

On the basis of the distinctive internal structures and lithologies and the sharp chilled contact between the DLS and the anorthositic series, Taylor (1964) concluded that these two rock series had very different modes of origin and that the anorthositic cumulates must have cooled signficantly by the time the DLS magma was emplaced. The series classification he proposed and its temporal implications were subsequently found to be generally applicable to other parts of the Duluth Complex (Phinney, 1972b; Bonnichsen, 1972; Davidson, 1972). Further embellished by the development of two-stage models to explain the petrogenesis of the two series (Weiblen and Morey, 1980; Miller and Weiblen, 1990), the series classification came to be accepted as the basic paradigm for the Duluth Complex (Miller, 1992). However, the discovery of virtually identical U-Pb ages of anorthositic and layered series rocks from Duluth and from the northwestern part of the complex (Paces and Miller, 1993), severely undermined these models. Moreover, the realization that the DLS "chill" could not be parental to the entire layered series, and its interpretation as a decompression quench of the DLS magma at an intermediate stage of crystallization differentiation, instead of a thermal quench (Miller, 1995b), is consistent with a closer temporal if not comagmatic relationship between the two series.

In mineralogic, textural, and structural characteristics, the anorthositic series at Duluth is very similar to other parts of the complex where plagioclase cumulates predominate (Fig. 3.1). Miller and Weiblen (1990) proposed that the anorthositic series rocks formed by the multiple injections of plagioclase crystal mushes tholeiitic basalt magmas enriched with as much as 60% intratelluric plagioclase. The most salient features of anorthositic rocks consistent with a crystal mush origin are: 1) their nonstratiform, chaotic internal structure (lamination) that is locally conformable to conformable to internal contacts (e.g., inclusions of other anorthositic rock types); 2) an apparent lack of internal differentiation; 3) a variety of zonation patterns of cumulus plagioclase in individual rock samples; and 4) an intercumulus (poikilitic) texture of all mafic phases except some olivine, which has a granular texture (even so, the relative abundance of cumulus olivine to cumulus plagioclase is strongly noncotectic). If the porphyritic gabbro in the contact zone between the anorthostic series and the hanging wall volcanics in Duluth (Fig. 3.3) is a flow-differentiated contact zone (see Stop 1-8 discussion), this would also be consistent with emplacement of a crystal mush parent for the anorthositic series.

The petrogenesis of the stratiform layered series seems to be much more straightforward and conventional. The generally unidirectional (albeit locally cyclic) progression of cumulus phase layering and cryptic variation displayed by the DLS (Fig. 3.5) can be explained by crystallization differentiation of a moderately evolved tholeiitic basaltic parent magma that was periodically open to recharge and eruption. With the exception of (decompression-) quenched ferrodiorite along the upper contact with anorthositic series rocks, crystallization of the DLS proceeded largely by cumulus mineral accumulation on the floor of the magma chamber. This fractional crystallization drove the magma along a Fenner trend of differentiation that ultimately resulted in the evolution of a high-iron and high-silica differentiate now represented by the ferromonzodiorites of the upper contact zone and perhaps by he melanogranophyre body within the anorthositic roof. The fact that no inverted cumulate sequence, such as the upper border series of the Skaergaard, developed in the roof zone of the DLS, despite its shallow crustal depth, implies that the anorthositic series must have been hot, if not partially molten when the DLS magma underplated it.

In light of the contemporaneity of the anorthositic and layered series at 1099 Ma, which notably corresponds approximately to the onset of main-stage MCR magmatism (Fig. I.3), and evidence from the Duluth area that the anorthositic series must have been quite hot when DLS magmas were emplaced, new ideas for petrogenesis of the Duluth Complex need to be entertained. Some models that may be considered are:

1) The two series represent the contributions of two deeper magma systems that evolved independently. One, perhaps deeper, system produced plagioclase-charged magmas, whereas a crystal-poor, differentiated basaltic magma was extracted from the other. A major problem with this model is that it fails to explain why crystal mush emplacement would have preceeded crystal-poor magmatism.

2) The layered series and the anorthositic series formed from the same batch of moderately plagioclase-enriched magma. Upon emplacement, intratelluric plagioclase was segregated to the roof zone by flotation, and the underlying magma evolved by fractional crystallization from the floor up. This model seems to best explain some leucotroctolitic intrusions of the Duluth Complex (e.g., the Tuscarora intrusion, Morey and others, 1981) and the Mellen Complex in Wisconsin (Mineral Lake intrusion, Olmsted, 1968). However, it fails to explain the lithologic and structural complexity and multiple intrusive nature of the anorthositic series in the Duluth area and other parts of the complex.

3) Magmas were tapped from the same deep crustal chamber which was initially highly charged with suspended plagioclase. Whereas early extractions from this deep staging chamber had high crystal loads, such crystal suspensions were progressively flushed out, resulting in successively less crystal-rich magmas. This model seems best able to explain the lithologic and structural attributes and contact relationships of the Duluth Complex at Duluth. Moreover, in that the 1099-Ma age of the Duluth Complex approximately marks the transition from the dormant stage to the main magmatic stage of MCR evolution, this model fits well with the dormant stage being a period of extensive underplating of the crust (see Miller and others, this volume; Fig. I.5).

Field Stop Descriptions

STOP 1-1: BASAL CONTACT ZONE AND LOWER TROCTOLITE ZONE, DULUTH LAYERED SERIES, DULUTH COMPLEX.

Location: Skyline Parkway near Bardon Peak, Duluth. West Duluth 7.5' quadrangle (T49N, R15W, Sec. 34, SW of NW).

Duration: 60 min.

Description: This stop examines rocks comprising the lower part of the Duluth Layered Series in roadcuts and barren knob exposures at the south end of Skyline Parkway. The rocks in this area are approximately 500 m east of a moderately dipping (~45°) basal contact of the Duluth Complex against shallow-dipping (<15°) basalt flows. The flows, termed the Ely's Peak basalts, total about 1.5 km in thickness and comprise the lowermost volcanic rocks in the Keweenawan section. About 5 km to the northwest, they lie unconformably on strongly deformed Lower Proterozoic metagraywackes and slates.

This area is situated astride the transition from the basal contact zone with its macrolayering of coarse olivine gabbro to medium troctolite the more homogeneous troctolite of the troctolite zone (Fig. 3.3). The lithologic and structural features here are interpreted to reflect the interplay of factors attending the early stages of inflation and crystallization of the DLS magma chamber. These include frequent magma injection, rapid heat loss through the footwall, active and turbulent convection, and volatile fluxing from the footwall. Figure 3.7 shows the geology of the area with eight areas of interest (A-H) noted and generally described.

The PO to OP cumulate rocks exposed at this stop include ophitic olivine gabbro, augite troctolite, troctolite, melatroctolite, and feldspathic dunite. In general, the more olivine-rich melatroctolites and dunites tend to be finer grained than the more augite-rich troctolites and olivine gabbros. These rocks display a variety of types and scales of layering. Macrolayering of rock types is evident on a meter to decimeter scale—typically from coarser grained augite troctolite to olivine gabbro and from finer grained troctolite to melatroctolite (Fig. 3.7). Cryptic differences in olivine composition are evident between macrolayers of melatroctolite-feldspathic dunite (Fo_{62-65}) and augite troctolite-olivine gabbro (Fo_{48-60}). Contacts between macrolayers are abruptly (<10 cm) to narrowly (<1 m) gradational (areas C, D, and G, Fig. 3.7). Within macrolayers, finer scaled layering is evident. Subtle grain-size layering is common in the medium- to coarse-grained augite troctolite to olivine gabbro intervals (area E). Centimeter-scale isomodal layering and decimeter-scale graded modal layering is very common in the more melatroctolitic rocks and locally has very rhythmic alternations (area F). Trough layering is locally indicated by variable orientations of modal layering (areas A and C) and the tendency for the more melatroctolitic layers to pinch out along strike. However, some of what appears to be pinch out may be due to faulting along ENE-trending structures through the area (Fig. 3.7). Very abrupt reorientation and steepening of lamination in Area G suggests that some faulting may be contemporaneous with crystallization.

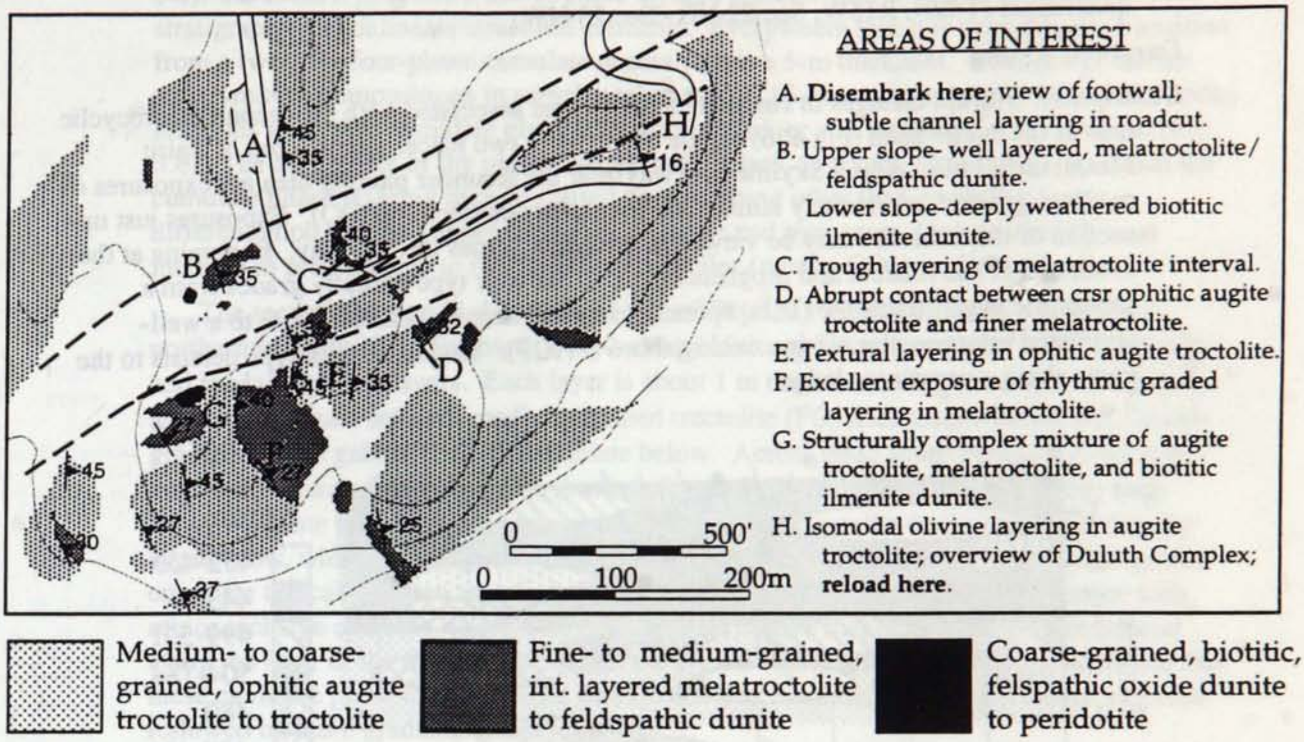


Figure 3.7: Outcrop geology of the Bardon Peak area denoting areas of interest at Stop 1-1. Dashed lines are inferred faults. Narrow lines are topographic contours portraying 50 foot intervals.

An interesting rock type in the lower part of the DLS is a deeply weathered, coarse-grained, biotitic oxide dunite to peridotite (areas B and G, Fig. 3.7). It is composed of 50-90% granular olivine, 3-20% Fe-Ti oxide (mostly ilmenite), 2-25% ophitic augite, 1-4% biotite, and 0-20% interstitial plagioclase. Primary minerals are commonly altered to chlorite, serpentine, talc, and actinolite. This rock type is found throughout the basal contact zone as small pipes or dike-like bodies and as subconformable lenses. Most occur at boundaries between coarse gabbro and finer troctolite. Olivine compositions in these bodies are Fo_{50-59} and therefore are similar to the augite troctolite and olivine gabbro cumulates in the area. Citing the crosscutting, pipe-like form of these bodies, the iron-rich composition, and the ubiquitous occurrence of secondary hydrous minerals, Ross (1985) suggested that these are metasomatic replacement bodies formed by volatile fluxing out of the footwall basalts. Severson (1994; Severson and Hauck, 1990) found similar bodies in the northwestern part of the Duluth Complex to be spatially related to iron-formation inclusion and suggested partial melting and assimilation of such inclusions may have generated the bodies. Another possible explanation is that they formed by mobilization of interstitial volatile-rich magma extracted from lower cumulates in the basal contact zone. Isotopic data would be useful in resolving the origin of these bodies.

STOP 1-2: CYCLIC ZONE, DULUTH LAYERED SERIES, DULUTH COMPLEX.

Location: Skyline Parkway and Spirit Mountain Ski Resort, Duluth. West Duluth 7.5' quadrangle (T49N, R15W, Sec 22 NE - Sec 23 NW)

Duration: 45 min.

Description: Abrupt changes in cumulate assemblages associated with the second macrocyclic unit of the cyclic zone (Fig. 3.6) can be observed in two places near the top of Spirit Mountain ski hill. Along Skyline Parkway near the Rounder parking area are exposures of medium-grained, moderately laminated, homogeneous troctolite (PO). Exposures just up-section of the troctolite may be viewed in a roadcut across the parkway. Beginning at the south end of the roadcut and progressing north, the rock type abruptly grades from a subophitic augite troctolite (POa) spotted with high-density clots of augite to a well-laminated intergranular olivine oxide gabbro (PAOF). This latter rock type persists to the north along a prominent ledge.

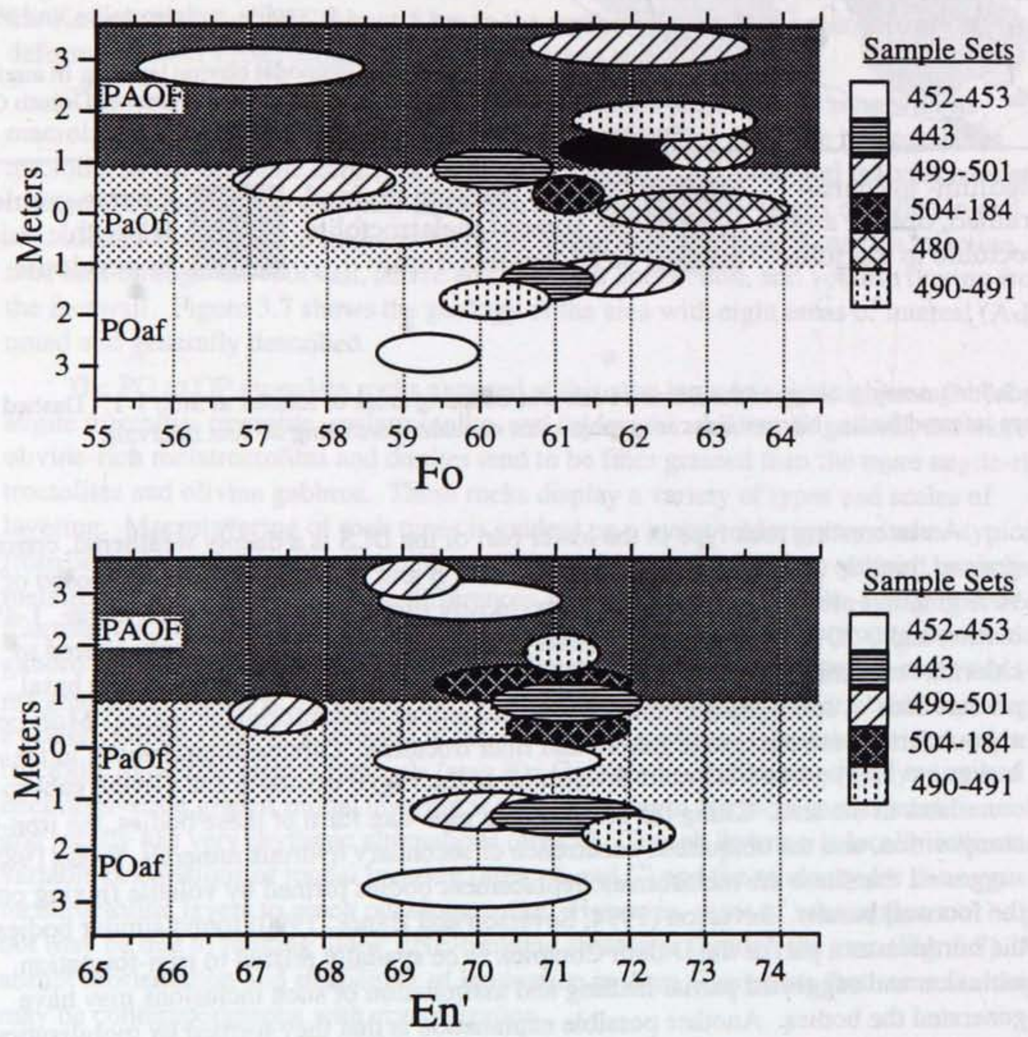


Figure 3.8: Fo and En' variation at different locations across the PO-PAFO boundary in macrocyclic unit II of the cyclic zone. Cumulate rock abbreviations as described in Figure 3.2.

The abrupt, unidirectional change in cumulus mineral assemblage observed here is characteristic of the middle part of the second macrocyclic unit throughout the area (Fig. 3.6). As such, it provides a useful phase layered horizon with which to correlate igneous stratigraphy and delineate structural elements. Everywhere that it is exposed, the transition from a two- to a four-phase cumulate occurs within a 5-m thickness. Changes in olivine and pyroxene compositions in samples taken within 15-m stratigraphically of and at various locations along this cumulate phase boundary are minor and inconsistent in direction (Fig. 3.8). The abruptness of the phase change and the lack of cryptic variation suggest that the cumulate phase layering may be caused by something other than crystallization differentiation bringing about saturation of oxide and pyroxene. One possibility is an increase in pressure due to the buildup of volatiles (mostly CO_2) in the roof zone.

The second location is about 300 m east of Skyline Parkway near the top of the northernmost chairlift. Exposed in a 10-m-high outcrop ledge are at least seven texturally and modally graded layers. Each layer is about 1 m thick and is bounded by a sharp lower and upper contact between medium-grained troctolite (PO) adcumulate above and coarse-grained olivine gabbro (PAFO) cumulate below. Across these sharp contacts, olivine is consistently more fosteritic (by as much as 3 mole% Fo) in the troctolite. Within each layer, troctolite grades upward into a subophitic augite troctolite containing high-density augite clots, which in turn grades into an intergranular olivine gabbro. The olivine gabbro cumulate typically makes up over half the layer. These relationships are consistent with minor recharge episodes whereupon new, more primitive magma pulses were emplaced along the floor of the magma chamber at the crystal-liquid interface. It is also possible that these cumulate phase changes could be explained by rapid decompression due to eruption followed by more gradual repressurization.

STOP 1-3: CYCLIC ZONE, DULUTH LAYERED SERIES, DULUTH COMPLEX.

Location: North of I-35, west of Boundary Ave., behind Skoglund Sign Co., Duluth. West Duluth 7.5' quadrangle (T49N, R15W, Sec 15 NW of SE)

Duration: 45 min.

Description: This stop examines the first transition from troctolite cumulates to olivine gabbro cumulates that defines the base of the cyclic zone (Fig. 3.6). From the driveway to Northern Engine and Supply Co., head north and uphill to outcrops of medium-grained ophitic augite troctolite (Fig. 3.9). This PO cumulate contains about 7-10% interstitial Fe-Ti oxide and augite, with the latter commonly forming 2-4 cm oikocrysts. Progressing about 70 m to the east over intermittent outcrop ledge (Fig. 3.9), the augite troctolite gradually becomes coarser grained and more enriched in augite and oxide.

Then crossing a 35-m gap in exposure, the next outcrop is of a medium- to very coarse grained, subophitic to ophitic olivine gabbro. In one area of the exposure, this rock type is in sharp, discordant, but unchilled contact with a coarse-grained, moderately laminated gabbroic anorthosite that is probably an inclusion of the anorthositic series.

Beyond an 8-m gap is a 10-m-high outcrop knob (Fig. 3.9) that grades from a medium-grained, subophitic olivine gabbro with 1/2-cm high density augite oikocrysts near the base to a texturally layered, medium- to fine-grained, intergranular olivine gabbro at the top. The textural layering is defined by 2- to 20-cm-thick, laterally discontinous layers of fine grained, moderately laminated, intergranular gabbro (microgabbro) that intermittently occurs within medium-grained, well-laminated, intergranular olivine gabbro(PAOF). Contacts between the medium gabbro and the microgabbro range from sharp to gradational on a scale of centimeters. Also in some areas, the textural layering drapes over football-sized blocks of a coarser grained, subophitic olivine gabbro, similar to the rock type seen just to the west (down section). About 150 m east of this knob, beyond a powerline cut (Fig. 3.9), low outcrops of medium-grained troctolite indicate a cumulate reversal.

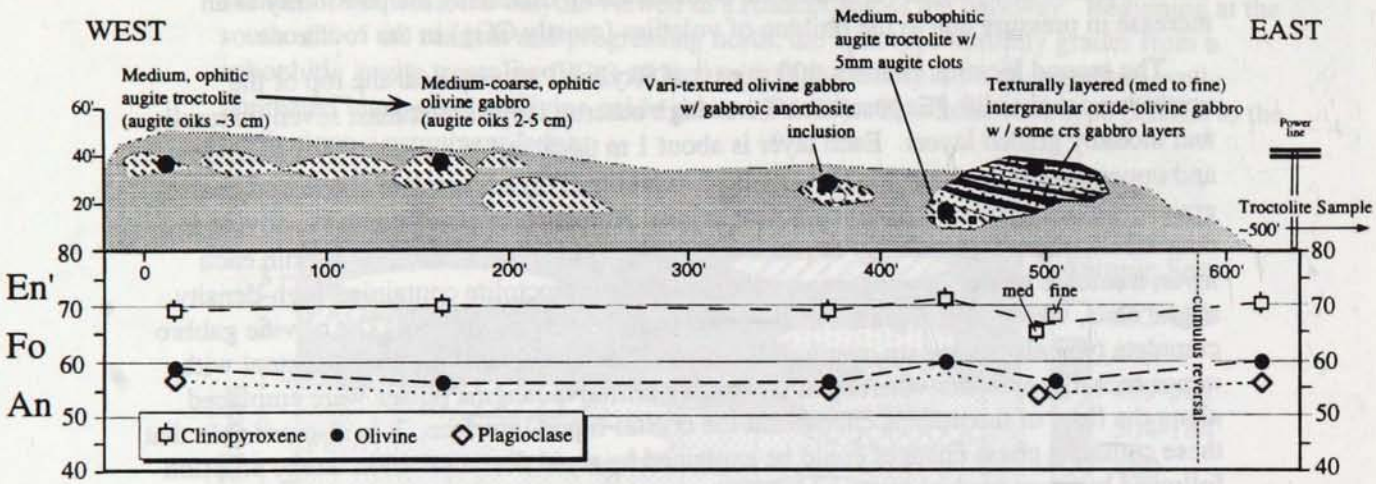


Figure 3.9: Schematic outcrop geology and cryptic variation along south facing slope at Stop 1-3. Igneous lamination and layering dip at 15-25° to the east. Exposure corresponds to the top of macrocyclic unit I marking the base of the cyclic zone (Fig. 3.6).

The cumulate stratigraphy observed here can be traced 1.5 km to the southern limit of exposure (Fig. 3.6). However, the texturally layered gabbro interval is only found here and in outcrops on the south side of the interstate highway. Elsewhere, coarse-grained, subophitic olivine gabbro, commonly containing gabbroic anorthosite inclusions, is in abrupt contact with medium-grained troctolite.

With the exception of the medium-grained, subophitic olivine gabbro below the texturally layered gabbro (Fig. 3.9), cryptic variations through this sequence show a general decrease in An, Fo, and En. Plagioclase, olivine, and augite compositions in the troctolite above the layered gabbro (Fig. 3.9) define a regression to more primitive values. Mineral compositions are also more primitive in the microgabbro layers compared to the enclosing medium-grained gabbro.

A possible interpretation of this sequence is that it represents progressive crystallization differentiation leading to multiple saturation in plagioclase, olivine, augite, and ilmenite followed by recharge of a more primitive magma causing a regression to saturation in just plagioclase and olivine. However, the occurrence of anorthosite inclusions and the microgabbro layers suggest that this recharge event may have been preceded by an eruption event from the chamber, which led to decompression quenching of magma (see Miller, 1995b and discussion at Stop 1-4).

STOP 1-4: CYCLIC ZONE, DULUTH LAYERED SERIES, DULUTH COMPLEX.

Location: Roadcut on Skyline Parkway below overlook at Thompson Hill Rest Area. West Duluth 7.5' quadrangle (T49N, R15W, Sec 14, center)

Duration: 90 min (including lunch).

Description: In the layered sequence of gabbroic rocks exposed along this 200-m-long roadcut (Fig. 3.10), a reversal in cumulus paragenesis can be seen that is characteristic of higher levels in the cyclic zone. The west end of the roadcut begins with a coarse-grained, moderately laminated, subophitic olivine gabbro, which locally is intergranular and elsewhere is leucocratic (sample A, Fig. 3.10). Augite in this rock is marginally cumulus but becomes definitely so quickly upsection where it consistently has an anhedral granular to subprismatic habit. This coarse-grained, intermittently layered (locally graded), olivine-poor (<5%) oxide gabbro (sample B) classifies as a PAFO cumulate. Beyond a poorly exposed interval, this gabbro includes a 2-m-thick interval wherein minor olivine becomes subpoikilitic and concentrated in layers (sample C). Another 3 m above this, the coarse gabbro passes into a medium-grained, well-laminated, subpoikilitic olivine-bearing oxide gabbro (samples D and D') which displays layering of olivine oikocryst concentration and elsewhere isomodal layering of Fe-Ti oxide+pyroxene abundance. The very strong alignment 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.

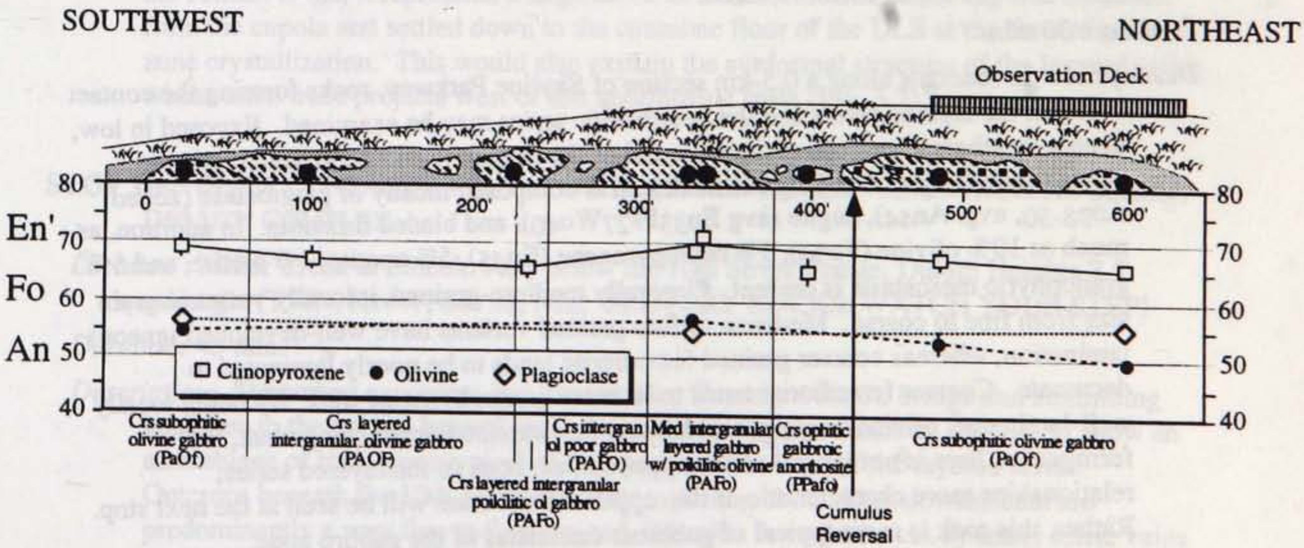


Figure 3.10: Schematic geology and cryptic variation along Skyline Parkway roadcut near Thompson Hill rest area; Stop 1-4. View is to the north. Dip of lamination and layering is exaggerated; averages about 20° to the east. Contacts between units are gradational over thicknesses of 10 cm to 1 m. Cumulate rock abbreviations as described in Figure 3.2.

As at Stop 1-3, a possible interpretation of this reversal in the cumulus status of the mafic mineral phases is that it represents a recharge of a more primitive magma that was undersaturated with respect to pyroxene and oxide. This intruding magma was emplaced and mixed with an evolved resident magma that was saturated in plagioclase, pyroxene, and oxide and undersaturated in olivine. The gabbroic anorthosite inclusions presumably mark the floor of the magma chamber at the time this recharge event occurred. However, the cryptic variation of En and Fo through this sequence (Fig. 3.10) is not consistent with such an explanation. The expected decrease in this ratio, while observed in samples A, B, and C, actually increases to its greatest value in the most "differentiated" cumulates at samples D and D'. This MgO enrichment may reflect the adcumulate nature (i.e., low trapped-liquid component) of this medium-grained, well-laminated gabbro or reflect a shift in the equilibrium compositions of mafic silicates due to the cumulus crystallization of Fe-Ti oxide. Regardless, the expected increase in MgO/MgO+FeO above the recharge horizon is not obvious in either the pyroxene or the olivine. An alternative explanation 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 and elsewhere throughout the cyclic zone. The hydrothermally altered nature of the gabbroic anorthosite is consistent with a volatile-rich environment in the anorthositic series cupola of the layered series magma chamber. The cumulus reversal to ophitic olivine gabbro without an increase in mg# could be explained by repressurization of a devolatized magma.

STOP 1-5: UPPER CONTACT ZONE, DULUTH LAYERED SERIES—ANORTHOSITIC SERIES, DULUTH COMPLEX.

Location: Roadcuts on Skyline Parkway above Oneota Cemetery. West Duluth 7.5' quadrangle (T49N, R15W, Sec 1, SE)

Duration: 30 min.

Description: In roadcuts along a 0.5-km section of Skyline Parkway, rocks forming the contact between the layered series and the anorthositic series may be examined. Exposed in low, deeply weathered outcrops at the west end of the section is an intergranular, apatitic olivine ferrodiorite (PAFOAp cumulate). It is composed mostly of plagioclase (zoned An_{38-50}, avg. An_{44}), augite (avg $En_{33}Fs_{27}Wo_{40}$), and bladed ilmentite. In addition, as much as 10% olivine (Fo_{30}), 5% orthopyroxene (En_{45}), 5% apatite, 3% biotite, and 5% granophyric mesostasis is present. Generally medium-grained, it locally varies in grain size from fine to coarse. Medium- to fine-grained variants have well-developed igneous lamination, whereas coarser grained ferrodiorite tends to be poorly laminated to decussate. Coarser ferrodiorite tends to be more granophyric, as well. By being locally well-laminated, olivine-bearing, and lacking ferromonzodioritic component, the ferrogabbro here is not typical of the upper contact zone of the layered series; relationships more characteristic of the upper contact zone will be seen at the next stop. Rather, this rock is more typical of gabbroic cumulates of the gabbro zone.

About 35 m east of the ferrogabbro, a prominent roadcut displays several common variants of anorthositic rocks that compose the anorthositic series. The western end of the roadcut is composed of a poorly to moderately laminated, poikilitic olivine gabbroic anorthosite (OGA) that contains olivine (Fo_{42-46}) oikocrysts up to 6 cm across, as well as interstitial augite and oxide. Progressively eastward, this rock type grades into a subpoikilitic to granular olivine gabbroic anorthosite. At the high point in the roadcut, this granular OGA contains an inclusion of olivine(Fo_{60})-bearing anorthosite. Whereas the enclosing OGA contain about 80-85% plagioclase (avg. An_{62}), this inclusion contains more than 95% plagioclase (unzoned An_{65}). The contact between OGA and the anorthosite inclusion is sharp with some concordant alignment of plagioclase in OGA.

Another roadcut just 5 m to the east and around a bend exposes several other varieties of anorthositic rock. Coarse-grained subpoikilitic to granular OGA at the west end of this roadcut gives way to a more leucocratic gabbroic anorthosite (Pl~90%) that locally contains 5% granophyric mesostasis. Near the center of the outcrop, the gabbroic anorthosite is cut by a 30 cm gabbroic pegmatite. Farther along,, the gabbroic anorthosite is in abrupt contact with a medium-grained leucotroctolite (Pl~75%). It is unclear whether the gabbroic anorthosite is a layer within the OGA and leucotroctolite and or an inclusion.

The inclusion relationships and the outcrop-scale variability in internal structure and lithology (mode, texture, olivine habit, and lamination development) of the anorthositic rocks here are ubiquitous features of the anorthositic series. These complexities led Taylor (1964) to characterize the anorthositic series as an igneous breccia. He and later Miller and Weiblen (1990) concluded that these rocks formed by repeated intrusions of plagioclase crystal mush.

This stop is at the western end of a westward projection of the layered series-anorthositic series contact (Fig. 3.3). A possible explanation for the irregular trace of the contact is that it reflects the original shape of the DLS roof. This would seem to be a difficult shape to maintain if the anorthositic series rocks were hot and lighter than the DLS magma. Moreover, given the truncated cryptic variation in the DLS leading up to this contact compared to the thicker DLS section to the north (compare Proctor and Morris Thomas profiles., Fig. 3.5) and the atypical character of the ferrodiorite in contact with the anorthositic series here, a more likely explanation for the westward projection of the contact is that it represents a large block of the anorthositic series cap that detached from the cupola and settled down to the cumulate floor of the DLS at the time of gabbro zone crystallization. This would also explain the synformal structure of the layered series whose axial trace projects west of this anorthositic mass (Fig. 3.3).

STOP 1-6: UPPER CONTACT ZONE, DULUTH LAYERED SERIES—ANORTHOSITIC SERIES, DULUTH COMPLEX.

Location : Miller Creek at Lincoln Park below the 10th Street Bridge. Duluth Heights 7.5' quadrangle (T50N, R14W, Sec 32, NE). SLIPPERY ROCKS; WATCH YOUR STEP!

Duration: 45 min.

Description: Streambed exposures just downstream from the 10th St. bridge and continuing upstream to the culvert beneath the intersection of Skyline Parkway and US 53 show an assemblage of rock types typical of the upper contact zone of the layered series. Outcrops beneath the 10th St. bridge and continuing about 40 m downstream are predominantly a very fine to fine-grained, ilmenitic ferrodiorite cut by many felsic veins and dikes. Between 25 and 40 m down from the bridge numerous inclusions of gabbroic anorthosite occur. Commonly , these inclusions are mantled by granophyre. Below 40 m all exposures are of gabbroic anorthosite. The ferrodiorite of the contact zone has a felty texture formed by decussate, subprismatic plagioclase, bladed ilmenite, and anhedral pyroxene and amphibole. Typically, this rock contains up to 5% subpoikilitic biotite clots, though here such clots are rare. The felsic veins and dikes cutting the ferrodiorite are typically medium-grained and intergranular, but may range in mafic mineral content (mostly Fe-pyroxene, amphibole, and iron oxide) from 5 to 30%.

Upstream from the bridge, the felsic dikes become somewhat less abundant, but otherwise the ferrodiorite remains fine-grained and unstructured. Rock type in the vicinity of the footbridge, about 250 m upstream of the 10th St. bridge, is a reddish, medium-grained, intergranular quartz ferromonzodiorite, which is composed of, in decreasing order of abundance, altered plagioclase, amphibole (primary and secondary after pyroxene), orthoclase, quartz, ferroaugite, iron oxide, biotite, and apatite. The nature of the transition from ferrodiorite to ferromonzodiorite is not clear here, though elsewhere the ferromonzodiorite is observed to intrude and be incompletely mixed with the fine-grained ferrodiorite to form a range of hybrid compositions.

TABLE 3.1: Comparison of DLS "Chill" and Cyclic Zone Microgabbro

	Analyses of DLS "Chill" samples							Microgabbro
	MD268	MD279A	MD293A'	MD247A	DG 103	DG 241	**Avg. "Chill"**	**MD455A**
SiO_2	48.3	48.5	49.1	47.2	~51	~51	**49.32**	**48.8**
TiO_2	3.35	2.97	3.10	3.49	3.92	3.74	**3.43**	**3.49**
Al_2O_3	13.3	13.1	13.2	12.5	10.6	10.5	**12.2**	**12.3**
Fe_2O_3	2.3	2.3	3.0	2.8			**2.61**	**2.6**
FeO	12.0	10.6	10.8	12.3	17.0	16.7	**13.2**	**11.9**
MnO	0.22	0.18	0.22	0.23	0.20	0.19	**0.21**	**0.22**
MgO	5.0	4.8	4.8	4.9	4.2	5.5	**4.9**	**4.6**
CaO	9.2	9.3	8.8	9.3	7.6	8.6	**8.8**	**8.5**
Na_2O	3.10	2.89	2.70	2.68	2.78	2.36	**2.75**	**2.65**
K_2O	0.42	1.05	0.93	0.71	1.19	0.47	**0.80**	**1.26**
P_2O_5	0.58	0.53	0.55	0.65	0.75	0.71	**0.63**	**0.60**
TOTAL	98.9	98.2	98.9	98.1			**98.9**	**98.5**
mg#	38.8	40.1	38.9	37.0	30.8	37.0	**37.1**	**36.4**
Trace Elements (ppm)								
Cr	110	110	100	110	59	120	**102**	**90**
Ni	76	77	73	63	55	83	**71**	**60**
Rb	15	46	34	19	<15	50	**30**	**44**
Sr	156	134	266	288	190	209	**207**	**338**
Ba	180	186	180	181	400	280	**235**	**168**
Zr	306	341	343	278			**317**	**386**
Y	56	55	56	61	88	56	**62**	**64**
La	35.6	35.6	38	33.9	61	28	**38.7**	**41.2**
Sm	10.5	10.0	11.0	10.7	14	9.2	**10.9**	**11.4**
Yb	5.0	4.9	5.2	5.2	9.4	5.5	**5.9**	**5.8**

(Analyses by ActLab, Ancaster, Ontario; Miller unpublished data)

The composition of the fine-grained ferrodiorite is similar all along the upper contact zone with the anorthositic series and is similar to some microgabbro layers in the cyclic zone (Table 3.1). The ferrodiorite is thought to represent expulsion from a shallow (~4 km deep) chamber causing decompression quenching of a volatile-rich magma in the roof zone (Miller, 1995b). The ferromonzodiorite and perhaps the granophyre are interpreted to be late-stage differentiates of the DLS system that intruded the ferrodiorite "chill." However, the presence of granophyre between the anorthosite and the fine-grained ferrodiorite suggest that some felsic melt may have been present in the roof zone at the time of ferrodiorite quenching.

STOP 1-7: ANORTHOSITIC SERIES INTRUDED BY OFFSHOOT OF DULUTH LAYERED SERIES, DULUTH COMPLEX.

Location : Roadcuts on Skyline Parkway on the south and north sides of Twin Ponds swimming area. Duluth 7.5' quadrangle (T50N, R14W, Sec 21, E central).

Duration: 60 min.

Description: Outcrops and roadcuts north and south of Twin Ponds expose contact relationships between very altered anorthositic series rocks and an irregular composite body of intermediate to felsic rocks. Exposed in the roadcut south of the T-juntion is an altered, coarse-grained, granophyric gabbroic anorthosite. All ophitic pyroxene has been replaced by fibrous amphibole, and a granophyric mesostasis locally composes 5-15% of the rock. The altered and granophyre-enriched character of anorthositic rocks is common over most of the upper part of the anorthositic series in the Duluth area (Fig. 3.3).

About 15 m east of the T-juntion, an abrupt but unchilled contact is exposed between the gabbroic anorthosite and a medium-grained, apatitic olivine ferromonzodiorite. As seen in thin section, all the olivine and most of the clinopyroxene has been replaced by amphibole, chlorite, and iron oxide, and plagioclase is commonly mantled by dusty K-feldspar which becomes intergrown with quartz to form a granophyric mesostasis. Locally the ferromonzodiorite is moderately laminated, and in one area it displays a streaky modal layering that is moderately inclined to the southeast. About 75 m east of the junction, a very complex contact between the ferromonzodiorite and a plagioclase-porphyritic quartz ferromonzodiorite is encountered which appears to be a hybrid mixture of gabbroic anorthosite and ferromonzodiorite. Seifert and others (1995) studied the petrography and geochemistry of the hybrid rock and report that it can be modeled to be a mixture of anorthositic gabbro, quartz monzodiorite, and a Na-rich hydrothermal component. Over a 25 m interval, this hybrid rock grades into variably altered and granophyric gabbroic anorthosite. With the exception of small irregular bodies of quartz monzodiorite about 140 m from the junction, the remainder of the roadcut is altered and granophyric gabbroic anorthosite.

On the north side of the Twin Ponds, a roadcut exposes a pink, mafic quartz monzonite (or melanogranophyre) cut by a 2-m-wide tholeiitic diabase dike. The melanogranophyre is composed of about 50% subhedral plagioclase, 5-10% skeletal prisms of Fe-pyroxene and amphibole, and 5% granular Fe-oxide in a felsic matrix of micrographically intergrown dusty K-feldspar and quartz. This rock type makes up 80% of the irregular composite body label as melanogranophyre in Figure 3.3 . Uphill (north) from the roadcut in intermittent pavement outcrops, the melanogranophyre can be observed to grade into more mafic compositions (quartz ferromonzodiorite to ferrodiorite) as sharp but unchilled contacts with gabbroic anorthosite are approached.

The alteration, contact relationships, and variety of rock types observed here indicate that evolved magma and hydrothermal fluids emanating from the underlying DLS passed through the anorthositic series over a protracted period of time and at a variety of scales. The very altered and granophyre-rich character of anorthositic rocks observed here and over much of the upper part of the anorthositic series suggests that late-stage melts and hydrothermal fluids fluxed through the intercumulus pore spaces of the (partially molten?) anorthositic series over a large area. In addition to this widespread percolation, the composite melanogranophyre body appears to represent a discrete conduit through which DLS-originating magmas passed and partially crystallized on their way to higher level intrusions or surface eruptions. The crude zonal distribution of composition more ferrodioritic at the margins and more quartz monzonitic in the interior—suggests that this conduit was open during crystallization of at least the upper third of the DLS. Moreover, the hybridization between the composite body and the gabbroic anorthosite host suggest that the anorthositic series was partially molten at least in the vicinity of this conduit.

STOP 1-8: UPPER CONTACT OF DULUTH COMPLEX AND NORTH SHORE VOLCANIC GROUP.

Location : Radio Tower Hill. East of corner on gravel road corresponding to 5th Ave and 9th Street. Duluth 7.5' quadrangle (T50N, R14W, Sec 28, NE).

Duration: 45 min.

Description: The east side of Radio Tower Hill overlooking downtown Duluth approximates the slope of the upper contact of the Duluth Complex with the overlying lava flows of the North Shore Volcanic Group, which now underlie all the lower ground to the northeast along the shore (Fig. 3.11). As observed at the southeastern crest of the hill, the rock exposed over most of this slope is a medium- to fine-grained, plagioclase-porphyritic (5-15%), subophitic to ophitic olivine gabbro that contains many decimeter- to meter-sized basaltic hornfels inclusions. In thin section, the gabbro typically displays intense hydrothermal alteration resulting in sericitic breakdown of plagioclase and uralitic, chloritic, and oxide replacement of pyroxene and olivine, though its primary subophitic to ophitic texture is typically preserved. The basaltic inclusions are locally intensely recrystallized and the gabbro shows little to only a weak chill around the inclusions. Less commonly, an inclusion of gabbroic anorthosite is present in the gabbro. To the west (downsection), basaltic inclusions become less common and plagioclase phenocrysts become more abundant, but in a nonsystematic way. The contact with the main suite of gabbroic anorthosite is not well exposed but seems to be abrupt and may actually finger into the anorthositic series as dikelike apophyses (Fig. 3.11).

Along a telephone line down the slope to the northeast, the gabbro gradually becomes finer grained and less porphyritic. At the base of the slope, the gabbro abruptly grades into a pink, medium-grained, intergranular to micrographic hornblende quartz monzonite (Fig. 3.11). The quartz monzonite is composed of ~15% amphibole and ferroaugite, 5% granular Fe-oxide, 40% plagioclase, 30% dusty K-feldspar, and 10% graphic to granular quartz. Abundant miarolitic cavities indicate a shallow depth of crystallization. The contact relationship between the gabbro and the quartz monzonite is best seen in a roadcut along the driveway to 415 W. Skyline Parkway. Here, the two phases, along with some basaltic inclusions, occur in a complex mixture over a 15-m-wide zone. The melanogranophyre probably formed by partial melting of intermediate to felsic lavas of the overlying North Shore Volcanic Group which form the hanging wall of the Duluth

Complex . A strongly recrystallized porphyritic felsic (icelandite) volcanic rock can be observed in low roadcuts about 40 m northeast of Skyline Parkway on W. 8th St.

A possible interpretation of the porphyritic gabbro is that it represents the upper border phase of the anorthositic series that formed by flow differentiation of plagioclase crystals away from the contact with overlying volcanic rocks. An alternative explanation is that this body represents a discrete intrusion, perhaps comagmatic with the layered series, which was emplaced at the anorthositic series-volcanic contact. Except for its highly altered state, its primary mineral compositions and texture are similar to the basal contact zone rocks of the layered series.

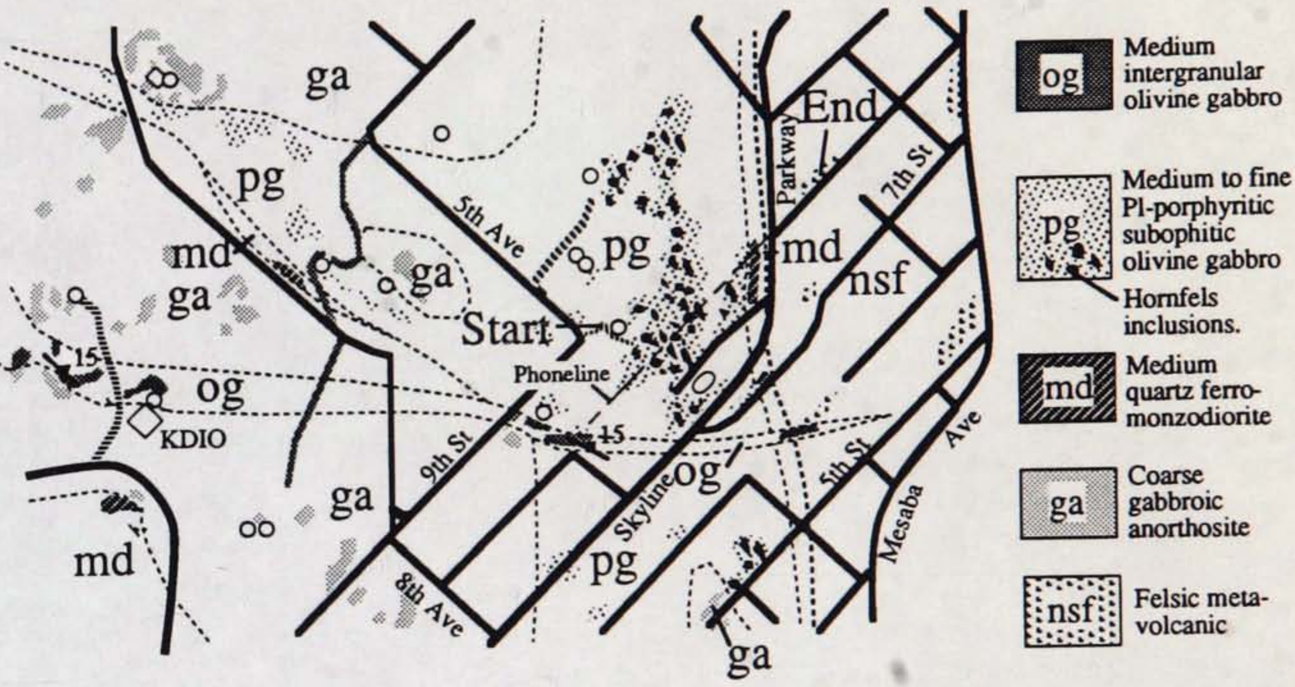


Figure 3.11: Outcrop geology of Radio Tower hill area, showing the starting and ending locations of Stop 1-8. Open circles indicate the location of radio towers.

STOP 1-9: ANORTHOSITIC SERIES, DULUTH COMPLEX.

Location : Outcrop behind Windwood Townhouses, just west of Pecan Avenue and 0.1 mile north of Hwy 194 (Central Entrance). Duluth 7.5' quadrangle (T50N, R14W, Sec 21, E central).

Duration: 30 min.

Description: This glacially scoured outcrop (and another across Pecan Avenue behind the Marshall School) shows some of the typical textural and structural features of the anorthositic series that give evidence for its multiple intrusive origin from plagioclase crystal mushes (basaltic magma laden with intratelluric plagioclase). Several intrusive cycles of erratically flow-laminated olivine gabbroic anorthosite are evident in this exposure. Intrusive relationships defined by crosscutting plagioclase lamination indicate that more leucocratic compositions are generally older. Also, some gabbroic anorthosites are graded in plagioclase concentration away from contacts with older anorthositic rocks, perhaps the result of flow differentiation of the crystal mush. An unique feature here is the occurrence of "snowball" inclusions composed of concentrated plagioclase tablets that display lamination oriented independent of that in the surrounding gabbroic anorthosite. Overall, the relationships here indicate several, irregularly shaped, intrusive pulses of gabbroic magma choked with previously crystallized plagioclase and cognate inclusions of varied size.

FIELD TRIP 3

DAY 2

THE BEAVER BAY COMPLEX

James D. Miller, Jr.
Minnesota Geological Survey

Overview

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. 3.1). 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 BBC, southern, northern, and eastern, are distinguished on the basis of distinctive rock types and intrusion form (Miller and Chandler, in press). This field trip will investigate some of the units composing the southern BBC (Fig. 3.12). The relationship of BBC intrusions to other subvolcanic intrusions within the NSVG (Fig. 3.1) 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 lease six major intrusive events (Miller and Chandler, in press). 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.

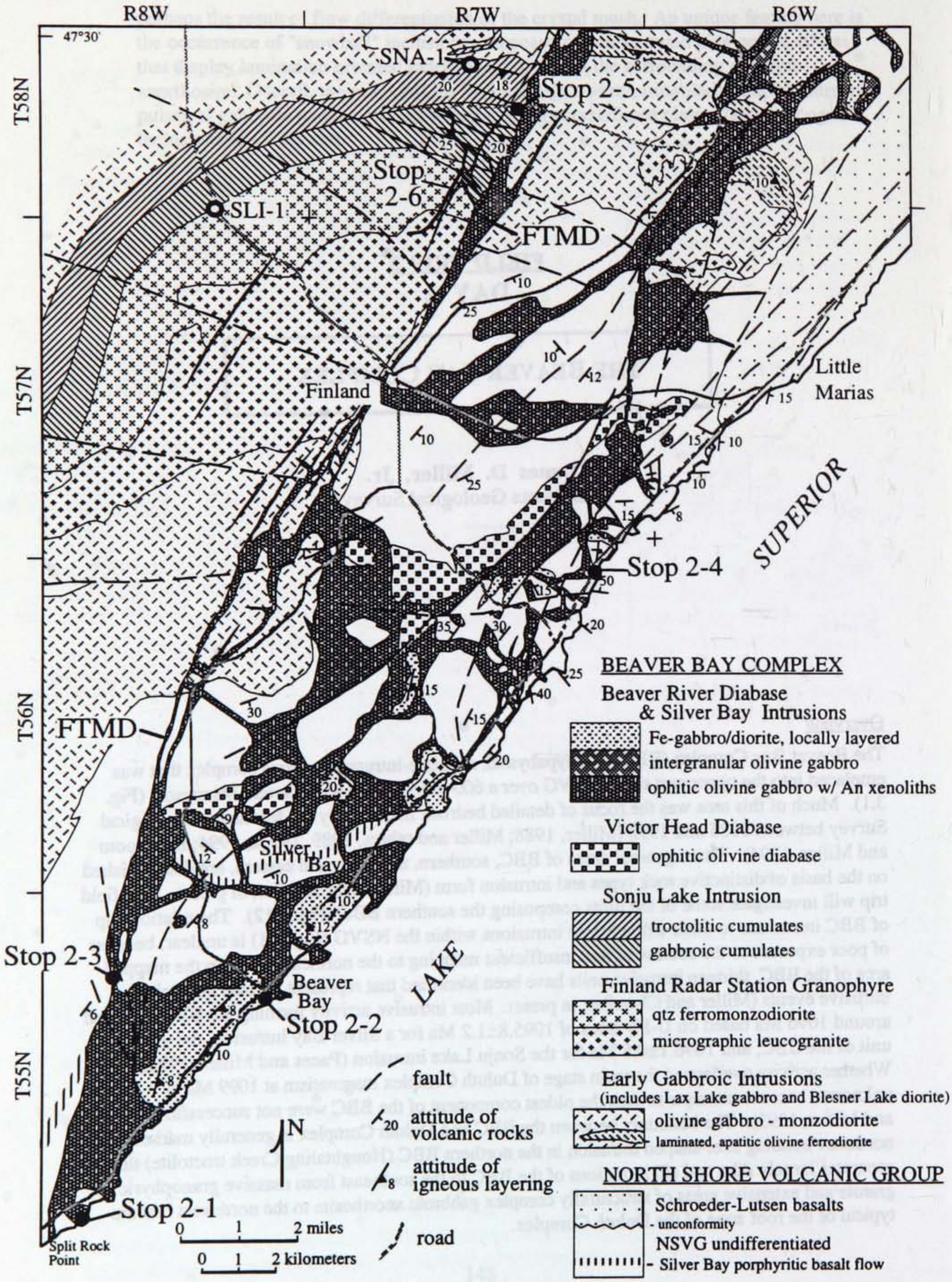

R8W
R7W
R6W
47°30'
T58N
T57N
T56N
T55N
SNA-1
Stop 2-5
Stop 2-6
SLI-1
FTMD
Finland
Little Marias
SUPERIOR
Stop 2-4
FTMD
Silver Bay
Stop 2-3
Beaver Bay
Stop 2-2
LAKE
Stop 2-1
Split Rock Point
N
0 1 2 miles
0 1 2 kilometers
fault
attitude of volcanic rocks
attitude of igneous layering
road
BEAVER BAY COMPLEX
Beaver River Diabase & Silver Bay Intrusions
Fe-gabbro/diorite, locally layered
intergranular olivine gabbro
ophitic olivine gabbro w/ An xenoliths
Victor Head Diabase
ophitic olivine diabase
Sonju Lake Intrusion
troctolitic cumulates
gabbroic cumulates
Finland Radar Station Granophyre
qtz ferromonzodiorite
micrographic leucogranite
Early Gabbroic Intrusions
(includes Lax Lake gabbro and Blesner Lake diorite)
olivine gabbro - monzodiorite
laminated, apatitic olivine ferrodiorite
NORTH SHORE VOLCANIC GROUP
Schroeder-Lutsen basalts
- unconformity
NSVG undifferentiated
- Silver Bay porphyritic basalt flow

Figure 3.12: Geology of the southern Beaver Bay Complex showing the locations of Stops 2-1 to 2-6. Generalized and slightly modified from 1:24,000-scale geologic maps of the Silver Bay and Split Rock Point NE 7.5' quadrangles (Miller, 1988), the Illgen City quadrangle (Miller and others, 1989), the Finland and Doyle Lake quadrangles (Miller and others, 1993a), and unpublished data from the Little Marais quadrangle.

The range of BBC parent magma compositions is similar to the olivine tholeiite and transitional basalt compositions that dominate the NSVG (Fig. 3.13A). 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. 3.13B), assimilation of footwall rocks, and/or composite intrusions of evolved magma from deeper staging chambers (Fig. 3.13C). The tightly clustered trend of BBC parent magma compositions evident on an AFM diagram (Fig. 3.13A) 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 MCR magmas, especially in later stages of magmatism (Green, 1983; 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. 3.13A) 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 which trends northwest across the BBC (SFCR, Fig. 3.1). 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 BBC which is situated over the gravity minimum. The eastern BBC opens up again into thick sheet intrusions. In the larger context of the Midcontinent rift, geologic, geochronologic, geophysical, and geochemical data consistently indicate 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, represented by the Schroeder-Lutsen basalts (Fig. 3.1), during the main stage of rift volcanism and graben formation (Miller and others, 1995; Miller and Chandler, in press).

This trip will investigate four intrusive units that comprise part of the southern BBC (Fig. 3.12). Stops 2-1 through 2-4 will investigate the composite intrusions of the Beaver River diabase and Silver Bay intrusions, the youngest intrusive components of the BBC, and Stops 2-5 and 2-6A-Q will investigate the upper section of the differentiated Sonju Lake intrusion and the overlying Finland granophyre.

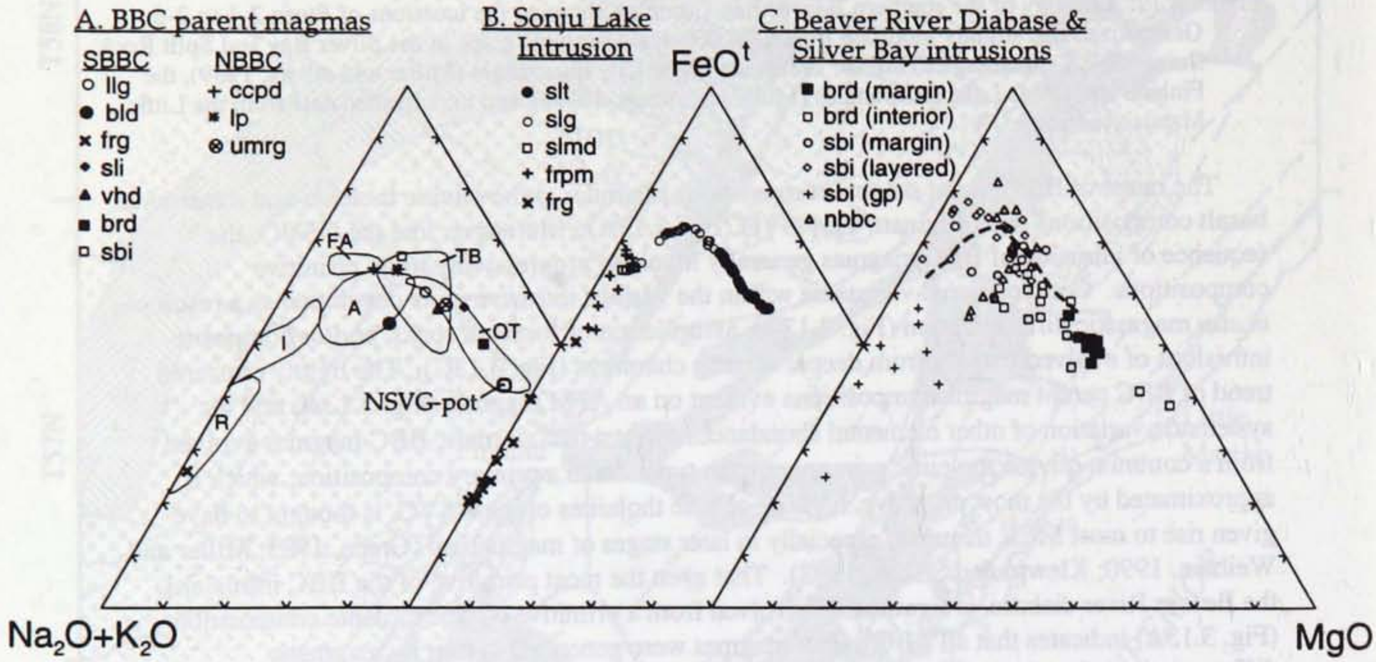


Figure 3.13: AFM diagrams of Beaver Bay Complex (BBC) intrusions. A) Plot of estimated parental magma compositions to various BBC intrusions (see Miller and Chandler (in press) for unit descriptions and details) compared to major NSVG lava compositions: OT - olivine tholeiite, TB - transitional basalt, A - andesite, FA - ferroandesite, I - icelandite, and 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 (**slt=slmt+sld+slt** units), gabbroic (**slg=slg+slfg+slad**) and monzodiorite (**slmd**) intervals of the layered sequence (see Fig. 3.16). 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 Stop Descriptions

STOP 2-1: BEAVER RIVER DIABASE WITH ANORTHOSITE INCLUSIONS, BEAVER BAY COMPLEX

Location: Split Rock Lighthouse State Park. Entrance 45 miles northeast of Duluth on Highway 61. Split Rock Point NE 7.5' quadrangle (T55N, R8W, Sec. 33).

Duration: 1 hour, 15 min.

Description: 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. 3.12), 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. With the exception of Lawson (1893), who thought the anorthosite to be the tops of deeply rooted Archean mountains around which Keweenawan rocks had flowed, most workers have long recognized their xenolithic nature, although their source has been in doubt.

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 which 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 (Table 3.2; Fig. 3.13). The prominent point just to the northeast (Rusty Point) is held up by a very large (~200 m) inclusion of medium-grained granite lying at the base of the sill. The diabase is slightly chilled around this granite inclusion, which is probably a xenolith of Finland granite.

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 [$An_{60\text{-}80}$]; 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). 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 (e.g., Stop 1-5). The highly disordered structural state of plagioclase (Miller, 1986, Fig. 4-6) 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).

To the southwest of the lighthouse along the shore, the base of the sill can be observed to conformably overlie a rubbly, silty, amygdaloidal top of a basalt flow. 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. Looking to the northeast, one can see that the near-vertically layered inclusion beneath the lighthouse extends to lake level.

STOP 2-2: SILVER BAY INTRUSION—BEAVER RIVER DIABASE, BEAVER BAY COMPLEX

Location: Near the town of Beaver Bay. PRIVATE PROPERTY, exact location not given to protect privacy of landowners. Silver Bay 7.5' quadrangle.

Duration: 60 min.

Description: The youngest intrusions recognized in the southern BBC are a suite of small, isolated bodies of varied shape and size, collectively termed the Silver Bay intrusions (Fig. 3.12). The bodies are intermediate in composition and form composite intrusions in Beaver River diabase dikes and sheets. The largest of these is located southwest of the town of Beaver Bay where it occurs as a gently southeast-dipping, concentrically-zoned, lensoid body emplaced into the upper part of a thick diabase sheet (Figs. 3.12 and 3.14). The two major zones of the body— an outer ring of coarse, vari-textured ferromonzodiorite and an inner core of strongly laminated and locally layered olivine ferrogabbro to ferromonzodiorite (Fig. 3.14)— were interpreted by Gehman (1957) to represent two distinct intrusions. However, recent mapping (Miller, 1988) and petrologic studies (Shank,1989; Chalokwu and Miller, 1992) concluded that the outer unit is a marginal facies to the inner sequence of iron-rich cumulates.

Geochemical modeling demonstrates that fractional crystallization of the Beaver River diabase could have generated the magma compositions that gave rise to the various Silver Bay intrusions. In the southern BBC, six such zoned bodies and several smaller massive bodies (Fig. 3.12) span a considerable range of intermediate compositions (Fig. 3.13), suggesting that these bodies were derived either from a deeper staging chamber at various times during its differentiation or from many different chambers at varied depths and degrees of differentiation. Upon emplacement many of the larger intrusions continued to differentiate *in situ* to produce iron-rich cumulates interior to decussate ferrogabbroic margins. (Fig. 3.14, Table 3.2).

This stop traverses the northern margin of the zoned Beaver Bay body. The traverse begins in a medium-grained, well-laminated, oxide-olivine gabbro (or diorite) that composes most of the interior of the intrusion. As observed over about 150 m of shoreline exposure, the rock is a four-phase cumulate composed of lath-shaped andesine plagioclase, subequant olivine (weathered pits), subprismatic ferroaugite, and granular Ti-magnetite. The gabbro contains 2-10% granophyric mesostasis and locally may contain significant apatite, hornblende, or hypersthene. Alignment of plagioclase and augite defines a pronounced subhorizontal lamination, but modal layering is uncommon. Thin granophyre dikes cutting the gabbro become more common as the contact with the marginal gabbro to the northeast is approached. Significant cryptic layering is evident in the laminated sequence along the shore and upslope as exemplified by a range in olivine composition from Fo_{35} to Fo_{2}. As shown in Figure 3.14, olivine in the cumulates becomes more evolved toward the margins of the intrusion. However, because lamination and layering are approximately parallel to the shoreline and dip at approximately the same low angle as the topography, very little vertical section is exposed here. This implies that cryptic layering is not concordant with internal structure.

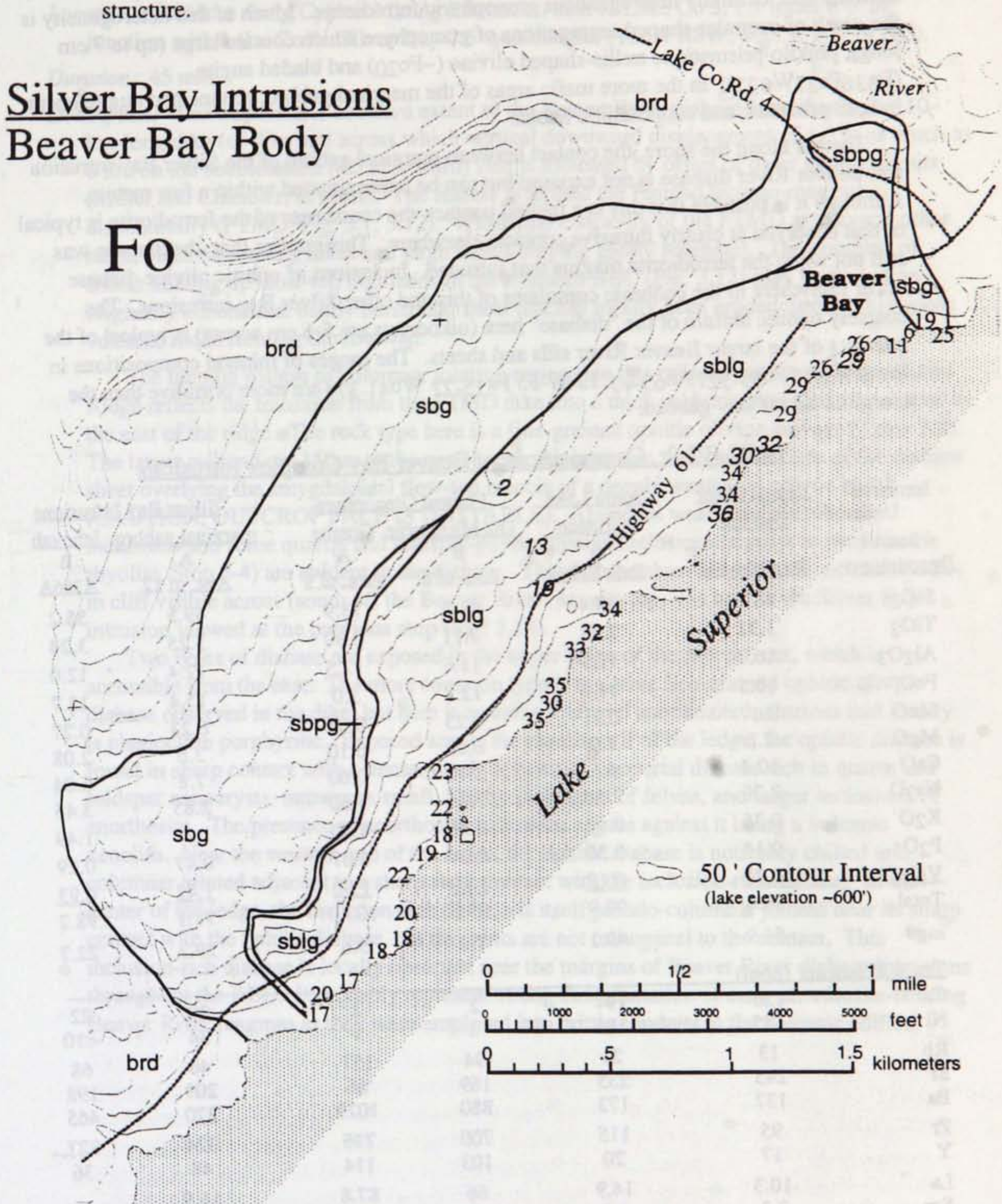


Figure 3.14: Geology of the Beaver Bay body of the Silver Bay intrusions showing the distribution of Fo contents of olivine (data from Shank, 1989 [italics] and Chalokwu, unpublished data [plain]).

The contact between the laminated oxide-olivine gabbro and the marginal rock is defined by an abrupt loss of lamination and increase in grain size. The marginal facies of this intrusion, which occurs across the next 100 m of shoreline exposure, is a coarse, decussate, modally and texturally heterogeneous granophyric ferrodiorite. Much of this heterogeneity is the result of irregular-shaped segregations of granophyre which contain large (up to 7 cm long), poikilo-prismatic to trellis-shaped olivine (~Fo_{20}) and bladed augite ($En_{36}Fs_{27}Wo_{37}$). In the more mafic areas of the marginal gabbro, olivine and augite occur in both prismatic and subpoikilitic habits.

Farther along the shore, the contact between marginal gabbro of the Silver Bay intrusion and Beaver River diabase is not exposed but can be approximated within a few meters. Although it is possible that this is a faulted contact, the coarseness of the ferrodiorite is typical of that observed at clearly intrusive contacts elsewhere. This implies that the diabase was still hot when the ferrodiorite magma was intruded. Inclusions of ophitic olivine diabase occur elsewhere in the gabbroic cumulates of this and other Silver Bay intrusions. The coarsely ophitic texture of the "diabase" here (oikocrysts are 5-6 cm across) is typical of the interiors of the larger Beaver River sills and sheets. The ranges of mineral compositions in the diabase (Fo_{59-32}; An_{65-40}; En_{47-35} Fs_{15-27} Wo_{41-37}) are more primitive than the marginal Silver Bay gabbro.

TABLE 3.2: Compositions of Beaver Bay Complex Intrusions

Intrusive Unit:	Beaver River Diabase	Sonju Lake Intrusion	Finland Granophyre: Qtz Mnzdiorite	Finland Granophyre: Granite	Silver Bay Intrusions: marginal gabbro	Silver Bay Intrusions: Jehovah
Source:	A	A	B	A	A	B
Description:	Avg.Margin	Bulk Comp	Avg. of 8	Avg. of 9	Avg. of 12	A386A
SiO_2	48.0	48.4	61.2	73.9	49.5	55.2
TiO_2	1.31	2.22	1.11	0.30	3.25	3.24
Al_2O_3	16.7	13.9	11.6	12.5	12.4	12.0
FeO_t	10.5	14.6	12.8	3.0	15.4	12.7
MnO	0.17	0.21	0.23	0.05	0.23	0.27
MgO	7.8	7.9	0.62	0.10	3.6	2.08
CaO	10.4	9.1	3.70	0.60	7.8	6.04
Na_2O	2.30	2.54	3.65	3.60	2.85	3.45
K_2O	0.35	0.68	2.86	4.90	1.17	1.49
P_2O_5	0.14	0.30	0.24	0.07	0.47	0.29
Volatiles	1.71	0.18	1.80	1.00	1.42	0.93
Total	99.4	99.9	99.8	100.0	98.1	98.2
mg#	56.9	49.1	7.9	5.6	29.3	22.7
Trace Elements (ppm)						
Cr	156	194	2	2	29	22
Ni	185	185	8	9	34	<10
Rb	13	21	94	157	40	68
Sr	245	235	169	66	209	198
Ba	137	173	880	1029	370	465
Zr	95	115	700	736	324	327
Y	17	20	103	114	48	36
La	10.3	14.9	66	87.8	16.9	-
Sm	3.3	4.2	17.1	16.7	5.0	-
Yb	2.03	2.1	10.0	9.8	3.7	-

Source: A) Miller and Chandler (in press); B) Miller, unpublished data.

STOP 2-3: FINLAND TECTONO-MAGMATIC DISCONTINUITY, BEAVER RIVER DIABASE, BEAVER BAY COMPLEX

Location: Roadcut along Cyprus Mining RR tracks, next to Lake Co 3, 1.7 miles SW of junction with Lake Co 4. Silver Bay 7.5' quadrangle. (T55N R8W, Sec. 9, SW of NE).

Duration: 45 min.

Description: 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-ward) side is indicated by offset of older geologic units (Miller and Chandler, in press). The feature is termed the Finland tectono-magmatic discontinuity (FTMD, Fig. 3.1, 3.12). 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 table-top highlands in the southern BBC (Fig. 3.12) 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 subhorizontal sheet developed to the east of the ridge. The rock type here is a fine-grained ophitic olivine diabase (Table 3.2). 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, Stop 5-4) are evident in the diabase. This diabase sheet holds up the extensive 120-m cliff visible across (south of) the Beaver River; near its top, it is host to the Silver Bay intrusion viewed at the previous stop (Fig. 3.12)

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 noticably 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 orthoganal to the contact. This inclusion-rich diabase is locally observed near the margins of Beaver River diabase intrusions throughout the BBC. It probably represent strong contamination of early anorthosite-bearing Beaver River magmas as they were emplaced into brittle conduits in the volcanic edifice.

STOP 2-4: SILVER BAY INTRUSION, BEAVER BAY COMPLEX AND NORTH SHORE VOLCANIC GROUP

Location: Roadcut on Minn. Highway 61 1.3 miles NE of junction with Minn. Highway 1 at Illgen City. Illgen City 7.5' quadrangle. (T56N R7W, Sec. 1, SW).

Duration: 30 min.

Description: A roadcut on Highway 61 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 which plunges about 15° to the east and measures about 450 m north to south (Fig. 3.15A). It is intrusive into the axial portion of a Beaver River diabase dike about 500-600 m wide, which dips steeply to the north.

On either end of the roadcut west of Highway 61 (Fig. 3.15B) are exposures of a deeply weathered, vari-textured but generally coarse-grained, decussate, apatitic, granophyric gabbro/diorite forming the margins of the intrusion (Table 3.2). Toward the center of the roadcut, this rock abruptly grades into a medium-grained, well-laminated, subprismatic, poikilitic olivine ferrodiorite. This laminated ferrodiorite typically consists of about 50-55% prismatic to lath-shaped plagioclase (An_{45-15}); 15-20% subprismatic and partially uralitized augite (EN'_{62-42}) and 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 that also contains prismatic apatite and brown amorphous alumino-silicate material. In thin section, the glassy black alteration product after iron-rich olivine, commonly termed hissingerite, 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 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 lamination. 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 indicative of fractional crystallization are noted through this sequence (Fig. 3.15B). Although the marginal gabbros of this body and the larger body at Beaver Bay (Stop 2-2) are similar in composition (Table 3.2), the compositions of olivine and augite indicate that the gabbroic cumulates of the Jehovah body are less evolved and therefore may represent exposure of a lower section of the cumulate pile.

A 1.5-m-wide dike of coarse-grained, decussate ferrodiorite, similar to the marginal rocks, cross cuts the internal structure of the laminated ferrodiorite about midway through the roadcut (Fig. 3.15B). Gehman (1957) noted similar dikes in the ferrodiorite intrusions near Silver Bay and suggested that the coarse, decussate ferrodiorite (his Black Bay gabbro) was therefore significantly younger than the laminated ferrodiorite (his Beaver Bay ferrogabbro). However, typically concentric zonation of the coarse, decussate rock grading inward to laminated cumulates and the overlapping compositions of these rocks suggest that they are comagmatic and coeval. We interpret the coarse-grained, decussate rocks to represent equilibrium crystallization of a stagnant margin around a convecting, fractionally crystallizing magma body. The occasional dikes of coarse-grained, decussate ferrodiorite internal to the zoned intrusion may represent late, minor intrusions of parental magma or remobilization of interstitial magma from within the intrusion.

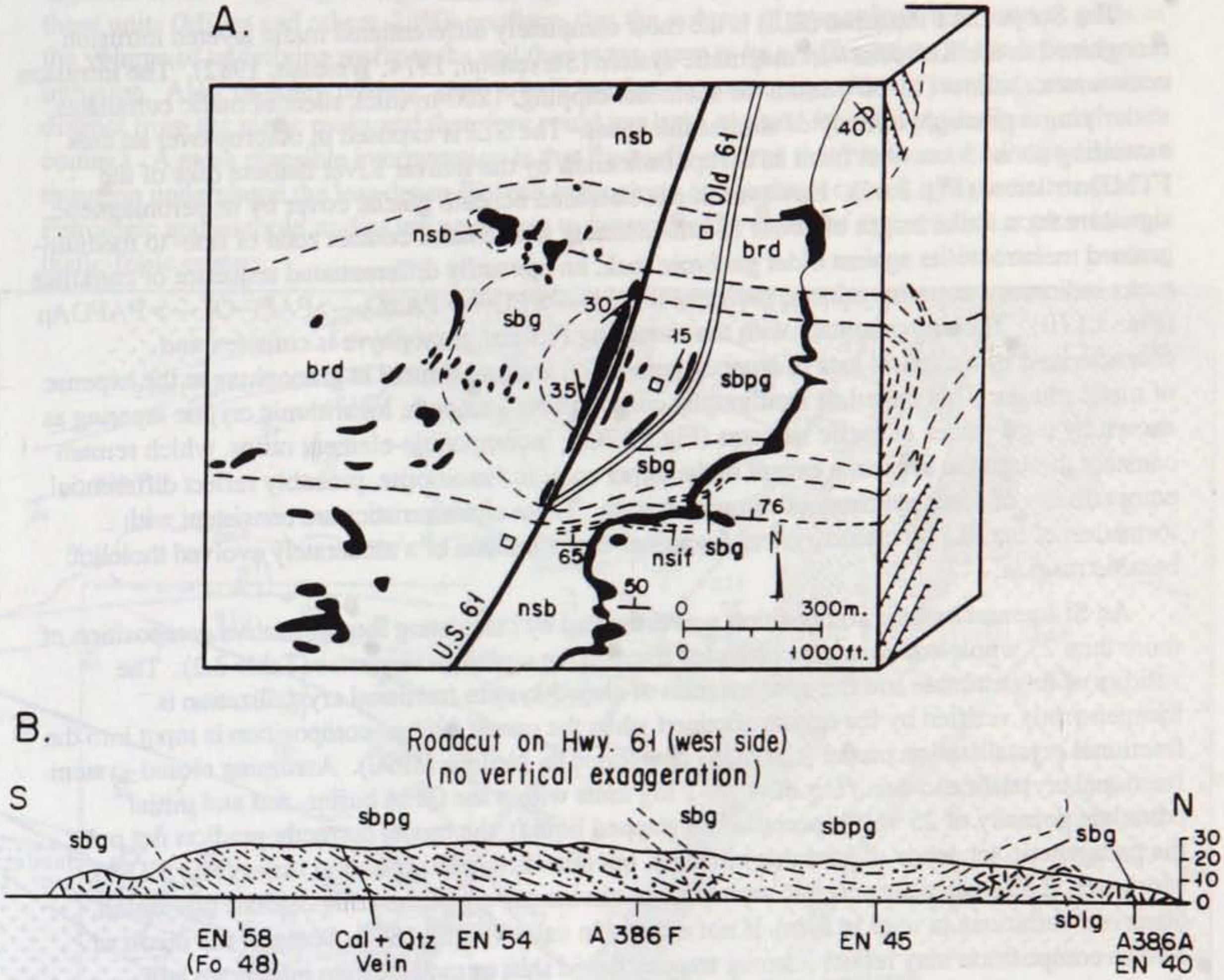


Figure 3.15: A) Geology of the Jehovah body of the Silver Bay intrusions (Stop 2-4). Outcrop areas shown in black. **sbpg** - laminated, poikilitic olivine ferrodiorite; **sbg** - coarse-grained decussate, granophyric olivine ferrodiorite: **brd**- ophitic olivine diabase (Beaver River); **nsb** - basalt flows. Interpretive cross section illustrates possible three-dimensional structure of intrusive units and attitude of volcanic flows (dots indicate flow tops). B) Sketch of the roadcut on west side of Minn. Highway 61. Tick marks represent the relative size and orientation of plagioclase. Dots represent the relative size and concentration of pseudomorphic olivine oikocrysts. **sblg**-nonpoikilitic laminated ferrodiorite. Average EN' values (= Mg/Mg+Fe) of augite and Fo of olivine are denoted as is the location of analyzed sample A386A (Table 3.2).

In the next roadcut about 400 m to the north, a sequence of seven overturned basalt flows may be examined. The flows range from rubbly-topped basaltic andesite to billowy-topped olivine tholeiitic basalt and contain a thin interval of interflow siltstone. This exposure is part of a large (400 x 700 m) block of basaltic lavas with the same orientation throughout that is surrounded by a complex network of Beaver River diabase dikes and later Silver Bay intrusions. The block faulting and rotation evident here is an extreme example of the brittle deformation that attended the emplacement of these intrusions.

Overview of the Sonju Lake Intrusion and Finland Granophyre

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. 3.16). However, it can be traced beneath glacial cover by its aeromagnetic signature for a strike length of about 15 km. Starting with a basal contact zone of fine- to medium-grained melatroctolite against older gabbroic rock, an upwardly differentiated sequence of cumulate rocks indicates a cumulus mineral paragenesis of O—> PO—> PA±O —>PAF(~O) —> PAFOAp (Fig. 3.17B). The upper contact with the overlying Finland granophyre is complex and characterized by a cyclical loss of igneous lamination and enrichment in granophyre at the expense of mafic phases. This cumulate stratigraphy compliments a smooth, logarithmic cryptic layering as shown by mg# values of mafic silicates (Fig.3.17C). Incompatible-element ratios, which remain constant through the sequence except in the upper apatitic ferrodiorite, probably reflect differential compatibility of some elements in cumulus apatite. These characteristics are consistent with formation of the SLI by closed-system fractional crystallization of a moderately evolved tholeiitic basaltic magma.

An SLI parent magma composition was estimated by calculating the cumulative composition of more then 75 whole-rock samples collected through the cumulate sequence (Table 3.2). The validity of this estimate and the interpretation of closed-system fractional crystallization is independently verified by the results obtained when the parent magma composition is input into the fractional crystallization model (CHAOS) developed by Neilsen (1990). Assuming closed-system fractional crystallization, an fO_2 of -1 to -2 log units within the QFM buffer, and and initial cumulate porosity of 25 vol% (occupied by trapped liquid), the model correctly predicts not only the paragenetic sequence of cumulus minerals, but also their relative arrival times (Fig. 3.17B). Moreover, it also calculates cryptic variations of olivine and pyroxene compositions that match observed variations in mg# in form, if not exactly in value (Fig. 3.18C). Some of the misfit of olivine composition may reflect a strong trapped liquid shift or may indicate inaccurate bulk distribution coefficients.

Overlying the SLI and exposed over a 40-km^2 area in the northwestern part of the southern BBC (Fig. 3.12) is a distinctive, oval-shaped mass of leucogranite and quartz ferromonzodiorite, termed the Finland Radar Station granophyre Miller and others (1993a), or simply the Finland granophyre. The main phase of the granophyre is a homogeneous, salmon-colored, micrographic leucogranite with abundant miarolitic cavities. 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. The intrusive relationships of the Finland granophyre, the older Lax Lake gabbro along its southern margin, and the younger Beaver River diabase at its eastern extent are well-established by several exposures of sharp contacts. However the genetic relationship of the granophyre to the Sonju Lake layered intrusion along its gradational northern boundary is more problematic. Stevenson (1974) considered the monzodioritic and granitic phases of the granophyre to be intrusive into the Sonju Lake intrusion, although he cites no evidence for that interpretation. Given the parallel zonation of the two phases of the granophyre with the strike of Sonju Lake cumulate units (Fig.3.12; also see Miller and others, 1993a) and the generally smooth compositional variations across these units (Fig. 3.17), a comagmatic relationship resulting from crystal fractionation and magmatic 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 (Fig. 3.13). 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, ongoing isotopic studies indicate that the granophyre has an isotopic composition distinct from the mafic rocks and therefore could not have evolved from them (A. Basu, oral comm.). A more plausible interpretation is that the mafic magma that produced the Sonju Lake intrusion underplated the less dense Finland granophyre and perhaps caused it to partially melt. Petrologic and isotopic studies are underway to determine the extent of assimilation across the mafic-felsic contact.

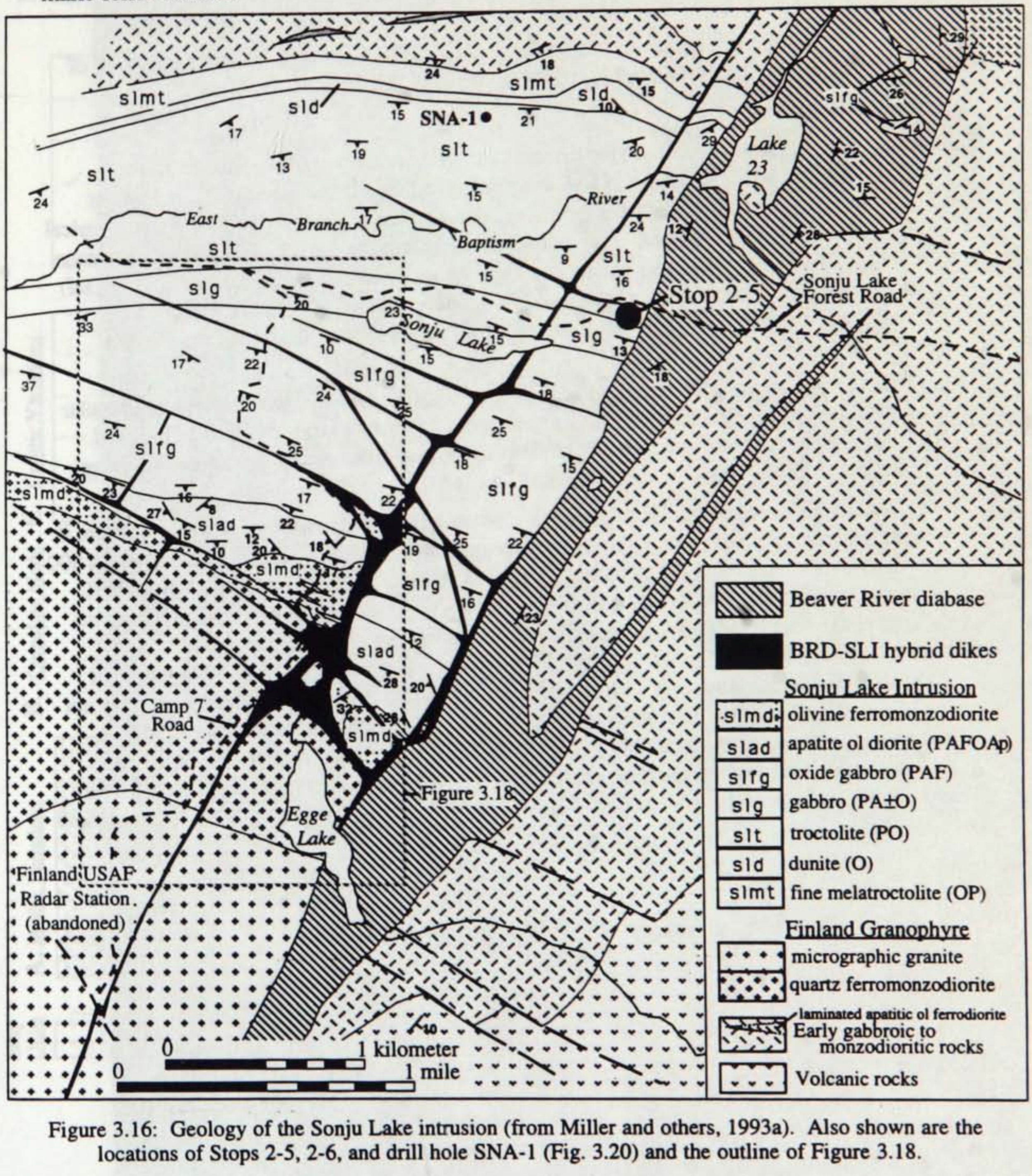


Figure 3.16: Geology of the Sonju Lake intrusion (from Miller and others, 1993a). Also shown are the locations of Stops 2-5, 2-6, and drill hole SNA-1 (Fig. 3.20) and the outline of Figure 3.18.

Stops 3-5 and 3-6 are intended to provide a detailed view of the upper half of the Sonju Lake intrusion and its transition into the overlying Finland granophyre. At the conclusion of the Stop 3-6 traverse, drill core will be on display that shows the rock types of 1) the lower contact zone including units **slt, sld, slmt** and the footwall gabbro (SNA-1, Fig. 3.16) and 2) the complex upper contact zone between the Finland granophyre and the apatite ferrodiorite (SLI-1; Fig. 3.12).

STOP 2-5: PO—PAO PHASE BOUNDARY, SONJU LAKE INTRUSION, BEAVER BAY COMPLEX

Location: Sonju Lake Forest Road, 3 miles west of Lake Co 7 and about 9 miles northeast of Finland. Finland 7.5' quadrangle. (T58N R7W, Sec. 27, NE).

Duration: 30 min.

Description: At this stop we will see the lowest easily accessible part of the SLI cumulate sequence. (Fig. 3.16). Exposed in low outcrops on either side of the Sonju Lake Forest Road is coarse-grained, moderately laminated, ophitic olivine gabbro (PaOf cumulate) typical of the upper part of the **slt** unit. Augite oikocrysts as large as 8 cm in diameter are common. Downsection from this horizon, the rock is more typically an ophitic augite troctolite (POaf) with olivine about twice as abundant as augite and the latter as 1-3-cm oikocrysts. The increase in mode and size of augite oikocrysts reflects the approaching saturation of the magma in augite.

The arrival of cumulus augite, as indicated by its abrupt change from ophitic to granular habit, can be observed in intermittent outcrop about 70 paces (~100 m) south of the road. This cumulus phase transition from PO to PAO is unique and is traceable across the entire exposure area of the SLI (**slt-slg** contact, Fig. 3.16).

The subtle valley east of these exposures marks the eastern boundary of the SLI where it is truncated by a Beaver River diabase dike occupying the FTMD. The dike here is about 500 m wide. On the other side of the dike is dioritic to melagranophyric rocks of an early BBC intrusive suite termed the Blesner Lake diorite. The only rocks similar to SLI cumulates on the east side of the FTMD are laminated, apatitic olivine ferrodiorite about 2.5 km north-northeast of this location (Figs. 3.12 and 3.16). These rocks are similar to the uppermost PAFOAp cumulates (**slad** unit) of the SLI (Fig. 3.16), but lower SLI-like cumulates are not found downsection of the ferrodiorite. If the ferrodiorite is correlative with the **slad** unit, this implies vertical displacement of 1.5 to 2 km on the FTMD and requires that a major unexposed fault has cut out the lower part of the cumulate section east of the FTMD (Miller and Chandler, in press).

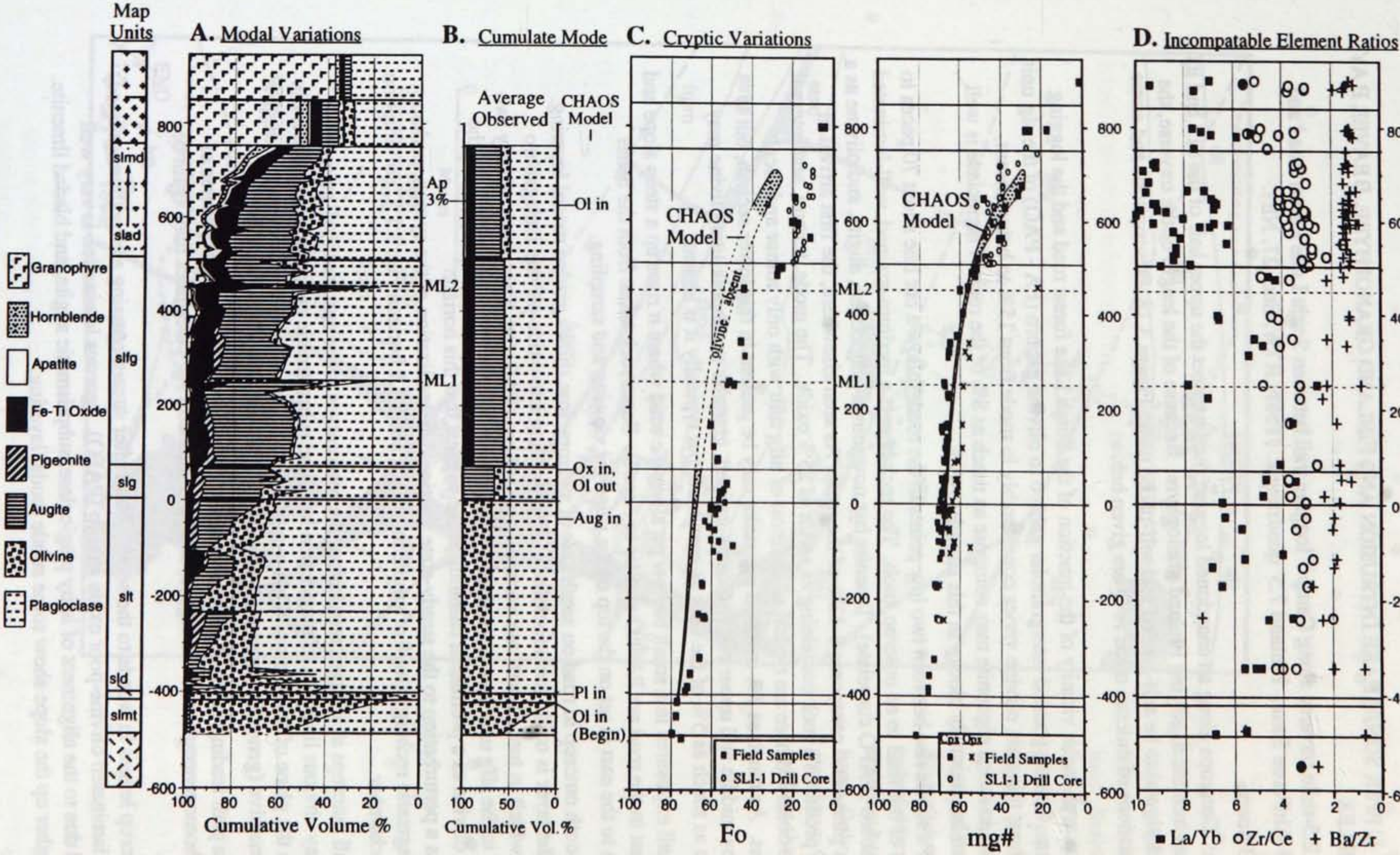


Figure 3.17: Stratigraphic variations of (A) modal mineralogy, (B) average observed mode of cumulate units, (C) cryptic variations in mg# of mafic silicates, and (D) variations of La/Yb, Zr/Ce, and Ba/Zr determined from whole-rock analyses in the Sonju Lake intrusion. Map units as in Figure 3.16 and described in text. Stratigraphic height of samples collected along five sampling profiles and from two drill cores are measured relative to the horizon marking the cumulus arrival of augite. Model results of the cumulus arrival and compositions of cumulus minerals were calculated by applying the SLI bulk composition (Table 3.2) to the CHAOS fractional crystallization model of Neilsen (1990) assuming closed-system conditions, 25% trapped liquid, and fO_2 = -1 to -2 log units from QFM.

STOP 2-6: UPPER SONJU LAKE INTRUSION AND FINLAND GRANOPHYRE, BEAVER BAY COMPLEX

Location: 2.5-mile traverse along Camp 7 logging trail between Sonju Lake Forest Road and Finland AirForce Base. Finland 7.5' quadrangle. (T58N R7W, Sec. 27, NE).

Duration: 3 hours.

Description: Outcrops along an abandoned logging road transect the upper half of the SLI and its transition into the overlying Finland granophyre. Because of the length of the traverse, the stop is designed to be self-guided and self-paced using Figure 3.18. Descriptions and some interpretations of outcrop areas A-Q are given below:

Area A: Outcrops in the vicinity of the junction of the Sonju Lake forest road and the logging road area. Coarse-grained intergranular gabbro to olivine gabbro (PA - PAO) of the **slg** unit. Throughout this unit, olivine varies considerably in mode from 15% to being absent. Interstitial inverted pigeonite may compose as much as 8% of the rock. Plagioclase is well laminated but uniquely blocky in this cumulate rock.

Area B: On a subtle rise between two low points in the road, follow a cut line about 70 paces to the west of the road to an outcrop knob. The rock here is a medium-grained, well-laminated oxide gabbro (PAFO cumulate). Ilmenite has now joined plagioclase, augite, and olivine as a cumulus phase and composes 8-12% of the rock. As seen elsewhere, the first arrival of ilmenite produces a rock containing as much as 25% oxide. The mode, texture, and internal structure observed here are typical of the most of **slfg** unit with only minor and local variations. As much as 5% inverted pigeonite may be present in this lower section, but is rare in the middle and upper parts. Coarse-grained, granular to subpoikilitic olivine may compose as much as 5% of the rock as seen here, but typically it is absent.

Area C: Small exposures in a small borrow pit along the road where it rises up a steep slope and at the crest in the road are PAF±O similar to Area B. Some rock taken from the better exposure to the east is set on the top of the ridge for viewing and sampling.

Area D: Smooth outcrop in road on south side of swampy low shows graded modal layering. Base of the layer is oxide melatroctolite (OPFa) passing into oxide-olivine melagabbro (APFO) which in turn passes into normal oxide gabbro (PAF±O). This is one of only two intervals in the **slfg** unit that show significant modal layering. This observation and the return of olivine as a prominent cumulus phase suggest that this horizon represents a perturbation to the steady-state fractional crystallization of the system, perhap due to magmatic recharge (mg# of augite increases, see ML1 horizon Fig. 3.17C) or eruption from the chamber.

Area E: Small outcrops at a ridge crest, before the road curves to the east, are PAF cumulates that display obvious lineation of plagioclase laths and subprismatic augite trending down-dip (south) in the plane of lamination. This would suggest that convection was operative despite the very massive (generally unlayered) character of the cumulates.

Area F: Up a path leading north of the trail before it crosses a stream, good exposures of massive, homogeneous, well-laminated, medium-grained PAF cumulates may again be observed.

Area G: Outcrop ledges in woods to the south of trail after stream crossing are medium-grained, very well laminated olivine-poor oxide gabbro (PAFO). Igneous lamination is very well developed due to the alignment of lathy plagioclase, subprismatic augite, and bladed ilmenite. Ledges higher up the slope show some subtle modal layering.

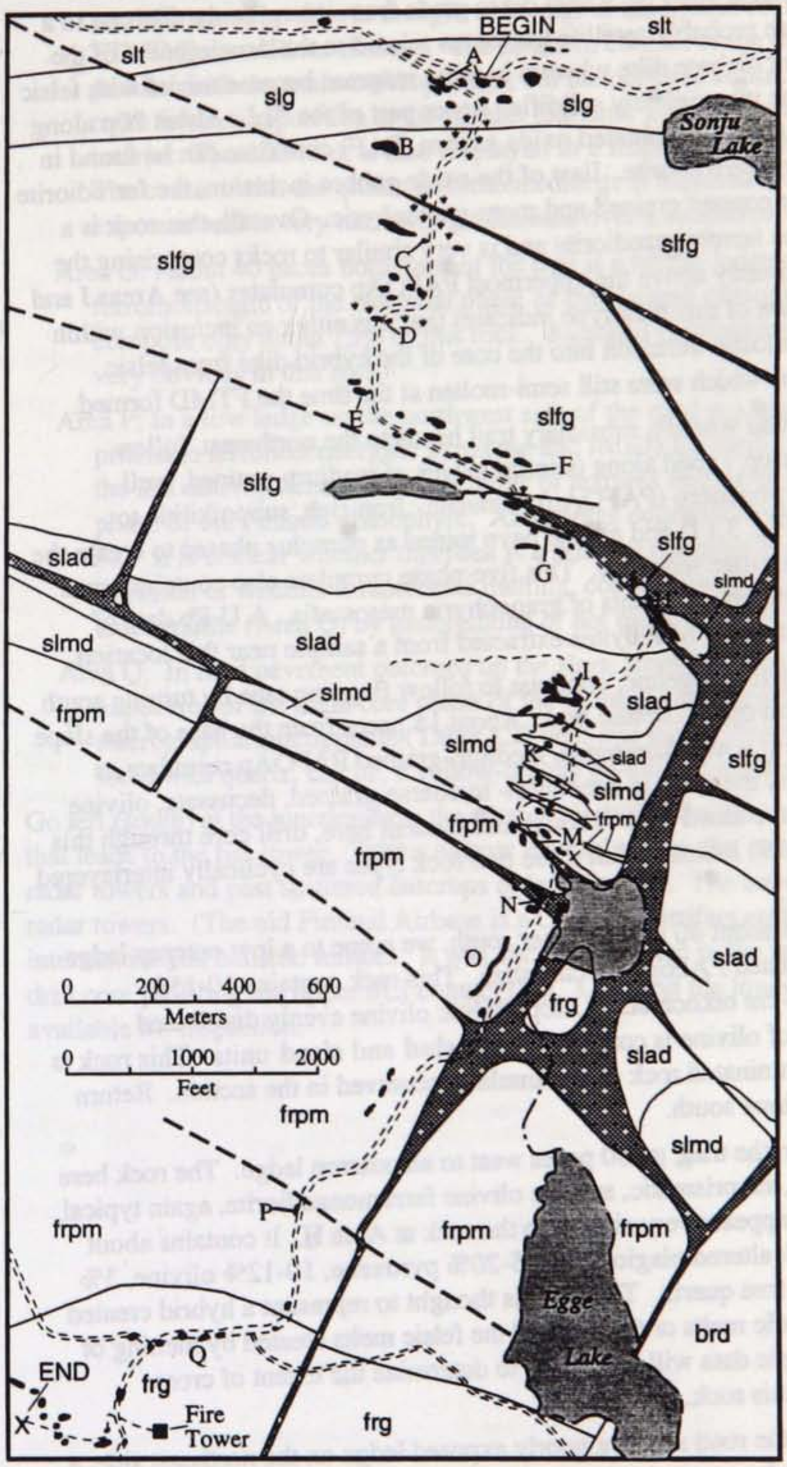


Figure 3.18: Locations of outcrop areas A-Q for Stop 2-6 traverse along the Camp 7 road. Sonju Lake intrusion units as in Figure 3.16. Finland granophyre units : **frpm** - prismatic quartz ferromonzodiorite and **frg** - leucogranite; **brd** - Beaver River diabase. BRD-SLI hybrid dikes shown by dark shading; faults by dashed lines. Only outcrops (black areas) near Camp 7 logging road are shown.

Area H: Traversing west to east for about 80 paces along a 2- to 4-m-high outcrop ledge just north of the trail, one encounters several different rock types. The western third of the exposure is composed of a fine-grained intergranular quartz ferrodiorite that is typical of a network of hybrid dikes that orthogonally cut the eastern part of the SLI (Fig. 3.16). A similar rock will be observed at Area N. These dikes grade from intergranular diabase to a mafic granophyre. They are probably satellite intrusions related to the development of the FTMD-hosted Beaver River diabase dike wherein basaltic magmas became mixed with felsic melts that still resided in the incompletely solidified upper part of the SLI. About 20p along the slope, a 5-m-wide inclusion of laminated oxide gabbro (PAF) cumulate can be found in sharp contact with the quartz ferrodiorite. East of the oxide gabbro inclusion, the ferrodiorite nonsystematically becomes coarser grained and more granophyric. Overall, this rock is a vari-texured, apatitic olivine ferromonzodiorite and is very similar to rocks comprising the **slmd** unit that is situated directly above the uppermost PAFOAp cumulates (see Areas J and L). The stratigraphic position of this rock type indicates that it is either an inclusion within the hybrid dikes or is a composite intrusion into the core of the hybrid dike from felsic magmas in the SLI roof zone which were still semi-molten at the time the FTMD formed.

Area I: From the crest of the road where a subsidiary trail heads to the northwest, follow flagging tape west of the Camp 7 road along outcrop ledges of medium-grained, well-laminated apatitic olivine ferrogabbro (PAFOAp) cumulate. Iron-rich, subpoikilitic to granular olivine (Fo_{22-10}, Fig. 3.17) and apatite have joined as cumulus phases to create the most evolved rock of the cumulate sequence. This five-phase cumulate also contains some interstitial hornblende and variable amounts of granophyric mesostasis. A U-Pb date of 1096.1±0.8 Ma was determined for badellyites extracted from a sample near this location.

Area J: After 40-50 paces along the ridgeline, continue to follow flagging tape by turning south and descending the ridge to an open forested area. About 15 paces from the base of the slope are two small outcrops. One is a well-laminated, medium-grained PAFOAp cumulate as observed just to the north, and the other is a medium- to coarse-grained, decussate, olivine ferromonzodiorite typical of the **slmd** unit. Though not evident here, drill core through this part of the upper SLI (Fig. 3.19) indicates that these two rock types are cyclically interlayered with one another.

Area K: Following flagging tape about 40 paces farther south, we come to a low outcrop ledge of moderately to poorly laminated PAFogp Ap cumulate. This rock contains 10-15% granophyric mesostatis and 1-cm oikocrysts of subpoikilitic olivine evenly distributed through the rock. This habit of olivine is common in the **slad** and **slmd** units. This rock is the stratigraphically highest laminated rock (i.e., cumulate) observed in the section. Return east to Camp 7 road and continue south.

Area L: From the crest of a rise in the trail, go 30 paces west to an outcrop ledge. The rock here is a coarse-grained, decussate, subprismatic, apatitic olivine ferromonzodiorite, again typical of the **slmd** unit. Note that it appears very similar to the rock at Area H. It contains about 30% pinkish granophyre, 30% altered plagioclase, 15-20% pyroxene, 10-12% olivine, 3% apatite, 5% Fe-oxide, and 5% free quartz. This rock is thought to represent a hybrid created at the interface between the mafic melts of the SLI and the felsic melts created by melting of the Finland granophyre. Isotopic data will be needed to determine the extent of cross-contamination represented by this rock.

Area M: In outcrop pavements in the road and in a poorly exposed ledge on the northeast side, a similar rock as seen at the Area L outcrop is observed, but here it is quite variable in granophyre content (20-50%). The transition from the **slmd** unit to the prismatic quartz ferromonzodiorite phase of the Finland granophyre is taken to be where the abundance of mafic silicate and oxide phase is less than about 25%. As evident here, that transition is very irregular.

Area N: Exposed in outcrop west of a crest in the road is the contact between a fine-grained intermediate dike and coarse-grained prismatic quartz ferromonzodiorite of the Finland granophyre. The dike is thought to be part of the hybrid set related to the development of the FTMD/Beaver River diabase. The southern contact of the dike trends westnorthwest nearly parallel with the main outcrop ledge, but the northern contact is not exposed. It grades from a more felsic composition at the contact to a finer grained more mafic composition toward the interior. The dike rock is also displayed in a sloping pavement outcrop southeast of the road. The contact with the quartz ferromonzodiorite is exposed low on the south end of the exposure and is very narrowly gradational over a centimeter-wide zone.

Area O: About 40 paces northwest of the trail is a rubbly ledge of quartz ferromonzodiorite to ferromonzonite of the marginal phase of the Finland granophyre. Subprismatic mafics compose only about 15% of this rock. Bleached (kaolinitized and sericitized) plagioclase is very obvious in this rock.

Area P: In a low ledge on the northwest side of the road is a quartz ferromonzodiorite with prismatic ferrohedenbergitic pyroxene and locally fayalitic olivine. This rock and that seen at the last outcrop demonstrate the range of textures and modes of the quartz ferromonzodiorite phase of the Finland granophyre. An average composition from six samples is given in Table 3.2. It is unclear whether this rock is a primary marginal facies of the Finland granophyre intrusion or whether it represents melting, contamination, and recrystallization of the leucogranite (Area Q) by underplating of hot mafic magma forming the SLI.

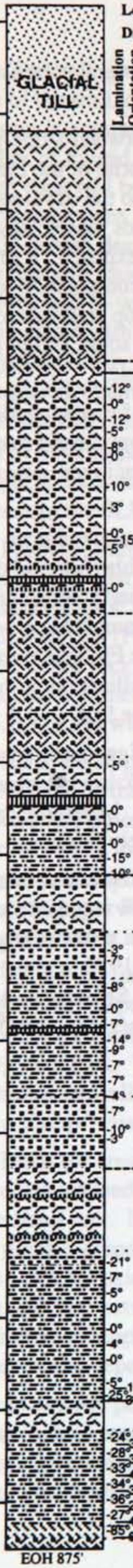
Area Q: In road pavement outcrops up the slope from the junction with Egge Lake road are good exposures of the main core phase of the Finland granophyre. The rock is a massive micrographic leucogranite (Table 3.2) with less than 5% mafic phases. Miarolitic cavities are lined with quartz, calcite, a yellow-orange Fe-silicate mineral, and rarely fluorite.

Go left (south) at the junction near the crest of the hill and continue about 130 paces to near the trail that leads to the fire tower. Take a narrow marked path to the west and follow it uphill toward the radar towers and past scattered outcrops of leucogranite. The traverse ends at near the largest of the radar towers. (The old Finland Airbase is a Cold War artifact established in 1950 to watch for intercontinental ballistic missles. It was decommissioned in the late 1970s.) Selected sections of drill core profiling the upper SLI contact (Fig. 3.19) and the lower SLI contact (Fig. 3.20) will be available for inspection.

Lithologic and Geochemical Log of SLI-1

Location: T58N, R8W, Sec 35, SE of SE, Lake County, Doyle Lake 7.5' quadrangle

Direction: 325° **Azimuth:** 65° (average) **Collar Elevation:** 1660' **Total Depth:** 875' **Core Size:** NQ



Map Units	Depth	Whole Rock	Microprobe #	Fo	En	Description
	100'					
FRPM		-120	-119		27.4	Pink, coarse, micrographic, sulfide-bearing, quartz ferromonzodiorite with prismatic to subprismatic Fe-mafic silicates
						-147' gradational
			-155		22.3	
		-188				Pink-dk green mottled, coarse, locally sulfide-bearing, micrographic ferromonzo- diorite with coarse subpoikilitic olivine and prismatic pyroxene and, locally, miarolitic cavities
SLMD	200'	-200				
		-222	-221		31.3	
						234-240' gradational
		-242	-241	14.4	40.4	
			-278	10.2	36.4	
SLAD	300'					Coarse, homogeneous, poorly to moderately laminated, granophyric (3-5%), apatitic olivine ferrodiorite
		-318	-317	10.3	40.8	
						—15' zone of granophyre diking and alteration
						grades to medium, laminated near contact
						-367' sharp
			-368	13.8	38.8	
		-382	-383	11.6	39.0	Coarse, poorly laminated to decussate, granophyric (5-10%), apatitic olivine ferrodiorite
SLMD	400'					
						—5' interval of well laminated ap-ol gabbro
			-442	10.5	40.6	444' narrowly gradational
		-457	-446	12.1	39.4	Medium to medium coarse, poorly laminated to decussate, weakly granophyric, apatitic olivine ferrodiorite
		-470				—8' prismatic melanogranophyre dike at 465-73'
SLAD			-479	18.5	46.4	-478 gradational
			-499	15.5	43.3	Medium coarse to medium, moderately to well laminated, apatitic, olivine ferrodiorite
	500'	-499	-503	32.6	51.9	
		-505	-506		43.5	512' sharp
SLMD		-518				Medium, moderately laminated grading to coarse, decussate, weakly granophyric, apatitic, subpoikilitic olivine ferrodiorite
		-538	-538	22.2	46.3	-542' gradational
		-559				Medium coarse, moderately laminated, apatitic, crs olivine ferrodiorite
						-566' gradational
		-586				Medium, moderately to well laminated, very apatitic, olivine ferrodiorite
SLAD	600'		-603	19.4	42.7	—2' granophyre dike at 592-94'
						-630' gradational
		-641				Medium coarse to coarse, poorly laminated, weakly granophyric, weakly apatitic, crs olivine ferrodiorite
			-664	17.7	46.1	-667-670' narrowly gradational
SLMD	700'	-693	-692	19.4	46.3	Coarse, locally altered, decussate, apatitic, subpoikilitic to crs granular olivine ferrodiorite
						—cut by two 1' melanogranophyre dikes at 683' & 707'
						-715' gradational
SLAD		-748	-749	14.7	38.4	Medium coarse, moderately laminated, very apatitic, olivine ferrodiorite
		-787	-787	15.2	41.1	—becomes medium, well laminated, and sulfide-bearing near contact
	800'					-795' sharp
SLMD		-802				Coarse, decussate, granophyric, sulfide-bearing, apatitic olivine ferrodiorite
						812' narrowly gradational
			-816	17.6	45.0	
SLAD		-830				Medium to medium coarse, moderately to well laminated, locally modally layered, very apatitic, olivine ferrodiorite.
			-851	18.8	47.4	
		-862				868' sharp
			-870.5	20.0	46.2	Steeply oriented, well laminated olivine ferrodiorite
						871' sharp
		-874	-873.5		43.4	Coarse, decussate, altered, very granophyric (15%) apatitic coarse olivine ferrodiorite

Figure 3.19: Lithologic and geochemical log of drill core SLI-1 through the upper part of the SLI (see Fig. 3.12 for location). From Meints and others (1993).

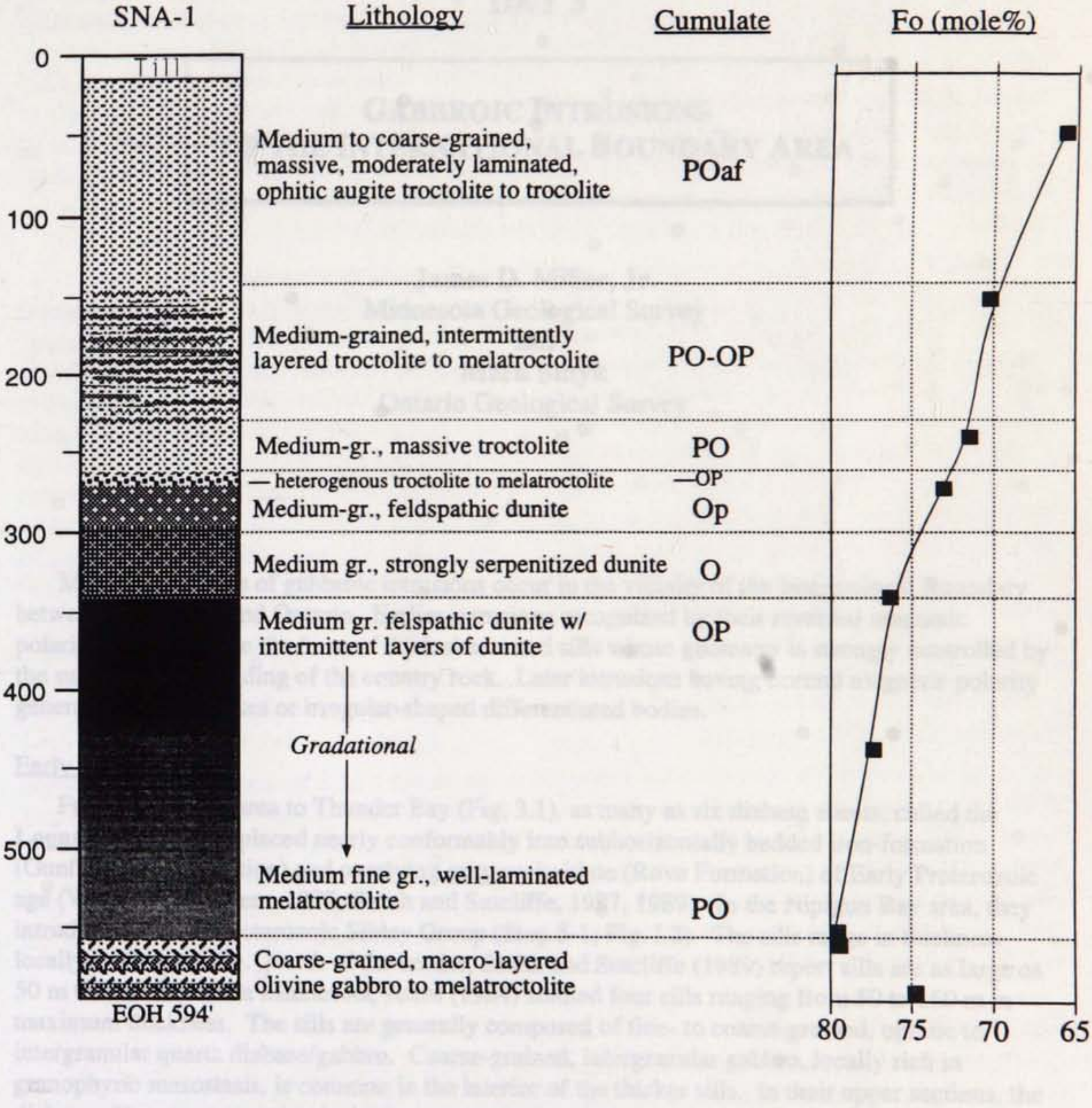


Figure 3.20: Generalized lithologic and geochemical log of drill core SNA-1 (see Fig. 3.16 for location).

FIELD TRIP 3
DAY 3

GABBROIC INTRUSIONS OF THE INTERNATIONAL BOUNDARY AREA

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Many generations of gabbroic intrusions occur in the vicinity of the International Boundary between Minnesota and Ontario. Earlier intrusions recognized by their reversed magnetic polarity generally take the form of thick sheets and sills whose geometry is strongly controlled by the subhorizontal bedding of the country rock. Later intrusions having normal magnetic polarity generally occur as dikes or irregular-shaped differentiated bodies.

Early Intrusions

From the border area to Thunder Bay (Fig. 3.1), as many as six diabase sheets, called the **Logan sills**, were emplaced nearly conformably into subhorizontally bedded iron-formation (Gunflint Iron Formation) and overlying graywacke/slate (Rove Formation) of Early Proterozoic age (Weiblen and others, 1972; Smith and Sutcliffe, 1987, 1989). In the Nipigon Bay area, they intrude the Middle Proterozoic Sibley Group (Stop 5-1; Fig. I.2). The sills range in thickness locally and regionally. North of the border, Smith and Sutcliffe (1989) report sills are as large as 50 m thick, whereas in Minnesota, Jones (1984) studied four sills ranging from 50 to 160 m in maximum thickness. The sills are generally composed of fine- to coarse-grained, ophitic to intergranular quartz diabase/gabbro. Coarse-grained, intergranular gabbro, locally rich in granophyric mesostasis, is common in the interior of the thicker sills. In their upper sections, the diabase sills are commonly plagioclase porphyritic containing as much as 60% phenocrysts that are 1-2.5 cm wide. Chilled margin and bulk compositions are iron-rich, quartz tholetiitic basalt (Table 3.3). A U-Pb zircon date of 1108(+4/-2) Ma of a Logan sill in the Lake Nipigon area of Ontario (Davis and Sutcliffe, 1985) confirm paleomagnetic data (Palmer, 1970), which imply that these sills were emplaced during the early volcanic stage of the Midcontinent rift (Fig. I.3).

Reversely polarized layered gabbro intrusions also form the northern prong of the Duluth Complex (Fig. 3.21). These intrusions, informally termed **Nathan's layered series** (Nathan, 1969; Weiblen and Morey, 1980), are composed of a variety of mafic cumulates that monoclinally dip about 15-20° to the south. Although Logan sills and Rove Formation form the footwall to these intrusions, they are capped by North Shore Volcanic Group (NSVG) lavas and contain basaltic hornfels inclusions in their lower margins (Stop 3-1). This suggests that the focus of their emplacement occurred at the discontinuity between the Early Proterozoic sediments and the base of the Keweenawan volcanic pile. A recent U-Pb zircon date of 1106.9±0.6 Ma (Paces and Miller, 1993) for a sample from the contact zone of Nathan's layered series confirms earlier interpretations of field relationships and magnetic polarity, which suggested that these reversed gabbros were roughly contemporaneous with or just younger than the Logan sills, but older than the anorthositic and layered series of the Duluth Complex (Weiblen and others, 1972). Indeed, if the ages of troctolitic and anorthositic rocks that cut out the western end of Nathan's layered series (Fig. 3.21) are similar to the 1099 Ma ages of the anorthositic and layered series rocks farther to the west and in the Duluth area (see Day 1 description), an 8-million-year hiatus between the emplacement of the reversed gabbros and the main phase of the Duluth Complex is implied.

What is known of the gabbroic rocks comes from a doctoral dissertation mapping and petrographic study by Nathan (1969) wherein he characterized the layered sequence as composed of multiple sheetlike intrusions of mafic cumulates (Fig. 3.21). Three general packages of rocks comprise the layered series: 1) a lower sequence of oxide-olivine gabbro to oxide troctolite that grades from a decussate vari-textured rock at the basal contact (Nathan's unit F) up to a well-laminated and modally layered rock (unit G) (Stop 3-1); 2) a middle sequence of troctolite (unit P; Stops 3-2 and 3-3), olivine gabbro (unit Q, Stop 3-3) and leucogabbro that is locally oxide-rich (unit J; Stop 3-4); and 3) an upper sequence of predominantly gabbronorite interlayered with troctolite intervals. The layered series is capped by monzodioritic to granitic rocks of the felsic series that in turn are roofed by lavas of the NSVG (Fig. 3.21). The petrogenetic relationship of a thinner sheet of poorly mapped gabbroic to granophyric rocks east of Nathan's layered series is unclear (Fig. 3.1). The abundance of oxide-rich gabbroic and orthopyroxene-bearing cumulate types composing the layered series is consistent with the gabbros having formed from iron-rich, quartz tholeiitic magmas similar to those parental to the Logan sills (Table 3.3). However, geochemical data are nonexistant, and a rigorous petrologic study has yet to be conducted on these interesting mafic cumulates.

Another group of reversed-polarity intrusions in the border area is a swarm of narrow diabase dikes that cut the lower reversed lavas of the North Shore Volcanic Group (Fig. 3.22). Green and others (1987) refer to these as the **Grand Portage dikes** and describe them as having an east-northeast to east trend and a range in thickness of 1 to 15 m, though some are as thick as 68 m. The diabase is typically fine grained, intersertal, and locally porphyritic. Like the Logan sill, to which they may be related, the Grand Portage dikes are quartz normative, iron rich (avg. mg# 32), and enriched in incompatible elements (Table 3.3).

<u>Younger Intrusions</u>

Closer to Lake Superior, the border area is cut by mafic dike swarms that are younger than the Logans sills as indicated by intrusive relationships and by their normal magnetic polarity and paleopole positions (Green and others, 1987). Although radiometric ages are not available for these younger intrusions, crosscutting relationships and distinctive orientations of each dike swarm indicate several episodes of emplacement among them.

The oldest and most pervasive intrusions are the **Pigeon River dikes**, which trend east-northeast to northeast and dip steeply to the southeast (Geul, 1970; Green, 1977; Smith and Sutcliffe, 1989). Displacement and warping of the Rove Formation is evident along many of the dikes. Composite intrusions are noted in several dikes (e.g., Stop 3-5). Dike widths may be as much as 150 m across in Ontario (Smith and Sutcliff, 1989) and 500 m in Minnesota (Green and others, 1987). The dikes are typically composed of ophitic diabase that may be weakly plagioclase porphyritic. Average whole rock compositions of Pigeon River dikes are moderately evolved (mg# = 52) olivine tholeiitic basalt (Table 3.3).

A 15-km-long, northwest-trending diabase dike, termed the **Arrow River dike** by Smith and Sutcliffe (1989) crosscuts Pigeon River dikes in Ontario. This dike and two shorter, similarly oriented dikes are composed of intergranular, quartz diabase that is commonly plagioclase-phyric.

TABLE 3.3: Gabbroic Intrusions of the International Boundary Area

	1	2	3	4	5
Intrusive Unit	Logan Sills		Grand Portage dikes	Mt Josephine	Pigeon River dikes
Source	A	B	C	D	C
Description	Chilled Margin	Bulk Comp	Avg. of 17 dikes	Small composite dike	Avg. of 14 dikes
SiO_2	49.0	50.1	52.5	47.9	48.9
TiO_2	3.40	3.6	2.48	3.66	1.65
Al_2O_3	13.1	13.1	13.4	11.8	16.3
FeO_t	15.6	14.3	13.3	16.2	11.5
MnO	0.22	0.16	0.19	0.24	0.18
MgO	5.6	3.9	3.68	4.04	6.46
CaO	7.45	7.2	6.72	8.9	10.21
Na_2O	2.52	3.4	3.17	2.63	2.42
K_2O	1.16	1.5	1.79	1.29	0.52
P_2O_5	0.38	0.43	0.48	0.53	0.18
Volatiles	0.92	2.5	--------	-------	-------
Total	99.35	100.19	99.11	99.6	99.8
mg#	38.4	32.2	34.1	33.0	50.1
Trace Elements (ppm)					
Cr	50	--	44	50	100
Ni	80	--	61	22	136
Rb	--	--	55	28	15.7
Sr	700	--	420	359	280
Zr	250	--	281	280	--
Hf	--	--	9.0	--	2.94

Source: A) Geul (1970), Table 3, Sample 2; B) Jones (1984), Table A-1, Sill A; C) Green and others (1987), Table II; D) Green (1986), Appendix E, Sample GP-60.

The youngest intrusions in the area tend to be more irregularlyshaped and internally zoned. One of these is the **Crystal Lake gabbro** (Stop 3-6), which is Y-shaped in plan view, with a west-northwest-striking limb 5 km long and an east-northeast-striking, southern limb 2.75 km long (Fig. 3.23). Internal layering and foliation suggest that the surface geometry of the northern limb may result from the tilting of a canoe-shaped body, open on its western end (Smith and Sutcliffe, 1987, 1989). The intrusion was subdivided into three major zones by Reeve (1969) and Geul (1970) and has been further subdivided by Smith and Sutcliffe (1987, 1989) and Cogulu (1990) into four major, roughly equivalent, lithologic zones:

(1) an upper zone (60 to 80 m) of sulfide-barren troctolite, olivine gabbro and anorthositic gabbro;

(2) a middle zone (30 to 42 m) of cyclic, layered anorthositic and olivine gabbro, Cr-spinel-bearing anorthosite, and olivine gabbro;

(3) a lower, unlayered zone (50 m) of vari-textured gabbro and leucotroctolite, which hosts the bulk of the Cu-Ni-sulfide deposit; and

(4) a basal zone (1 to 7 m) of fine-grained, chilled melagabbro and hornfelsed country-rock xenoliths.

Orthocumulate rocks predominate in the Crystal Lake intrusion; adcumulates occur only in the cyclic, layered rocks, in association with Cr-spinel mineralization (Cogulu, 1990). Plagioclase, with lesser olivine and Cr-spinel, is the main cumulus mineral, while clinopyroxene, magnetite and Cu-Ni-sulphides are intercumulus (Cogulu, 1990; Smith and Sutcliffe, 1987, 1989). Total REE, LREE/HREE ratios, and sulfide content decrease from the basal zone to the cyclic sequence, while the modal proportion of olivine increases (ibid). Smith and Sutcliffe (1987, 1989) have suggested the following crystallization sequence: Cr-spinel - olivine - plagioclase - clinopyroxene - magnetite - apatite. Biotite is epitaxial to olivine and augite, and it may also rim intercumulus sulfides and oxides (Cogulu, 1990). See Stop 3-7 for more discussion of the mineralization associated with the Crystal Lake gabbro.

Projecting east of the Crystal Lake gabbro and arcing to the northeast to hold up a string of islands in Lake Superior is a major composite dike called the **Pine River-Mount Mollie intrusion**. Smith and Sutcliffe (1989) describe this 35-km-long body as composed of gabbroic, dioritic, and granophyric rocks. Gabbroic rocks in the margins of the dike commonly have modal layering and sulfide mineralization similar to that in the Crystal Lake gabbro. Inward from the gabbro, quartz-bearing diorite cores much of the dike. In the central and western portions, granophyre occupies the core of the dike and shows contact relationships with diorite that indicates mixing of felsic and mafic liquids (Smith and Sutcliffe, 1987, 1989).

Field Stop Descriptions

STOP 3-1: BASAL MARGIN OF NATHAN'S LAYERED SERIES, DULUTH COMPLEX

Location: Roadcuts along Gunflint Trail between entrance and exit of Cook County Road 82. South Lake 7.5' quadrangle (T64N, R2W, Sec. 1).

Duration: 1 hour.

Description: Roadcuts along this section of the Gunflint Trail expose some of the rock types composing the marginal zone of Nathan's layered series (Fig. 3.21). Beginning from the junction of the Gunflint Trail and the west end of Cook County Road 82, proceed east along the north side of the Trail back toward the Rockwood Lodge.

About 150 m from the junction is a low, deeply weathered outcrop of coarse-grained, decussate, ophitic, biotitic oxide-olivine gabbro. This noncumulate, vari-textured (taxitic) gabbro forms the lowermost unit of Nathan's (1969) layered series (his unit F). A sample taken in the next outcrop west of the junction yielded a U-Pb zircon age of 1106.9±0.8 Ma (Paces and Miller, 1993). Locally, Cu-Fe sulfide is present and gives rise to the very rusty appearance of the outcrop. Sulfide mineralization in the lower contact zone is characteristic of many parts of the Duluth Complex (Bonnichsen, 1972; Hauck and others, in press). Along its northwestern margin, stable isotope data indicate that the sources of the sulfur are the shale and graywacke of the Early Proterozoic Virginia Formation, which forms the footwall of the gabbro (Ripley, 1981). Although the same slates form the footwall (as the Rove formation), the amount of mineralization here is much less. The difference may be due to a lower initial sulfur content of the Rove Formation or to depletion of sulfur by earlier intrusion of the Logan Sills.

Progressing east past intermittent outcrop for about 500 m, a prominent roadcut is predominantly composed of a very fine grained gabbro. This is Nathan's (1969) unit C, which he suggested could be recrystallized inclusions of Keweenawan lavas, Logan sills, or more likely, the chilled contact of a precursor intrusion (his units A and B). In thin section, the rock is a strongly recrystallized, poikiloblastic gabbronorite with small (<5 mm) high-density oikocrysts of augite and hypersthene enclosing granular plagioclase. Because the texture of Logan sills remains unaffected as the contact with the layered series approached (Nathan, 1969), it is unlikely that this body is such an inclusion. Rather, the zones rich in ovoid clots of coarse feldspar are similar in appearance to what have been interpreted elsewhere in the complex as recrystallized amygdules (Bonnichsen, 1972), and this inclusion is most likely a basalt flow. The abundance of these types of inclusions and the relative lack of quartz- or cordierite-bearing inclusions of Rove Formation suggest that the focus of emplacement of Nathan's layered series was at the horizontal to shallow-dipping discontinuity between Early Proterozoic sedimentary rocks and Keweenawan lava flows.

About 700 m farther east past a bend in the road, another deeply weathered roadcut exposes a medium- to coarse-grained, moderately laminated oxide troctolite with ophitic augite. This belongs to Nathan's unit G which he characterized as "coarse-grained, olivine-plagioclase and augite-plagioclase rocks with strongly foliated plagioclase and abundant tironals (Fe-Ti oxides) . . . bioite is conspicuous . . . density graded layering is prominent" (Nathan, 1969, p. 68). Alhough they form structurally lowest cumulates in the layered series, Nathan interpreted this unit to be younger than structurally higher troctolitic to gabbronoritic units on the basis of a complex sequence of discordant relationships (Units A and B, Fig. 3.21). The validity of Nathan's intrusive stratigraphy clearly needs to be tested.

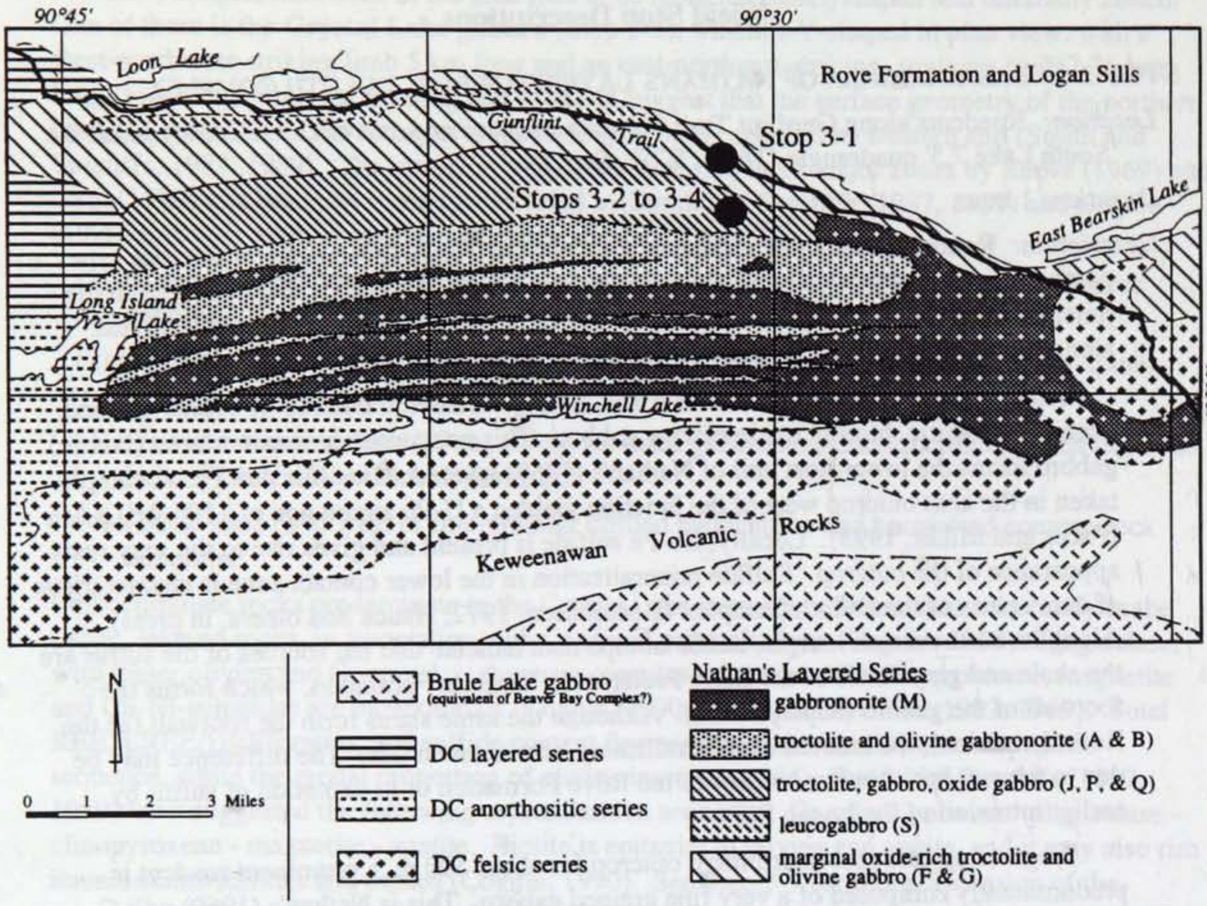


Figure 3.21: Generalized geology of Nathan's layered series showing locations of Stops 3-1 to 3-4. Letters after unit descriptions correspond to units defined by Nathan (1969). Modified from Phinney (1972b), Morey and Nathan (1977), and Green (1982a).

Stops 3-2 to 3-4 are on islands in the southwestern part of Poplar Lake and are accessible only by boat. Each site can host a maximum of about 20 people.

STOP 3-2: UNIT J OF NATHAN'S LAYERED SERIES, DULUTH COMPLEX

Location: Island in southwestern part of Poplar Lake. *Land canoes against outcrops on the southeast and soutwest sides of the island.* South Lake 7.5' quadrangle (T64N, R2W, Sec. 12, NE).

Duration: 30 minutes.

Description: Exposures on the south side of the island are composed of coarse-grained, laminated, ophitic leucogabbro (PPaf(±o) cumulate). Plagioclase ranges from 70 to 80% of the rock and is the only consistently cumulus phase. In poor exposures on the north side of the island (downsection), granular (cumulus) augite, olivine, and oxide are present, but the rock is still leucocratic. Layers and pods rich in granular Fe-Ti oxides and pyroxene are evident on the southwestern corner of the island. This rock is part of Nathan's Unit J, which he recognized as gradational downward into olivine gabbro of Unit G.

STOP 3-3: UNITS Q AND P OF NATHAN'S LAYERED SERIES, DULUTH COMPLEX

Location: Island with campsite in southwestern part of Poplar Lake. *Land canoes against outcrops on the south side or in cove on east side of the island.* South Lake 7.5' quadrangle (T64N, R2W, Sec. 12, NE).

Duration: 30 minutes.

Description: Island is composed of two east-west oriented outcrop areas. Exposures on the south side of the island are composed of medium-grained, well-laminated, intergranular olivine gabbro (PAOf cumulate). Olivine occurs as small (<1cm), subpoikilitic clots or as prisms 0.5 to 3 cm long (~1:10 aspect ratios) that tend to be concentrated in thin layers. This rock type is part of Nathan's unit P.

Outcrop on the northern half of the lake includes a variety of cumulate types interlayered with one another. The most common rock type is a medium-grained, intergranular to subophitic, augite troctolite (POAFip). This rock shows textural and modal layering to locally produce coarse-grained, intergranular oxide gabbro (PFA) and medium-grained, intergranular gabbronorite (PAIpF). These rocks belong to Nathan's unit Q which is gradational into unit P.

STOP 3-4: UNIT P OF NATHAN'S LAYERED SERIES, DULUTH COMPLEX

Location: Larger island in southwestern part of Poplar Lake. *Land canoes against outcrops on either the south-central or southeast side of the island.* South Lake 7.5' quadrangle (T64N, R1W, Sec. 7, NW).

Duration: 30 minutes.

Description: Island is composed of two outcrop areas on the south side of the island, between which access is difficult. Both exposure areas are composed of medium-grained, well-laminated, intergranular olivine gabbro (PAOf cumulate) as observed on the south side of the island at Stop 3-3. Locally, 2-5-cm clots of coarse pyroxene and oxide are present and elsewhere are pods (inclusions?) of leucogabbro. Again olivine occurs as small subpoikilitic clots or as prisms up to 3 cm long. Because prismatic olivine tends to be concentrated in thin layers along which joints tend to develop, this texture is preferentially displayed on the outcrop face. Note that on the plane of layering, prismatic olivine is commonly oriented parallel to the strike of layering. This observation and the delicate prismatic habit of the olivine argues against crystal settling by density currents.

STOP 3-5: PIGEON RIVER DIKE AT MOUNT JOSEPHINE

Location: Large roadcut on Minnesota Highway 61; 36 miles NE of Grand Marais, 3.5 miles from the International Boundary. Grand Portage 7.5' quadrangle (T64N, R6E, Sec. 34).

Duration: 20 minutes.

Description: This roadcut passes through a thick (~500 m wide) composite diabase that is nominally part of the Pigeon River intrusions (Green and others, 1987). It is part of a complex boxwork of intrusions that were emplaced into nearly flat-lying Rove Formation slates (Fig. 3.22). The extreme contrast in erodability of the diabase and the slate has given rise to very dramatic topography in the area (Mt. Josephine stands 770 feet above Lake Superior just a quarter mile (1200 m) away).

Mount Josephine has a unique north-northwest orientation compared to other northeast-trending dikes more typical of the Pigeon River intrusions. Instead, its orientation is more like the Arrow River dike observed about 12 km to the north. Unpublished mapping in the area by John Green has not discerned any intrusive relationships with the northeast-trending dikes (J.C. Green, oral comm.). Although chemical data do not exist for the main Mount Josephine diabase, it is similar in mineralogy (olivine gabbro) and texture (subophitic to ophitic) to most Pigeon River intrusions and unlike the intergranular quartz diabase of the Arrow River dike.

Exposed in this cut is evidence for two or three intrusive episodes. On the south side of the roadcut from east to west, slightly warped slates of the Rove Formation give way to a sharp chilled contact with diabase. The diabase quickly becomes medium grained and subophitic, but at about 7 m from the contact a thin aphanitic diabase dike cuts the subophitic diabase. This dike has an evolved high-Ti ferrobasalt composition (Table 3.3). About 35 m from this small dike, another intrusive contact is marked by calcite veining and columnar jointing in the fine-grained diabase west of contact. This rock also quickly coarsens to a medium-grained subophitic to ophitic olivine diabase. This rock type persists along the rest of the roadcut, though many mineralized shear zones are evident along the way.

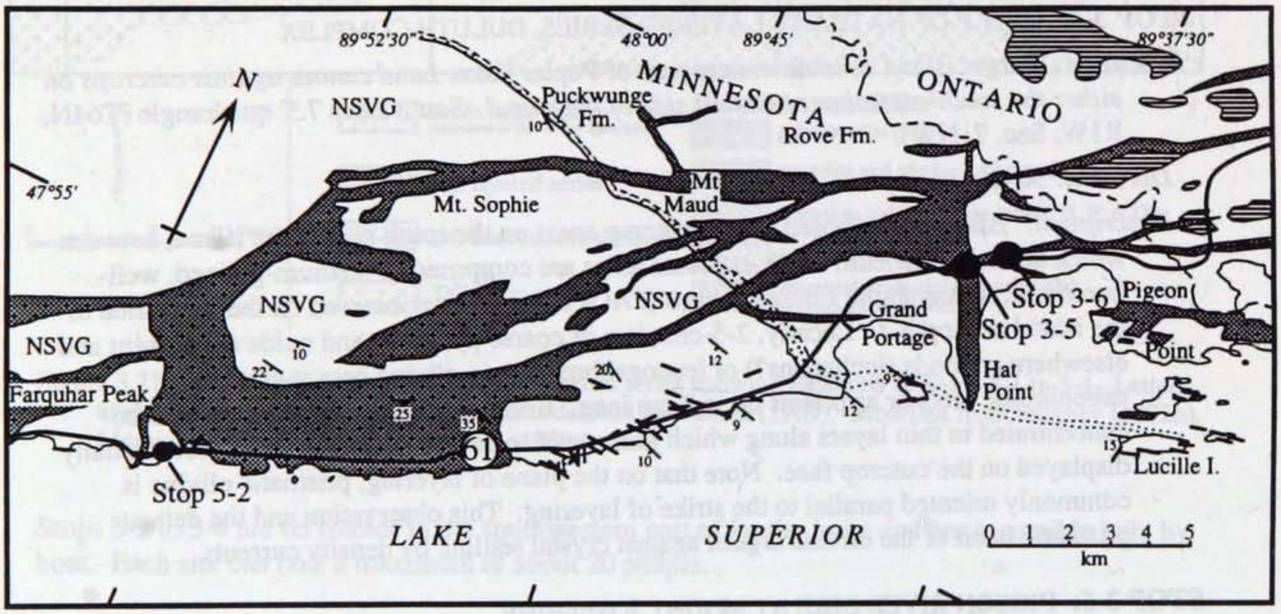


Figure 3.22: Generalized geology of the Grand Portage area showing the locations of Stops 3-5, 3-6 and 5-2. Dark shaded areas are Pigeon River diabase dikes; horizontal-ruled areas are Logan Sills; short heavy lines are Grand Portage dikes. Modified after Green and others (1987) and Smith and Sutcliffe (1987).

STOP 3-6: ROVE FORMATION AND PIGEON RIVER DIKE.

Location: Wayside pullout on Minnesota Highway 61 overlooking Waswaugonig Bay. . Grand Portage 7.5' quadrangle (T64N, R6E, Sec. 26).

Duration: 20 minutes.

Description: From this vantage, there is a view to the east of Pigeon Point, which is held up by a differentiated sill in the Rove Formation composed of olivine gabbro, leucogabbro, diorite, and granophyre. It was probably fed from the Pigeon River dike swarm. The basal Keweenawan lava flows make up the southernmost island (Lucille Island) from Pigeon Point (Fig. 3.22). The Mount Josephine dike holds up Hat Point visible to the southeast. On a clear day, Isle Royale can be seen some 35 km away.

About 15 m of section of the Rove Formation is exposed in the lower part of a 40-m-high roadcut opposite the pullout. The total thickness of the Early Proterozoic section (Gunflint Iron Formation and Rove Formation) in this area is about 1 km (Morey, 1969; Smith and Sutcliffe, 1987). Rarely is the bedding in these Early Proterozoic rocks tilted more than 20°, even here among the numerous intrusions of Pigeon River diabase dikes. The section of Rove exposed here is composed of thin-bedded argillite with intercalated beds of fine-grained graywacke dipping 15° to the southeast. Several steep, small-scale normal faults cut the sedimentary rocks that may be due to emplacement of a near-vertical, northeast-trending Pigeon River diabase that holds up the upper part of the cutface. Oxidation of the argillite is evident in the contact aureole adjacent to the diabase. The sloping cliff to the east exposes the contact, where it can be seen that the composition of the beds affects the width of the aureole. Sandy beds show the effects of heating farther away from the dike.

<u>Passing through Canadian Customs,</u> please have identification and visas available for inspection.

STOP 3-7: CU-NI-CR-PGE MINERALIZATION IN THE CRYSTAL LAKE GABBRO

Location: Great Lakes Nickel Property, 3.4 km west of Highway 61 on Forest Road 87, 47 km southwest of Thunder Bay, Ontario. South Lake quadrangle (T62N, R9E, Sec.4, NE of SE)

Duration: 2 hours

Description: A switch-back road extends from the base of a 140-m-high ridge to the lower adit, excavated at the contact between Rove Formation country rocks and the base of the Crystal Lake gabbro (see Day 3 overview for description of geology). An overgrown trail continues upward to the upper adit in "taxitic," vari-textured gabbro and talus boulders. The upper portions of the intrusion, including layered rocks, are visible from this vantage point, but are inaccesible for safety reasons. A stockpile of mineralized gabbro is available for sampling on the road at the base of the ridge. A second dump, farther up the road, consists mainly of Rove Formation shales and argillites with abundant zeolite and carbonate veins.

The sulfides which compose the Great Lakes Nickel deposit are disseminated, interstitial, and included grains and droplets. Pyrrhotite, chalcopyrite, cubanite, and pentlandite are the main sulfides. Accessory minerals include violarite, troilite, niccolite, maucherite, native bismuth, mackinawite, bornite, millerite, nickeloan pyrite, sphalerite, and marcasite (Cogulu, 1993a). Adcumulate and orthocumulate Cr-spinels have been subdivided by Cogulu (1993b) into compositionally and texturally distinct groups with complex re-equilibration histories. Heterogeneity between Cr-spinels was also noted by Whittaker (1986).

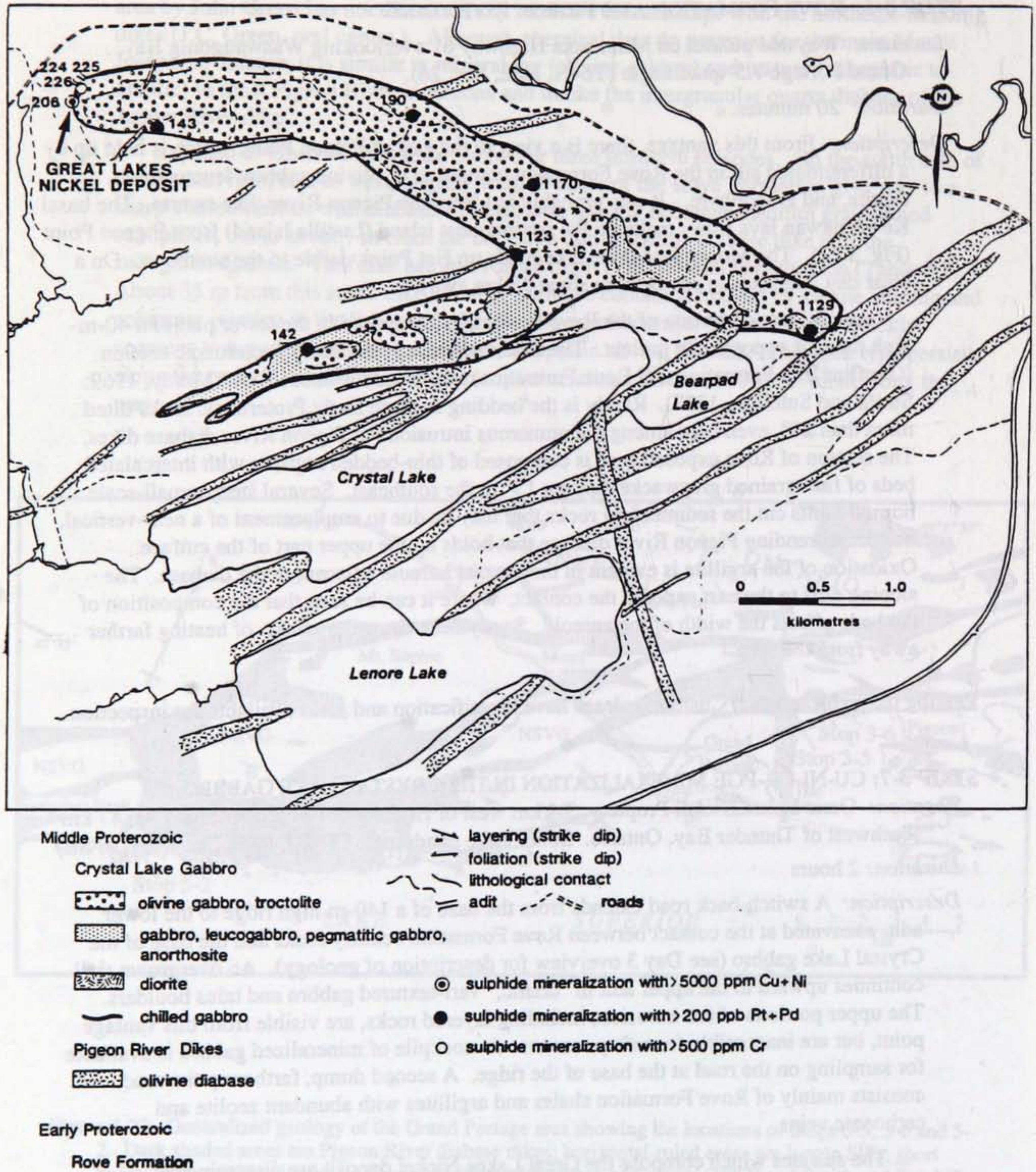


Figure 3.23: Geology of the Crystal Lake area (after Smith and Sutcliffe, 1989).

Exploration History

The discovery of copper- and nickel-mineralized float boulders in this area early in this century led to a concerted exploration effort to find their source. United States Smelting, Refining and Mining Company conducted exploration in 1936, and were followed by Frobisher Exploration Company in 1942. In the summer of 1952, J.S. Brodie and T.W. Page examined a large outcrop of gabbroic rocks, 6 km northeast of the original float discovery. Prospecting soon indicated a Cu-Ni-Pt mineralized zone, and the ground was staked for Mattawin Gold Mines. Because of lack of operating capital, the Mattawin property was optioned to Falconbridge Nickel Mines in late 1952. A trench was subsequently excavated to test the lower gabbro contact. Although trench samples returned copper and nickel values, they did not warrant further investigation and the option was allowed to lapse. Additional work by J.S. Brodie drew attention to the area north of the trench.

The property was optioned to R. Barker and W. Dawidowich in 1954. Six diamond drill holes, totalling 1058 m, were completed. One 55-foot section returned 0.54% Cu and 0.18% Ni with some precious metal values. Mogul Mining Corporation Limited held the property from 1954 to 1957 and drilled seven holes, totalling 1693 m. Intersections of the mineralized zone averaged 9 to 12 m and assayed 0.9% combined Cu and Ni.

Late in 1964, Great Lakes Nickel Corporation Limited acquired the property and initiated their exploration program in June 1965. From 1965 to 1970, 47,803 m of surface drilling was completed; 19 underground holes, totaling 392 m, were also drilled. Underground drilling was conducted from a newly constructed, 37-m-long adit, driven into the base of the hillside. In addition, Thunder Bay Nickel Mining Corporation Limited drilled 16 holes, totaling 13,579 m, on the down-plunge extension of the deposit, 2.5 km east of the main workings.

In 1972, access and development work was undertaken by Great Lakes Nickel to further test the deposit. This work included the excavation and driving of a 522-m development portal and drift and more than 12 000 m of surface and underground drilling. Plant-site surveys, bulk sampling, metallurgical, and feasibility tests were also conducted, financed largely by the Swedish company, Boliden Aktiebolag. By 1974, plans were made to mine the deposit at an initial rate of 1.8 million tons per year (subsequently increased to 2.5 million tons per year). Up to that point, about $10 million (Canadian) had been spent on the property on 58,689 m of surface drilling, 26,182 m of underground drilling and the driving of the adit, which eventually reached a length of 1041 m. This work had outlined a deposit of 32.8 million tons grading 0.36% Cu and 0.20% Ni, with a further potential reserve of 40 million tons of about the same grade.

However, cost escalations, high interest rates and uncertain metal prices forced suspension of mine development in October 1974. Rising interest in platinum-group elements in the mid-1980s prompted Fleck Resources Ltd. to re-evaluate the deposit, whose reserves then stood at 45.6 million tons at a grade of 0.334% Cu and 0.183% Ni. Between September 1986 and February 1987, Fleck completed geologic mapping and sampling, as well as the relogging and assaying of more than 9144 m of drill core. Six holes were drilled to test the deposit for its PGE potential. Sampling by Fleck returned the following assays on a 3.7-million-ton portion of the deposit:

> 0.006 oz./ton Pt, 0.57% Cu, 0.027 oz./ton Pd, 0.264% Ni, 0.003oz./ton Au, 0.016% Co, 0.04 oz./ton Ag. Based on February 1987 metal prices, this material was valued at $34.29 (Can.) per ton (Fleck Resources Ltd., Annual Report, 1987).

<u>FIELD TRIP 3</u>
DAY 4

GEOLOGY AND MINERALIZATION OF INTRUSIVE COMPLEXES OF THE MARATHON, ONTARIO AREA

Mark Smyk and Ronald P. Sage
Ontario Geological Survey

Overview

A variety of alkalic and carbonatitic rocks compose several intrusive complexes on the North Shore of Lake Superior. They include the Coldwell and Killala Lake alkalic complexes, the Prairie Lake carbonatite, and numerous diatremes and related dykes in the vicinity of Dead Horse Creek (Fig. 3.24). This part of the field excursion will examine geology and mineralization of the Dead Horse diatreme (Stop 4-1) and the Coldwell Complex (Stops 4-2 to 4-9).

These complexes are spatially localized and structurally controlled by the Trans-Superior Tectonic Zone (TSTZ), a north-northeast-trending structure that extends for over 600 km and includes the Thiel Fault in Lake Superior (Klasner and others, 1982). Carbonatitic magmatism has been recognized at Chipman Lake (Sage, 1985), 150 km northeast of Lake Superior. Magmatism related to the Midcontinent rift occurred along the TSTZ from approximately 1.2 to 1.0 Ga (Table 3.4). It has been postulated that the TSTZ may represent part of a failed arm of a Keweenawan triple junction (Weiblen, 1982; Mitchell and Platt, 1982b) or the intersection of a late fracture system with the rift (Mitchell and others, 1983). Local alkalic and carbonatite complexes have been emplaced at imflections in the trends of major structural zones, or at sites of cross-faulting (Sage, 1991). The Coldwell and Killala Lake complexes have both formed as the result of ring fracturing and caldera collapse. The abundance of xenolithic blocks and roof pendants suggests that these complexes are exposed at relatively high structural levels.

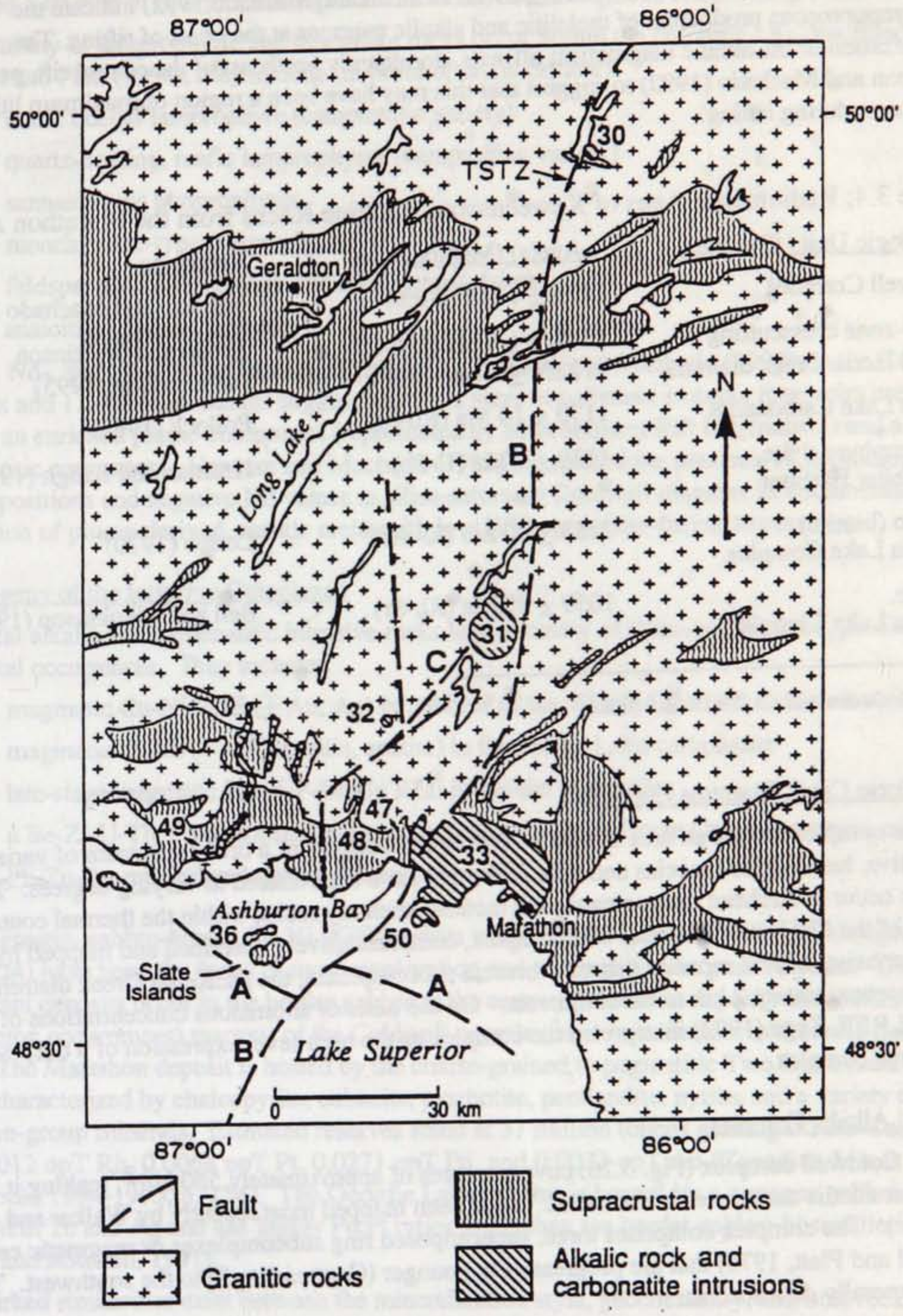


Figure 3.24: Schematic geologic map of the Marathon area showing the locations of alkalic rock and carbonatite intrusions along the northern part of the Trans-Superior Tectonic Zone (TSTZ). The major intrusions and structures in the area are: 31- Killalla Lake Alkalic Complex; 32- Prairie Lake Carbonatite Complex; 33- Port Coldwell Alkalic Complex; 36- Slate Islands diatremes and carbonatite dike; 47- Dead Horse Creek diatreme; 48- McKellar Creek diatreme; 49- Gold Range diatreme; 50- Neys diatreme; A- Michipicoten Island Fault; B- Theil Fault and its extrapolated northern extension; C- Killala Lake deformation zone.

Identical ages for Logan diabase sills (1108.8 +4/-2 Ma; Davis and Sutcliffe, 1985) and alkalic rocks of the Coldwell Complex (1108 ± 1 Ma; Heaman and Machado,1992) indicate the contemporaneous production of tholeiitic and alkalic magmas at the onset of rifting. The locallization of the alkalic magmatism off-axis, dominantly northeast of the central rift, prompted Heaman and Machado (1992) to suggest that this may have been a region of maximum lithospheric extension during rifting.

Table 3.4: Radiometric Ages of Keweenawan Alkaline Rocks from the Marathon Area.

Lithologic Unit / Complex	Age(s) (Method)	Reference
Coldwell Complex	1108 ± 1 Ma (U-Pb)	Heaman and Machado (1987)
Be-Zr-zone crosscutting Dead Horse Creek diatreme	1112.7 ± 4 Ma (Upb) [1] 1128.7 ± 6 Ma (U-Pb) [2]	Krogh and Wilkinson (pers. comm. 1995)
Prairie Lake Carbonatite	1130 ± 10 Ma (Rb-Sr)	Pollock (1987)
Lamprophyre dyke, McKellar Harbour	1130 ± ? Ma (U-Pb)	Heaman and Krogh (1986)
Gabbro (biotite), Killala Lake Complex	1185 ± 90 Ma (K- Ar)	Coates (1970)
Syenite, Killala Lake Complex	1050 ± 35 Ma (Rb-Sr)	Bell and Blenkinsop (1980)

1 -2.49% discordant; 2 -1.82% discordant

Dead Horse Creek Diatreme Complex

This complex covers an area of 400 x 1600 m and consists of a broad spectrum of variably radioactive, heterolithic breccias and dike rocks that have been altered to varying degrees. The breccias occur in Archean metavolcanic and metasedimentary rocks, within the thermal contact aureole of the Coldwell complex. The complex, comprehensively described and mapped by Sage (1982), consists of five separate diatreme breccia subcomplexes; the McKellar Creek diatreme occurs approximately 2 km to the southwest. On the basis of anomalous concentrations of K, Nb, Zr, and LREE, Sage (1982) interpreted the complex as the high-level expression of a deep-seated carbonatite complex.

Coldwell Alkalic Complex

The Coldwell complex (Fig. 3.26) covers an area of approximately 580 km^2, making it one of the largest alkalic complexes in the world. It has been mapped most recently by Walker and others (1993 b,c). The complex comprises three, superimposed ring subcomplexes or magmatic centres (Mitchell and Platt, 1978) that are progressively younger (Centres 1 to 3) to the southwest. They can be generally described as:

Centre 1: saturated alkaline rocks with oversaturated residue, chiefly gabbro and Fe-rich augite syenite;

Centre 2: miaskitic alkaline rocks with oversaturated residue, chiefly alkalic biotite gabbro and nepheline syenite; and

Centre 3: alkaline rocks with oversaturated residue, chiefly syenite and quartz syenite.

The superimposition of intrusive centres and a complex and protracted magmatic history have produced a myriad of hybrid rocks, igneous breccias, and ambiguous crosscutting relationships.

A variety of lamprophyric and other dike rocks occur within the complex. As described by Mitchell and Platt (1994), they include (in order of emplacement):

(1) mafic ocellar lamprophyre (camptonitic variety)

(2) quartz-bearing, mafic lamprophyres (camptonitic variety)

(3) sannaite-type lamprophyres

(4) monchiquitic-type lamprophyres

(5) feldspar glomeroporphyry and alkali basalt dykes

(6) analcime tinguaite (heronite)

Sr-, Nd-, and Pb-isotopic studies conducted by Heaman and Machado (1992) on the Coldwell complex and 1.1-Ga carbonatites suggest that there were two distinct isotopic reservoirs present at 1.1 Ga: an enriched plume-component (represented by Midcontinent Rift magmatism) and a carbonatitic component. Heaman and Machado (1992) interpreted the presence of nonradiogenic Pb compositions and negative Nd values in silica-saturated Coldwell magmas as documenting the interaction of plume-derived, mantle melts with low U/Pb, granulite-facies, lower crust.

Metallogeny of the Intrusive Complexes

Local alkalic and carbonatitic intrusive rocks host a variety of characteristic base, precious, and rare metal occurrences. They include:

(1) magmatic Cu-Ni-PGE ($\pm$ Au, Ag) in gabbros of the Killala Lake and Coldwell complexes

(2) magmatic U, Nb (+ wollastonite, apatite) in the Prairie Lake carbonatite

(3) late-stage magmatic Nb-Y-F-family REE in syenite pegmatites

(4) a Be-Zr-U-Th-Y mineralized zone crosscutting the Dead Horse Creek diatremes, and

(5) Pb-Zn-Ag mineralized quartz-carbonate veins.

Magmatic, gabbro-hosted Cu-Ni-PGE deposits in the Killala Lake and Coldwell complexes (Fig. 3.24) have been the focus of much exploration and research for the past 30 years. The most significant deposits occur in the border gabbro at the eastern (Marathon deposit) and western (Middleton occurrences) margins of the Coldwell complex, and in its interior at Geordie Lake (Fig. 3.26). The Marathon deposit is hosted by the coarse-grained to pegmatitic Two Duck Lake gabbro and is characterized by chalcopyrite, cubanite, pyrrhotite, pentlandite, pyrite, and a variety of platinum-group minerals. Estmated reserves stand at 37 million tonnes grading 0.31% Cu, 0.04% Ni, 0.0012 opT Rh, 0.0068 opT Pt, 0.0271 opT Pd, and 0.0023 opT Au (Canadian Mines Handbook, 1994-1995, p. 154). The Geordie Lake deposit is hosted by a younger gabbro, is enriched in Te and Ag and has higher Pd:Pt ratios (~19) than the border gabbro-hosted deposits (Mulja and Mitchell, 1991).

Marked similarities exist between the mineralization style, geochemistry, and host rocks of both Coldwell- and Duluth Complex-hosted deposits, as well as the Crystal Lake gabbro (Stop 3-7). These features include mineral textures, abundance, and compositions; crystallization paths for the host gabbros; silicate-sulfide associations; trace-element trends; and chalcophile-element fractionation trends (Good and Crockett, 1994a).

Recent research by Watkinson and Ohnenstetter (1992) and Good and Crockett (1994a,b) has produced debate about the relative importance of magmatic and hydrothermal processes in the Cu-Ni-PGE mineralization process. Watkinson and Ohnenstetter (1992) have presented field, petrographic, and mineral-chemical data that support the interaction of magmatic sulfide mineral assemblages with a Cl-rich mixture of magmatic (deuteric) fluid and volatile species generated by the breakdown of assimilated xenoliths at low temperatures. However, Good and Crockett (1994a,b) have contended that element migration took place over only very short distances and that the original bulk sulfides were not enriched in Cu and PGE by later fluids.

Sage (1987) has summarized the occurrence and exploration for uranium, niobium, wollastonite and apatite in the Prairie Lake carbonatite complex (Fig. 3.24). Uraniferous and nonuraniferous pyrochlore occurs as inclusions in pyroxene and in calcite-apatite segregations in ijolite. In sovite, it occurs in bands with other oxides and silicates and in association with clusters of magnetite-apatite (Watkinson, 1976).

REE mineralization in the Coldwell Complex has most recently been summarized by Walker and others (1992; 1993a) and McLaughlin (1990). The most common REE-bearing phases are bastnaesite and synchisite, in association with thorite, pyrochlore, feldspar, quartz, carbonate, and fluorite in pegmatitic syenite dikes and pods. The majority of the REE occurrences are in pegmatites associated with the intrusion of amphibole syenite between the Fe-rich augite syenite and a basaltic xenolith roof pendant. It has been suggested that the pegmatites resulted from the accumulation of ascending REE-enriched, residual fluids within cupolas at the base of the roof pendant.

Rare-metal mineralization associated with the Dead Horse Creek diatreme complex has been documented by Smyk and others (1993) and Knox (1987). A Be-Zr-Y-U-Th-mineralized zone crosscuts diatreme breccia of theWest Dead Horse Creek subcomplex. Rare-metal associations and HREE enrichment suggest that this mineralization is perhaps related to syenitic magmatism associated with the nearby Coldwell complex, rather than with the host diatremes, which appear to have a carbonatitic affinity.

A number of Pb-Zn-Ag-bearing, carbonate $\pm$ quartz veins occur in the vicinity of Dead Horse Creek (Walker, 1967; Kissin and McQuaig, 1988; McQuaig and Kissin, 1995). New occurrences have been recently noted by Walker and others (1992) in alkaline gabbro of the Coldwell complex and within the contact metamorphic aureole of the complex near Middleton (G. and D. Michano, prospectors, oral comm., 1995). These veins contain galena and sphalerite, with minor chalcopyrite, pyrite, freibergite ($Cu_6(Ag,Fe)_6Sb_4S_{13}$), digenite and covellite in a matrix of calcite $\pm$ quartz (Kissin and McQuaig, 1988; McQuaig and Kissin, 1995). Although no definitive age dating has been conducted on these veins, crosscutting relationships and similar metal associations with veins near Thunder Bay suggest that they are related to Midcontinent rifting.

Field Stop Descriptions

STOP 4-1: DEAD HORSE CREEK DIATREME COMPLEX

Location: Lat: 48°50' 30" N Long: 86° 40' 15" W; 2.9 km north on Dead Horse Road (east side) from junction with Highway 17.

Duration: 30 minutes.

Description: This site, east of the main access road, is the 1976 discovery outcrop of diatreme breccia. The breccia is gray-weathering, clast-supported, and produces a pitted weathered surface. The fragments are angular to rounded and generally less than 0.3 m in maximum dimension. Clasts are extensively altered along their margins. Fibrous white crystals of scapolite up to 5 cm long, in addition to carbonate and amphibole (riebeckite?), may replace parts of some clasts. The matrix to the fragments consists of quartz, carbonate, amphibole, opaque minerals and scapolite.

Another site, in the west Dead Horse subcomplex, comprises an area stipped in 1987 by Unocal Canada Limited during their exploration program for yttrium (Fig. 3.25). Exploration focused on a narrow, siliceous, Be-Zr-U-Th-Ymineralized zone that crosscuts heterolithic diatreme breccia. This zone, most recently described by Smyk and others (1993),

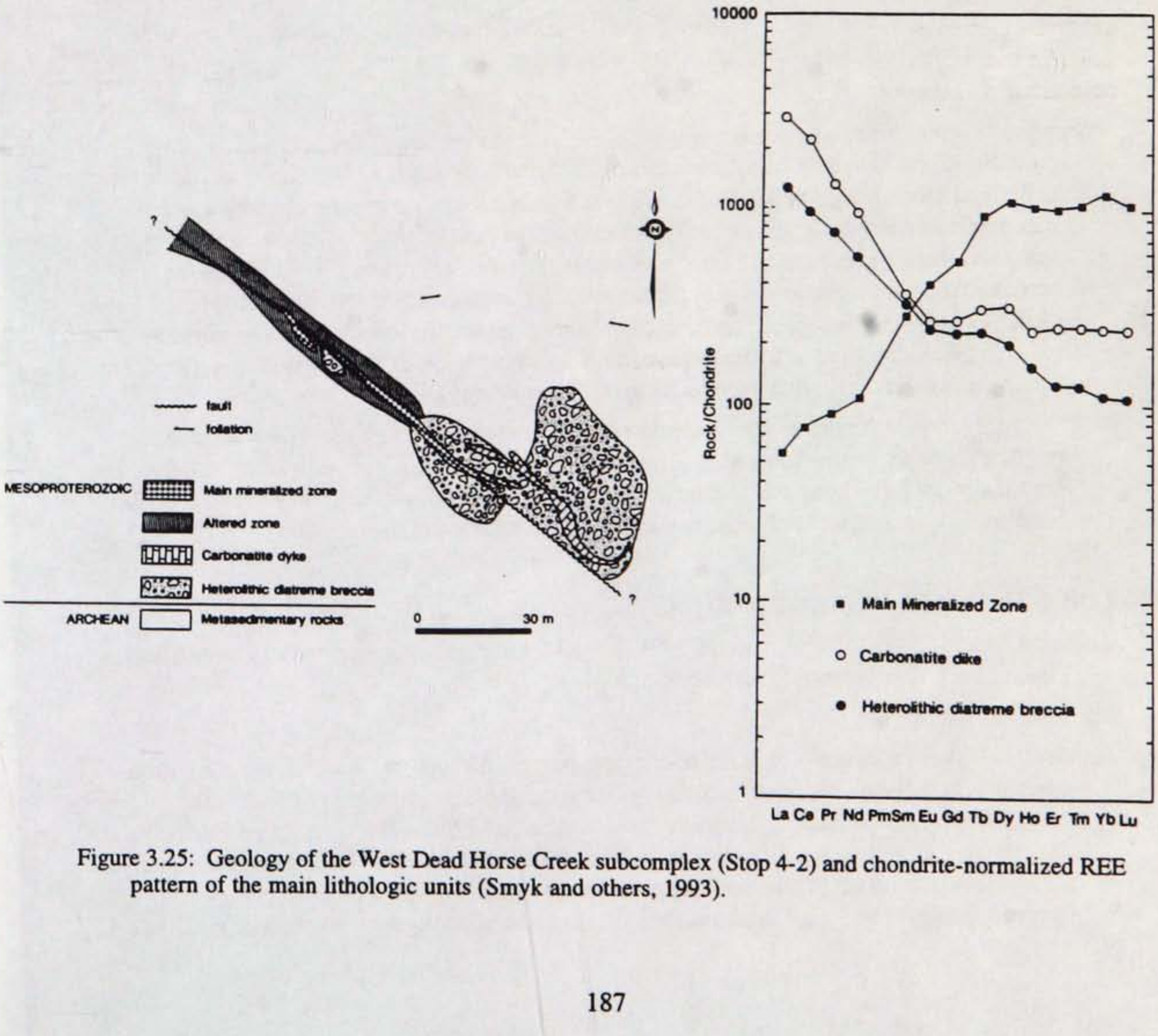


Figure 3.25: Geology of the West Dead Horse Creek subcomplex (Stop 4-2) and chondrite-normalized REE pattern of the main lithologic units (Smyk and others, 1993).

consists of highly radioactive, fine-grained, chocolate-brown, vitreous material. Most of this material is a fine-grained, metamict, calcium zirconosilicate, with accessory zircon, uraninite, thorite, monazite (Ce) and xenotime (Y). Robust pink phenakite, the beryllium-bearing phase, is visible in quartz veins that locally core the zone. Assays, returned over a width of 1.5 m and a length of 82 m, range up to 11.6% Zr, 0.6% Be, 2.5% Th, 250 ppm Sc, 1850 ppm Y, 300 ppm Nb, 1004 ppm ΣREE and 4600 ppm U. REE patterns and geochemistry suggest that the diatreme breccias were related to carbonatitic magmatism, whereas the later, rare-metal mineralization may be associated with an agpaitic type of magmatism, more akin to the Coldwell Complex.

The host heterolithic diatreme breccia at this site and elsewhere in the Dead Horse and McKellar Creek diatreme complexes, is notable in that it contains subrounded to subangular clasts of white to pink quartzite that has been ascribed to the Middle Proterozoic Sibley Group, the nearest outcrops of which are 60 km west of these sites. The majority of the clasts are of local derivation, consisting of variably hematitized and fenitized Archean supracrustal rocks. The radioactivity of the samples is apparently proportional to the degree of their hematitization. Riebeckite is a common accessory phase in the fine-grained, comminuted matrix of the breccia. Carbonate-rich lamprophyre and mafic silicocarbonatite dikes which intrude the breccia will be the focus of future geochronologic study.

STOP 4-2: CONTACT METAMORPHIC AUREOLE, COLDWELL ALKALIC COMPLEX

Location: Lat: 48°48' 40" N Long: 86°39' 45" W; 2.0 km east along Highway 17 from the Dead Horse Creek Road turn-off, both sides of highway

Duration: 15 minutes

Description: Theseroad cuts consist of hornfelsed Archean wackes of the Schreiber-Hemlo greenstone belt which strike approximately 050° and dip steeply to the south. Extreme variation in the attitude of bedding is evident, however. The contact metamorphic aureole was first described by Walker (1967), who noted the development of red-brown biotite, inclusion-free(r) feldspars, and locally garnet in the metasedimentary rocks within the aureole. The mafic volcanic rocks display hornblende (rather than tremolite-actinolite), biotite closer to the complex contact, and pyroxene at the contact. Acicular cummingtonite has recently been identified in mafic metavolcanic rocks in the Dead Horse Creek area (Resident Geologist's Files, Schreiber-Hemlo District, Thunder Bay).

Molybdenite occurs as disseminated grains and rosettes up to 2 mm along joint and fracture surfaces. Some molybdenite has been altered to yellow ferrimolybdite. The wackes are locally cut by gabbro and syenite dikes of the Coldwell Complex. These syenite dikes commonly host xenoliths of country rocks and other phases of the complex.

STOP 4-3: NATROLITE-BEARING SYENITE, COLDWELL ALKALIC COMPLEX

Location: Lat: 48°47' 45" N Long: 86°39' 15" W; 3.3 km east along Highway 17 from Dead Horse Creek Road turn-off, north side of highway

Duration: 15 minutes

Description: This exposure is of natrolite-bearing, pegmatitic syenite. Reddish-orange natrolite patches ($\leq$10-15 cm), perthitic feldspar ($\leq$30 cm), and black amphibole ($\leq$20-25 cm) comprise the bulk of this pegmatitic syenite; Mitchell and Platt (1994) report accessory pleochroic clinopyroxene, zircon, titanite, and biotite. Natrolite has locally been ascribed to the hydrothermal alteration of primary nepheline; it has also been referred to as "hydronepheline" by many local workers. The syenite is intruded by a lamprophyre dike (camptonite?), Mitchell and Platt, 1994) and also hosts medium-grained gabbro xenoliths. These gabbro xenoliths, up to 1 m in size, have a dark reaction rim 1 to 2 cm wide next to the enclosing syenite.

To the east, the pegmatitic syenite gives way to finer grained nepheline syenite in which chalky-weathering nepheline may be discerned. Rare natrolite grains are also present. Farther east, a variety of equigranular and pegmatitic syenites are exposed. Near the eastern end of the outcrop, a large xenolith of gabbro-hosted, massive, titaniferous magnetite has been exposed. Minor clinopyroxene, plagioclase, and apatite occur within the massive oxide unit. Analyses completed in 1951 and reported by Hinz and Landry (1994) indicated total Fe and Ti values ranging between 33% and 45%, and 4.5% and 13.5%, respectively; P contents ranged up to 0.371%. McGill (1980) has examined both the natrolite-bearing syenites and the titaniferous magnetite deposits in the gabbro.

Mitchell and Platt (1994) have most recently assigned these various rock units to their respective centres within the Coldwell Complex. Their associations are:

Gabbro, Fe-rich augite syenite — Centre 1 (oldest)

Natrolite-bearing syenite — Centre 2

Amphibole-bearing syenite — Centre 3 (youngest)

STOP 4-4: LITTLE PIC RIVER BRECCIA ZONE, COLDWELL ALKALIC COMPLEX

Location: Lat: 48°47'45" N Long: 86°37' 30" W; 5.5 km east along Highway 17 from Dead Horse Creek Road turn-off; first highway roadcuts east of Little Pic River bridge.

Duration: 30 minutes

Description: These roadcuts expose spectacular intrusive breccias within the youngest rocks of the complex, along the east side of the fault zone that the Little Pic River occupies. The breccias consist of angular blocks of fine- to medium-grained, equigranular, gabbroic rocks in a matrix of pink, medium-grained, quartz syenite. The mafic rocks were interpreted as oligoclase-bearing basalt by Mitchell and Platt (1982a). Subsequent discussion and study have led to the suggestion of perhaps two, texturally discernable, types of basic xenoliths: those with (i) sharp, angular margins, and (ii) those with lobate to cuspate margins. In this model, the angular xenoliths represent synplutonic basalts, which are now preserved elsewhere as megaxenoliths in younger intrusions (e.g., STOP 4-6). The cuspate-margined xenoliths may represent the effects of mixing between two contemporaneous gabbroic/basaltic and syenite magmas (i.e., magma mixing/co-mingling). Cuspate, possibly chilled margins with quench-textured clinopyroxene, plagioclase, and skeletal olivine have been noted in similar xenoliths to the south on the Coldwell Peninsula by G. Shore (oral comm. 1995); they suggest the quenching of the basic magma against the cooler, syenitic magma.

Although isolated xenoliths are common, there are areas of incipient or in-situ brecciation characterized by syenite dikes and "jig-saw puzzle" breccias, where brecciated fragments can be fit back together. Miarolitic cavities, up to several centimetres in width, contain euhedral quartz, feldspar and calcite crystals.

The breccia zone persists to the east, toward the scenic lookout (800 m east). The south side of the highway is underlain by oligoclase gabbro and quartz syenite, and various, xenolithic-bearing syenitic rocks are exposed on the north side. These pyroxene- and amphibole (ferro-edenite)-bearing syenites contain xenoliths of alkali gabbro, alkali diorite, and other equigranular to porphyritic syenites. Near the lookout turnoff, gray, nepheline-bearing syenite intrudes the mafic rocks and contains orange natrolite. Sannaite and ocellar, camptonitic lamprophyre dikes have been reported near this site by Mitchell and Platt (1994), who proposed the following order of local emplacement:

Mg-hornblende syenite—contaminated Fe-edenite syenite—Fe-edenite syenite—qtz syenite
(earliest) (latest)

Lukosius-Sanders (1988) classified the local rocks as miaskitic, metaluminous syenites, enriched in U, Th, REE, and Zr. These syenites have affinities to A-type granites and have been interpreted to be the result of fractional crystallization of mantle-derived, basaltic magma (Lukosius-Sanders, 1988; Mitchell and others, 1993).

STOP 4-5: CENTRE 2 GABBRO AND SYENITE, COLDWELL ALKALIC COMPLEX

Location: Lat: 48°46' 30" N Long: 86°37' 10" W; Prisoner Cove, Neys Provincial Park, 9.8 km east of the Dead Horse Road turn-off, 2.8 km off Highway 17 to the park headquarters and south along the shoreline trail. (**Sample Collecting Prohibited!**)

Duration: 1 hour

Description: The wave-washed, glacially polished outcrops along the shoreline of Lake Superior at Prisoner Cove and south along the western side of the Coldwell Peninsula exhibit a variety of lithologic, textural, and crosscutting features that characterize much of the Centre 2 magmatism in the Coldwell complex. In its simplest sense, this stop exposes the contact between alkaline gabbro and amphibole-natrolite-nepheline syenite, but the enigmatic effects of assimilation and hybridization have severely complicated and obscured many of the primary features.

Medium- to coarse-grained, olivine- and enclave-bearing, biotite gabbro composes much of the eastern portion of the outcrops. Gabbro xenoliths occur within the syenite and within hybrid phases along their mutual contact, which trends roughly north, parallel to the shoreline. Much of the outcrop has a pitted surface resulting from the preferential weathering of mafic enclaves consisting of biotite-olivine gabbro to biotite-clinopyroxene gabbro or leucogabbro (Walker and others, 1992) within a more syenitic matrix. The syenitic matrix consists of fine- to coarse-grained nepheline (+natrolite) syenite with acicular amphibole and poikilitic biotite. Mitchell and Platt (1994) have identified the amphibole as hastingsite.

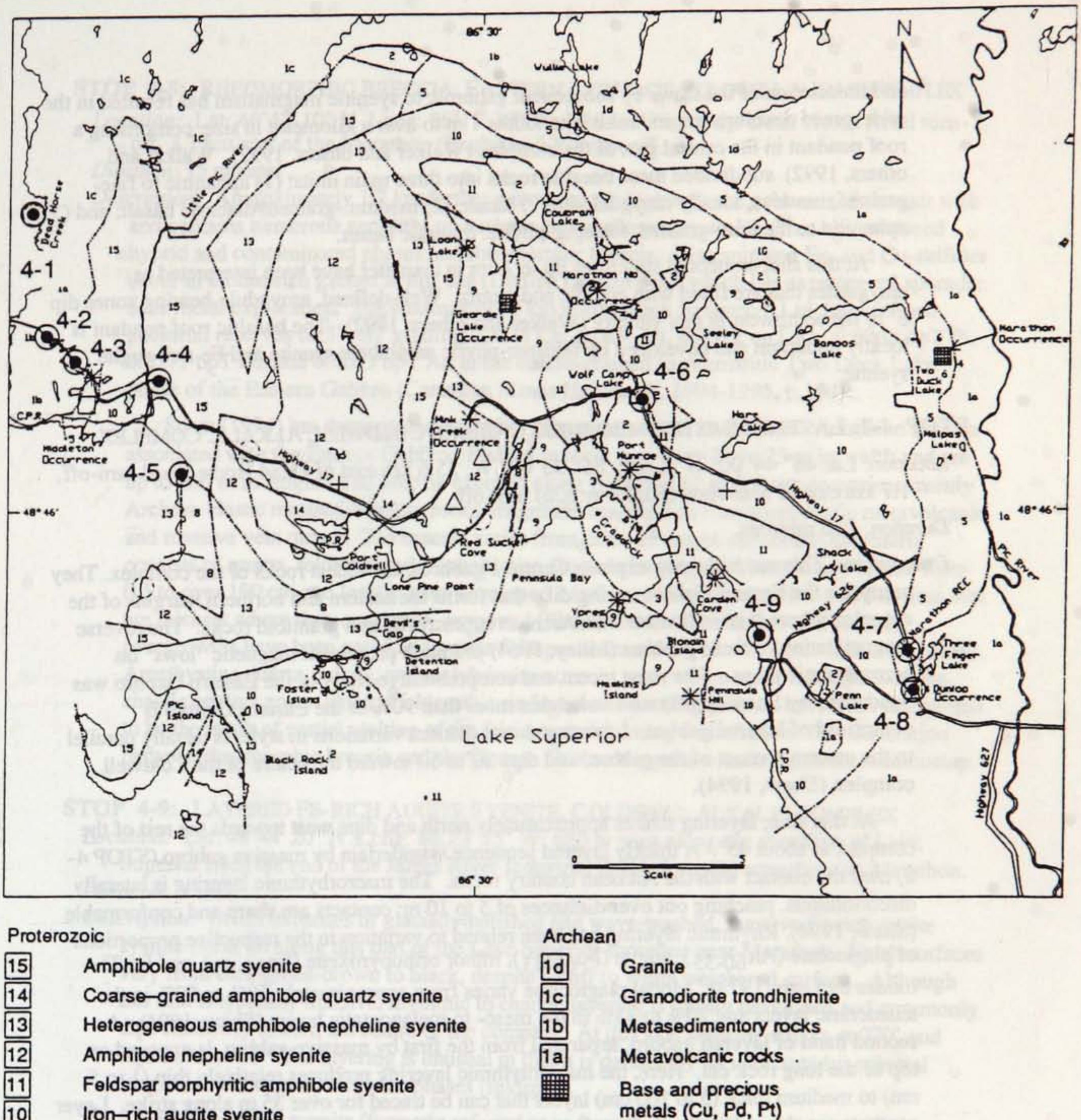
Distinct to nebulitic layering exists within the amphibole-natrolite-nepheline syenite. It is oriented parallel to the contact and dips very steeply to vertically at this location. Much discussion has focused on whether the observed structures have resulted simply from igneous process, syn- or post-intrusion shearing, or a combination of processes.

STOP 4-6: HORNFELSED BASALTIC ROOF PENDANTS

Location: Lat: 48°47' 40" N Long: 86°26' 05" W; 24.5 km east of Dead Horse Road turn-off, north side of highway (pull-off), road to Wolf Camp Lake.

Duration: 15 minutes.

Description: Hornfelsed basaltic rocks overlying the complex were recognized early in its mapping by Tuominen (1967) and Puskas (1970) and likely represent a volcanic edifice that has been subsequently eroded (Sage 1986). Mitchell and Platt (1994) and Nicol (1990) have considered these basalts to be a tholeiitic lineage, contemporaneous with the Coldwell complex. Fresh, metasomatized, and hornfelsed andesine-oligoclase basalt flows are estimated to attain a thickness of 5 m (Mitchell and Platt, 1994; Nicol, 1990). Assimilation



Proterozoic

- 15 Amphibole quartz syenite
- 14 Coarse–grained amphibole quartz syenite
- 13 Heterogeneous amphibole nepheline syenite
- 12 Amphibole nepheline syenite
- 11 Feldspar porphyritic amphibole syenite
- 10 Iron–rich augite syenite
- 9 Recrystallized amphibole quartz syenite
- 8 Alkaline gabbro
- 7 Geordie Lake Gabbro
- 6 Two Duck Lake Gabbro
- 5 Eastern and Western Gabbro
- 4 Rheomorphic breccia
- 3 Monzodiorite
- 2 Basaltic xenolith

Archean

- 1d Granite
- 1c Granodiorite trondhjemite
- 1b Metasedimentory rocks
- 1a Metavolcanic rocks
- Base and precious metals (Cu, Pd, Pt)
- Base metals (Cu)
- Rare metals (REE, Nb, Y, Zr)
- Industrial minerals (nepheline)
- building stone
- Geological contacts
- Diatremes

Figure 3.26: General geology of the Coldwell alkalic complex showing field trip stop locations. Geology from Walker and others (1992).

and brecciation of the flows by subsequent gabbroic to syenitic magmatism has resulted in the widespread development of basaltic xenoliths 1 m to over a kilometre in size, comprising a roof pendant in the central part of the complex (Walker and others, 1992). Walker and others, 1992) subdivided these basaltic rocks into three main units: (1) aphanitic to fine-grained, massive, locally amygdaloidal(?) basalt; (2) medium-grained, diabasic basalt; and (3) aphanitic to medium-grained, feldspar-phyric, diabasic basalt.

At this site, amoeboid structures up to 2 cm in diameter have been interpreted as amygdules that are filled with epidote and quartz. Well-defined, amygdule-bearing zones dip 8° to the southwest in this vicinity (Walker and others, 1992). The basaltic roof pendant is locally underlain and enveloped by feldspar-phyric amphibole syenite and Fe-rich augite syenite.

STOP 4-7: LAYERED EASTERN (BORDER) GABBRO, COLDWELL ALKALIC COMPLEX

Location: Lat: 48° 44' 00" N Long: 86° 20' 00" W; 35.6 km east of Dead Horse Road turn-off, 1.7 km east of Marathon (Highway 626) turn-off

Duration: 30 minutes.

Description: This roadside stop exposes Centre 1 gabbros, the oldest rocks of the complex. They comprise the Eastern Gabbro, a ring dike that forms the eastern and northern margins of the complex where it is in contact with Archean supracrustal and granitoid rocks. The reverse magnetization of these gabbros (Lilley, 1964) produces prominent magnetic "lows" on aeromagnetic maps. The most recent and comprehensive study of the Eastern Gabbro was conducted by Shaw (1994) who noted that more than 90% of the Eastern Gabbro is comprised of layered gabbro. Layering shows distinct variations in style, is usually parallel to the eastern contact of the gabbro, and dips 20°to 60°toward the centre of the Coldwell complex (Shaw, 1994).

At this stop, layering strikes approximately north and dips west towards the rest of the complex at about 45°. A thickly layered sequence is underlain by massive gabbro (STOP 4-8) near the contact with the Archean country rocks. The macrorhythmic layering is laterally discontinuous, pinching out over distances of 5 to 10 m; contacts are sharp and conformable (Shaw, 1994). Rhythmic layering has been related to variation in the respective proportions of plagioclase (An_{60-35}), augite (Fo_{67-43}), minor orthopyroxene (En_{55-66}), and Fe-Ti-oxides by Lum (1973). Modal plagioclase varies from approximately 60% to 80% in the leucocratic layers and 20% to 35% in the meso- to melanocratic layers (Shaw, 1994). A second band of layered gabbro, separated from the first by massive gabbro, is exposed on top of the long rock cut. Here, the microrhythmic layering produces relatively thin (1 to 5 cm) to medium thick (5 to 100 cm) layers that can be traced for over 35 m along strike. Layer contacts are sharp, locally scalloped, and conformable. Trough cross-bedding has been noted on vertical faces by Shaw (1994).

This stop is also close to the contact between the Eastern Gabbro and Fe-rich augite syenite. Locally pegmatitic syenite dikes intrude the gabbro at this locality and contain miarolitic cavities. McLaughlin (1990) has reported the presence of a variety of REE-bearing fluorocarbonates (bastnaesite, parisite, synchisite), Nb-bearing phases and zircon in pegmatitic syenite with quartz, feldspar and sodic amphibole.

STOP 4-8: RHEOMORPHIC BRECCIA, EASTERN CONTACT, COLDWELL ALKALIC COMPLEX

Location: Lat: 48°43' 10" N Long: 86°19' 30" W; 37.3 km east of the Dead Horse Road turn-off, 3.7 km east of the Marathon (Highway 626) turn-off.

Duration: 15 minutes

Description: Approximately 1.7 km farther east along the highway, the gabbro is in contact with and contains numerous xenoliths of Archean metasedimentary rocks. This has produced hybrid and contaminated phases and rheomorphic breccia. Disseminated Fe- and Cu-sulfides occur in biotite-rich gabbro at this site (Dunlop Occurrence), which has experienced sporadic base metal exploration. Approximately 10 km north, Fleck Resources Ltd. has outlined potential reserves of 37 mT grading 0.31% Cu, 0.04% Ni, 0.0012 opT Rh, 0.0068 opT Pt, 0.0271 opT Pd, and 0.0023 opT Au in the coarse-grained to pegmatitic Two Duck Lake phase of the Eastern Gabbro (Canadian Mines Handbook, 1994-1995, p.154).

Shaw (1994) has documented a number of occurrences of rheomorphic intrusion breccia associated with the Eastern Gabbro. Pods of breccia vary from 20 to 75 m in width and are up to 250 m in lenght. The breccia exposed along Highway 17 at this site comprises mainly Archean clastic metasedimentary rocks, magnetite-bearing iron formation, mafic metavolcanic and massive vein quartz. The breccia varies from clast- to matrix supported; the matrix consists of quartz, feldspar, and minor biotite. Rounded to angular clasts range in size from 0.5 to over 100 cm and locally have developed 1 to 2 cm wide, chlorite-rich reaction rims that are thickest where they are matrix-supported (Shaw, 1994). Tourmaline and prehnite overgrowths have been noted and quartz-feldspar-tourmaline veins cut both matrix and clasts. Amethystine quartz, calcite, and molybdenite occur in vugs within more felsic pods in the chaotic contact zone. Quartzofeldspathic rinds and crosscutting veinlets have been interpreted to be the result of partial melting of the felsic material during assimilation. The association between rheomorphic breccia and the Eastern Gabbro suggests a mutual genetic relationship.

STOP 4-9: LAYERED FE-RICH AUGITE SYENITE, COLDWELL ALKALIC COMPLEX

Location: Lat: 48°44' 20" N Long: 86°23' 25" W; 680 m west along the shoreline of Lake Superior from the end of the James River Industrial Road along the waterfront in Marathon.

Duration: 1 hour

Description: Broad expanses of glacially polished and wave-washed, massive Fe-rich augite syenite occur all along this part of the Lake Superior shoreline near Marathon. Fresh surfaces vary from dark green-brown to black, despite a buff to white weathered surface. Although this unit is typically massive, rhythmic to chaotic layering is locally developed and commonly dips shallowly toward the centre of the complex. At this site, layering strikes at 070° and dips north at 60° The layering is unusual in that it is defined by an intercumulus mineral (augite) rather than by cumulus phases (feldspar).

Fe-rich augite syenite (formerly referred to as ferroaugite syenite) composes a large portion of the exposure in the eastern half of the Coldwell complex. It appears to be a sheetlike intrusion that dips approximately 15° toward the centre of the complex, sandwiched between underlying Eastern Gabbro and overlying, recrystallized amphibole-quartz syenite; it also intrudes the basaltic roof pendants (Walker and others, 1992; 1993a). Crystallization of the syenite inwards from its upper and lower contacts produced mineralogical and compositional variations across it (Walker and others, 1993a). Constituent minerals include iridescent, lathlike, cryptoperthitic feldspar, up to 30% interstitial, Fe-rich augite, and variable amounts of fayalite, amphibole, aenigmatite, and rare quartz. Coarse-grained to pegmatitic portions of the syenite host a variety of REE-bearing fluor-carbonates, quartz, chalcedony, and molybdenite. Iridescent feldspar, known locally as "spectrolite," is commercially extracted on a small scale from pegmatite at Shack Lake near Marathon.

FIELD TRIP 3
DAY 5A
(morning)

LOGAN SILLS IN THE NIPIGON BAY AREA

Mark Smyk
Ontario Geological Survey

STOP 5-1: SIBLEY GROUP AND LOGAN SILLS AT KAMA HILL

Location: Lat: 49° 00' 00" N Long: 88° 01' 30"W; Highway 17, 21.9 - 22.9 km east of Nipigon.

Duration: 45 minutes

Description: This stop features a cross section through Middle Proterozoic sedimentary rocks of the Sibley Group as exposed in a large, highway rock cut. These rocks unconformably overlie Neoarchean migmatitic and granitoid rocks of the Quetico subprovince at this locality. The unconformity lies near the highway level at the base of Kama Hill, west of this stop and is exposed on the shore of Kama Bay.

This section exposed here includes the Rossport and Kama Hill Formations, the two uppermost Sibley Group formations. The Sibley Group has yielded Rb-Sr isochron ages of 1294 $\pm$ 33 Ma (Franklin *in* Wanless and Loveridge, 1978) and 1339 $\pm$ 31 Ma (Franklin and others, 1980). These ages are bracketed by U-Pb ages of nonconformably underlying, anorogenic granite (1537 + 10/-2.3 Ma), disconformably overlying, Keweenawan Osler Group rhyolites (1098 $\pm$ 4 Ma), and crosscutting Logan diabase sills (1109 +4/-2 Ma) (Davis and Sutcliffe, 1985; Sutcliffe, 1991). Intercalation of ca. 1537 Ma volcanic fragmental rocks and quartz arenite in northern Lake Nipigon has prompted Sutcliffe (1991) to suggest that Sibley Group sedimentation may have been initiated much earlier than previously thought.

The rocks exposed at highway level are interbedded, buff dolomicrite and red, dolomicritic mudstone of the Channel Island Member (Rossport Formation). They are interpreted by Cheadle (1986) to be the result of fluctuating water levels in a playa lake, proximal to an alluvial floodplain. These rocks are overlain by the Middlebrun Bay Member, a thin (80 to 140 cm), but laterally extensive chert-carbonate unit that is characterized by cryptalgal laminites and hemispherical, stromatolitic growth structures. The uppermost part of the Rossport Formation, the Fire Hill Member, consists of purplish-red to orange to buff, dolomicritic mudstone. The overlying Kama Hill Formation consists of purple, hematitic shale and buff, arkosic siltstone attributed to subaerial, floodplain deposition.

The sedimentary rocks, which dip southward at about 10°, are intruded by a number of stacked, Logan diabase sills that vary in thickness from 1 to 100 m. The columnar-jointed, erosion-resistant sills form the tops of the mesas and cuestas in the region. Features such as asymmetric folds and boudinage adjacent to the sills, ascribed in part to synsedimentary deformation (Franklin and others, 1982), have more recently been interpreted by Antonellini and Cambray (1992) as induced by shear strain (Fig. 3.27). They have described an early deformation phase that produced minor thrusts and mesoscopic "shear bands" [C' shears?] that suggest dextral or tops-to-the-south movement, parallel to bedding planes. These thrusts are confined to a single stratum and are in association with southwest-verging, asymmetric folds and tension gashes. These structures are well developed in coarser, buff-coloured Channel Island Member near the highway level. Extensive brecciation has occurred in overlying chert-carbonates. A second deformational event involved the rotation of the shear bands in the opposite (sinistral) sense and is localized at the basal contact of this lower sill. These observations support a model in which sills take advantage of bedding-parallel shear zones that are the surface expression of a south-dipping, steplike detachment plane related to extension during Midcontinent rifting. The second deformational phase may be the result of shear stress induced by the intrusion of diabase magma from the south.

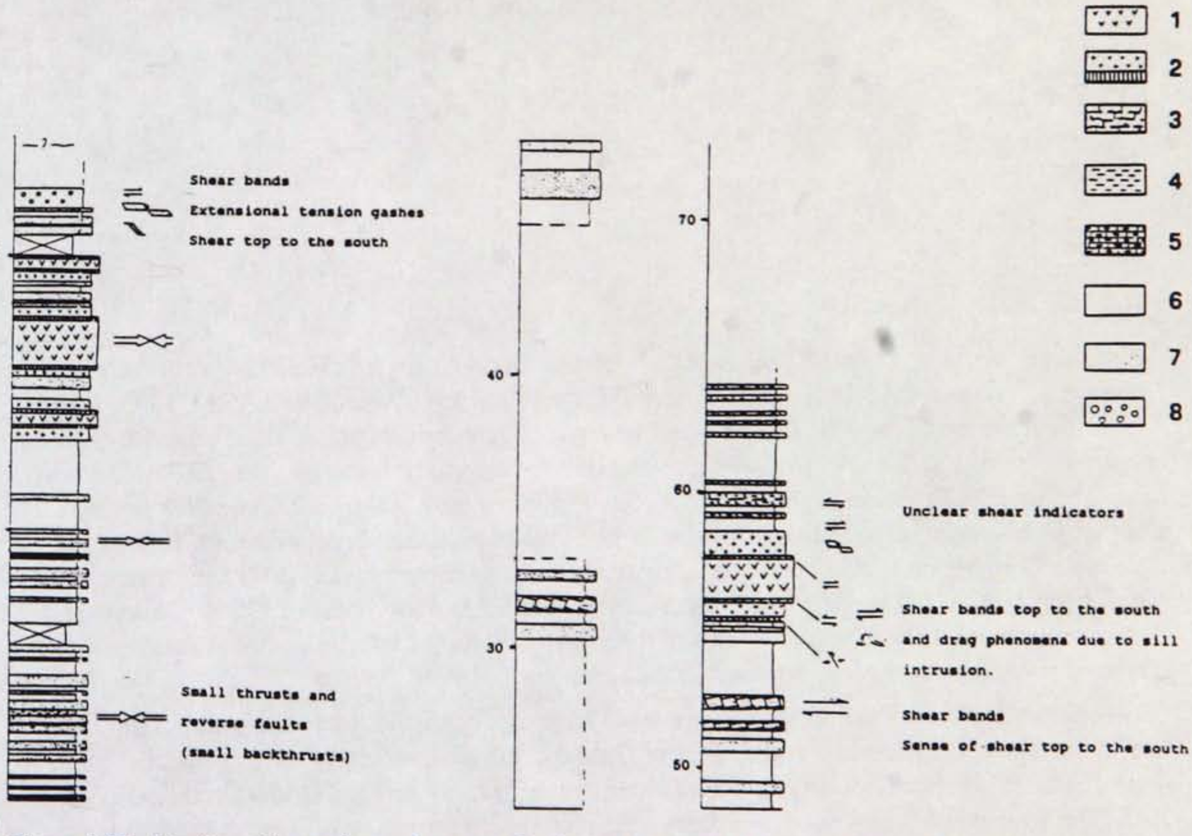


Figure 3.27: Stratigraphic section and structural relations at Kama Hill. The distribution and characteristics of the strain indicators are represented at the side of the stratigraphic column. 1 - diabase sill; 2 - chilled margin; 3 - sandy dolostone in clay and feldspar matrix; 4 - massive dolostone rich in clay and feldspar; 5 - central chert member; 6 - dolostone; 7 - quartz arenite; 8 - reduction spots. From Antonelli and Cambray (1992).

FIELD TRIP 3

DAY 5B

(Afternoon)

VOLCANIC ROCKS OF THE NORTH SHORE VOLCANIC GROUP, NORTHEASTERN MINNESOTA

John C. Green

University of Minnesota—Duluth

Overview

Volcanic rocks and interbedded sediments of the Midcontinent rift (MCR) in northeastern Minnesota compose the North Shore Volcanic Group (NSVG) (Green, 1972, 1982, 1983; BVSP, 1981). Correlation of the NSVG with other volcanic sequences around the Lake Superior basin is shown in Figure I.3. The NSVG includes parts of two major lava plateaus, separated by a hiatus of approximately 7 million years (Figure 3.28). The lower part, which is 2.6 km thick in Cook County, is reversely polarized and includes some of the oldest dated lavas in the MCR (1108 Ma). The upper part, which is over 9 km thick and is normally polarized, yields U/Pb zircon ages that range from 1100 Ma at its base to 1096.6 Ma near its top (Davis and others, 1995). This younger date overlaps slightly with the lower part of the exposed Portage Lake Volcanics of Michigan, and a tentative correlation has been proposed (Miller and others, 1992).

Structurally, the NSVG occurs as the eroded edge of a gently SE-tilted basin extending from Grand Portage at the northeast end to west of Duluth at the southwest end. Dips generally range from 10° to 25°, with a N-S strike at Duluth and a roughly E-W strike at Grand Portage (Figure 3.1). The stratigraphic sequence rises from each end toward the top near Tofte in southwestern Cook County, where the lavas strike parallel to the shoreline. Attitudes are disturbed locally near subvolcanic intrusions. At each end the basal lavas conformably overlie a thin quartz arenite, which in turn unconformably (to the southwest) or disconformably (to the northeast) overlies Lower Proterozoic metasediments of the Animikie Group.

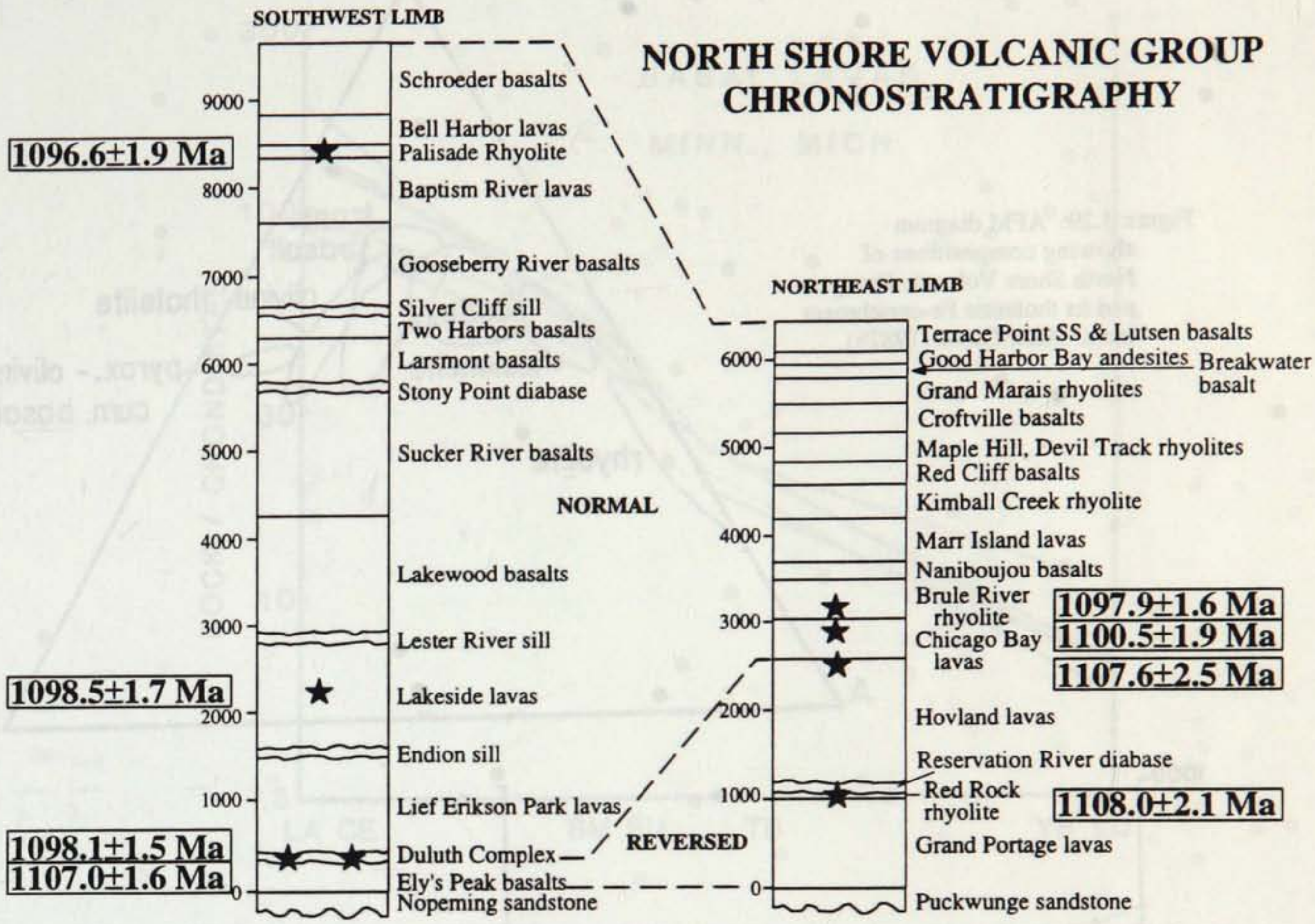


Figure 3.28: Generalized chronostratigraphy of the North Shore Volcanic Group. Dates from Davis and others (1995).

The volcanic edifice of the NSVG has been intruded at its base by the Duluth Complex (See Day 1) and at higher levels by the Beaver Bay Complex and other hypabyssal and subvolcanic bodies (see Day 2). At the southwest end of the NSVG, the Duluth Complex has intruded between a wedge of reversely polarized lavas (Ely's Peak basalts) and the much thicker, overlying, normally polarized sequence (Fig. 3.3). Duluth Complex intrusions occupy a similar horizon at the northeastern end of the area in Cook County (Fig. 3.1).

Detailed mapping along the lakeshore and tributary streams and inland reconnaissance has established a stratigraphy for both the "southwest limb" and the "northeast limb" (Figure 3.28). A wide range of compositions is represented, with a distinct bimodal character and a typical Fe-enriched tholeiitic trend on variation diagrams (Figure 3.29). Olivine tholeiite is the most abundant, followed by transitional basalts. The most primitive olivine tholeiites are high-Al basalts with mg#'s of 65-68. Representative analyses are given in Table 3.5. Trace elements show the basalts to be of undepleted, plume character (Figure 3.30).

The NSVG is atypical among the several MCR volcanic sequences in containing an exceptionally large proportion of felsic rocks: about 10% in the southwest limb and about 25% in the northeast limb. This may be related to the fact that it is the only MCR sequence that is underlain almost entirely by large mafic plutons (the Duluth Complex).

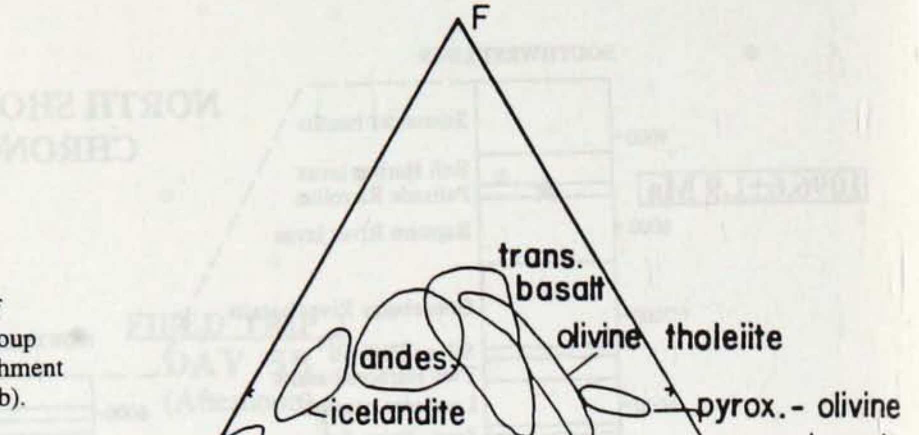


Figure 3.29: AFM diagram showing compositions of North Shore Volcanic Group and its tholeiitic Fe-enrichment trend. After Green (1982b).

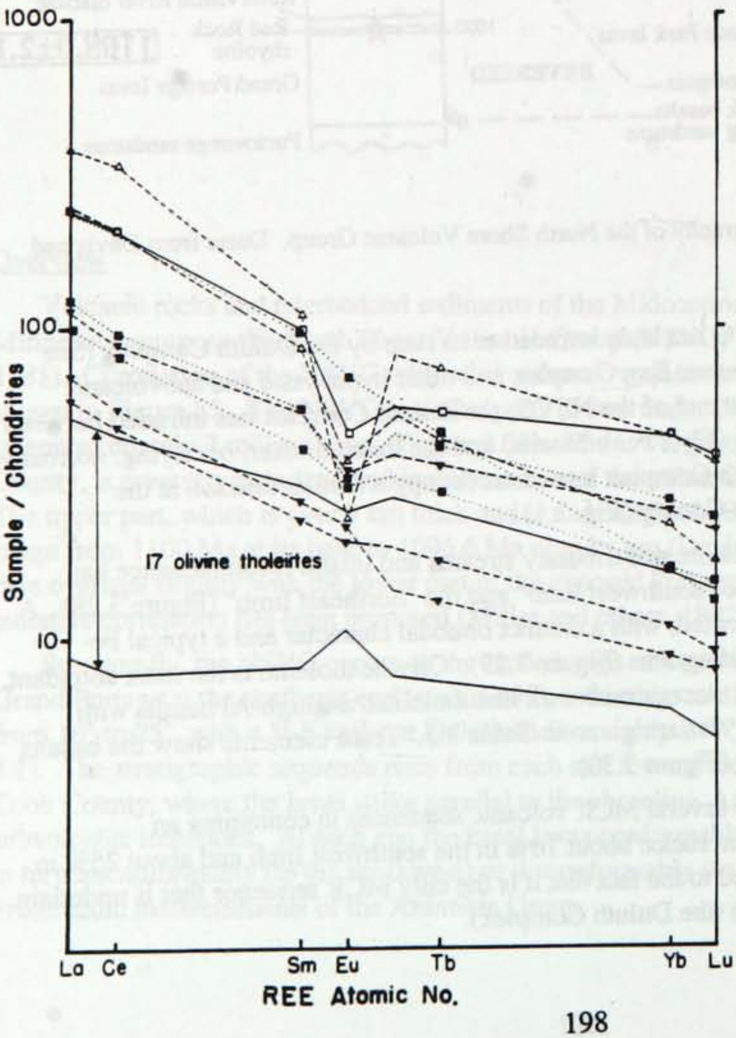


Figure 3.30: Chondrite-normalized REE diagram for North Shore Volcanic Group lava flows After Green (1982b).

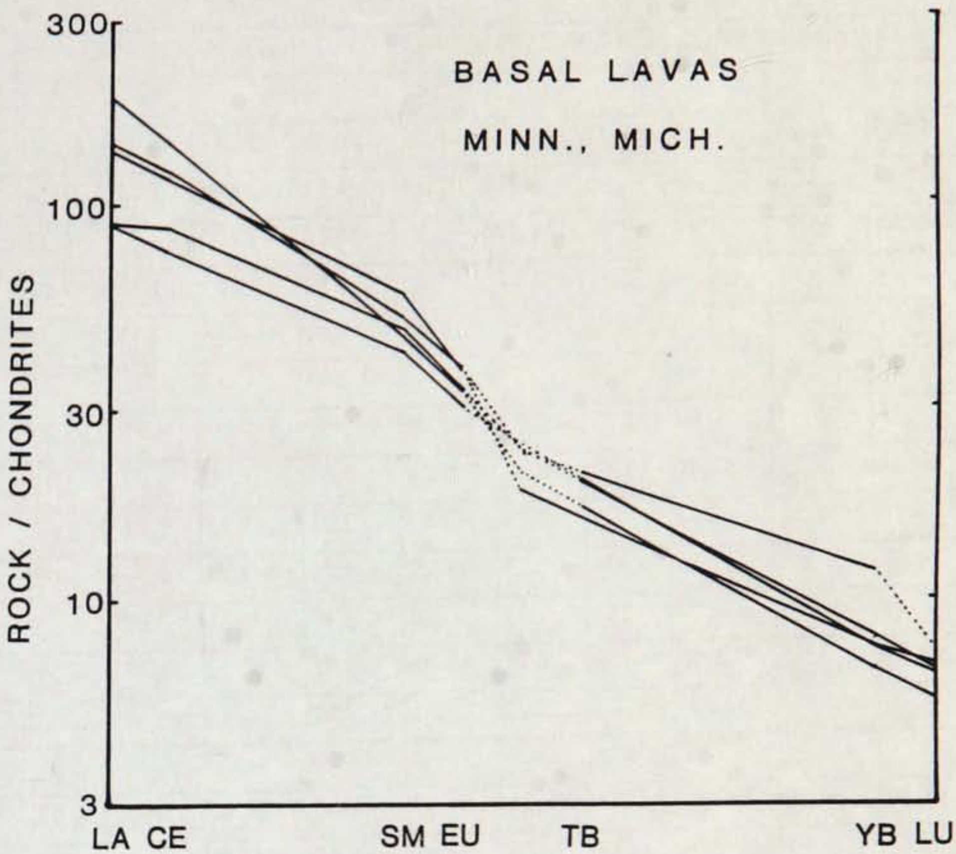


Figure 3.31: Chondrite-normalized REE diagram for basal, augite-phyric lavas of the North Shore Volcanic Group and Powder Mill Group, western Lake Superior.

Stratigraphic geochemistry shows a rather random intermixture of lava compositions, with three exceptions.

1) At the base, at both Grand Portage and Duluth (as well as northern Wisconsin), are basalts that have a unique geochemical and petrographic character, with augite phenocrysts and a steep chondrite-normalized REE pattern (Figure 3.31; Nicholson and others, 1991; 1995).

2) A 3.5-km km sequence between Duluth and Two Harbors (Brannon, 1984) gradually trends from most-evolved rhyolite, icelandite, and ferroandesite to primitive olivine tholeiite with stratigraphic height.

3) At Grand Portage, the lowest kilometer of lavas shows the opposite trend, from most primitive at the base to rhyolite at the top (Green and others, 1980; Green, 1995).

Trace-element ratios and Rb/Sr, Nd/Sm, and U/Pb isotopic systems (Table 3.6) indicate that NSVG basalts, particularly the olivine tholeiites, contain little or no crustal component, but reflect a mantle-plume source (BVSP, 1981; Nicholson and others, 1991; Nicholson and Green, 1995). Only some basal flows at Grand Portage suggest a possible Archean crustal influence on U/Pb relations. Rhyolites, on the other hand, have initial e_{Nd} values that range from -4 to -14, indicating a moderate to large contribution from crustal melting (Vervoort and Green, 1995).

TABLE 3.5: Chemical Analyses of Representative NSVG Lavas and Associated Dikes

Sample	GP-17	FR-10	SR-10	GH-25	KR-20	LW-21	TH-2	H-5b	SR-16	LW-27	MI-2	F-201	H-4
SiO_2	46.63	46.64	46.91	47.19	48.52	49.54	50.19	53.22	55.20	58.50	62.60	74.41	57.31
TiO_2	2.47	2.33	.89	.95	.72	2.14	1.51	2.05	1.61	1.54	1.09	.24	1.62
Al_2O_3	9.65	15.21	17.65	17.04	17.57	15.05	15.15	17.55	13.51	10.12	12.01	10.95	13.39
Fe_2O_3	5.34	9.49	6.43	2.63	2.18	10.05	5.51	4.63	4.16	8.25	8.18	1.64	4.23
FeO	8.34	4.86	3.38	7.69	6.10	3.23	5.82	6.18	6.43	6.33	2.02	2.91	8.05
MnO	.29	.14	.16	.14	.13	.17	.15	.13	.16	.30	.12	.05	.17
MgO	8.25	5.68	8.04	8.11	8.65	5.05	5.91	2.33	4.39	1.27	1.40	.30	1.89
CaO	10.38	9.21	11.46	10.76	10.83	8.38	9.13	4.79	5.53	3.59	2.22	.50	3.61
Na_2O	2.44	2.56	2.22	2.23	2.23	2.67	2.71	5.35	3.74	2.42	4.04	1.93	3.88
K_2O	.86	.54	.17	.35	.17	1.29	.62	1.76	1.90	3.69	4.15	5.64	2.66
P_2O_5	.26	.25	.04	.13	.03	.29	.17	.51	.22	.46	.28	.04	0.59
H_2O	3.14	3.05	2.98	2.55	2.25	2.53	1.92	1.32	2.75	2.50	1.72	.79	2.05
CO_2				.07			.13	.11			.04	.26	
Total	98.05	99.56	100.33	99.84	99.38	100.39	98.92	99.93	99.60	98.97	99.87	99.66	99.45
Mg	.55	.43	.61	.59	.66	.42	.49	.29	.43	.14	.23	.11	.22
Rb	17.2	13	12.4	4	4	50	8	54	19	125			
Sr	780	199	281	265	213	235	184	973	249	157			
Ni	170	170	249	220	240	90	110	n.d.	n.d.	n.d.			
Cr	611	154	307	292	357	207	124	1	48	6	11	12	
La	45.9	15.8	4.5	7.82	2.94	19.4	16.1	68	41.5	74.1	81.5	128	
Yb	1.7	3.20	1.4	1.70	1.15	3.6	3.24	2.9	5.5	11.9	9.0	9.1	
Th	3.7	1.83	0.4	0.84	n.d.	2.3	2.8	7.1	5.9	10.8	11.6	16.4	
Zr				69			150	412					
Hf	6.9	4.6	1.7	2.01	1.14	5.1	4.5	10.9	7.8	19.5	19.0	16.1	
Ta	2.7	.9	.5	.4	n.d.	1.0	.6	2.3	1.2	3.0	3.3	2.6	

Sample identification. KEW - numbers are more fully described in BVSP, 1991. All but last two are lava flows.

GP-17: **transitional basalt near base of section, Grand Portage Island.**

FR-10: **Fe-rich olivine tholeiite, near French River, NE of Duluth, KEW-2.**

SR-10: **olivine tholeiite, Gooseberry Falls.**

GH-25: **olivine tholeiite; Terrace Point basalt flow, near Good Harbor Bay, KEW-3.**

KR-20: **primitive olivine tholeiite, Knife River area NE of Duluth, KEW-6.**

LW-21: **transitional basalt, Lakewood quadrangle NE of Duluth, KEW-12.**

TH-2: **quartz tholeiite, Two Harbors, KEW-15.**

H-5b: **plagioclase-phyric andesite, SW of Reservation River, Cook County, KEW-17.**

SR-16: **aphyric andesite, near Split Rock River NE of Two Harbors, KEW-20.**

LW-27: **ferroandesite, Lakewood quad NE of Duluth, KEW-18.**

MI-2: **icelandite, Marr Island quad SW of Brule River, Cook County.**

F-201: **porphyritic rhyolite of Palisade flow (rheoignimbrite), Illgen City, Lake County.**

H-4: **icelandite/andesite dike, same locality as H-5b.**

Table 3.6: Initial Isotope Ratios of NSVG Olivine Tholeiites

$^{87}Sr/^{86}Sr$	ε_{Nd}	$^{238}U/^{204}Pb$
.7035-.7059	≈0	≈8.2

As in the other volcanic plateaus of the MCR, fissure eruptions predominated in the NSVG, producing flood lavas and minor low shields (Green, 1989). Several basaltic dike swarms, intruded during both reversed and normal-polarity intervals and probably feeders to these flows, are found around the Lake Superior basin (Green and others, 1987). Most basaltic flow surfaces are of pahoehoe type (smooth, billowy, or ropy), with flow breccia (*aa*) surfaces on some of the more siliceous basalts and andesites. Nearly all flow bases were fluid. A 188-flow sequence from Duluth to Two Harbors has a mean flow thickness of 25 m and a median of 15 m. The icelandites and rhyolites were emplaced mostly as very large, hot, fluid lavas and rheoignimbrites; the largest rhyolites had volumes estimated to be up to 600 km^3, with exposed strike lengths of 32-40 km (Green and Fitz, 1993).

Field Stop Descriptions

STOP 5-2: H OVLAND PORPHYRITIC ANDESITE, NSVG, AND ICELANDITE DIKE

Location: Pull off on Highway 61 west of Reservation River, ~12 WSW of Grand Portage, Hovland 7.5' quadrangle (T62N, R4E, Sec. 12, NW).

Duration: 1 hour, lunch stop.

Description: At this site an abandoned Nipissing beach cliff exposes a remarkable porphyritic basaltic andesite flow of the reversed polarity sequence. It contains tabular calcic andesine phenocrysts up to 10 cm across (along with broken chips thereof) that form a distinct flow structure in this lava. Flows containing tabular plagioclase phenocrysts, commonly glomeroporphyritic, are common only in the reversed-polarity plateaus around the basin (especially here and in the Powder Mill Volcanics, Michigan-Wisconsin). The magma for this flow had apparently accumulated plagioclase in a subvolcanic chamber (by floating?), as shown by the flow's high Al content and positive Eu anomaly. For a chemical analysis (H-5b) see Table 3.5.

Near the highway the porphyritic andesite is cut by a porphyritic icelandite/ferroandesite dike that can be followed for about 2 miles (3.2 km) roughly parallel to the shore. It also has reversed polarity and could have fed a higher, now-eroded flow. It contains small phenocrysts of ferroaugite and plagioclase. A chemical analysis (H-4) is given in Table 3.5.

STOP 5-3: OLIVINE THOLEIITE FLOW, INTERFLOW SEDIMENTARY ROCKS, NSVG

Location: Turnout on Highway 61 overlooking Good Harbor Bay, 4.8 miles southwest of Grand Marais, Good Harbor Bay 7.5' quadrangle (T61N, R1W, Sec.).

Duration: 20 minutes.

Description: In this large roadcut, the Terrace Point basalt flow, one of the major cliff-formers of the "Sawtooth Range," overlies a thick (40 m) section of interflow sedimentary rock. The flow is approximately 50 m thick, and can be traced inland (to the southwest) for at least 23 km. The shoal out in the lake to the east (Gull Rock) is also made of this flow. For a chemical analysis (GH-25) see Table 3.5.

The Terrace Point basalt is predominantly a massive, fine-grained, black, sparsely porphyritic, olivine tholeiite that characteristically contains thomsonite in amygdules. In its lengthy exposure (including this cut) several flow units and breccia zones of various character show complex relationships with the massive, basal portion. The relations here suggest that a small vent, coming through saturated lakebed sediments, built a small pile of hyaloclastite and pillow breccia; massive, fluid flood basalt then erupted between the breccia and the sediments (Fig. 3.32).

The interflow sedimentary rock in this cut are mainly thin-bedded siltstone and silty shale, but they overlie a thick sequence of medium-grained, well-sorted, arkosic-lithic (volcanic) sandstone of intrabasin derivation showing abundant ripples (Jirsa, 1984). They appear to be lacustrine, deposited in a large depression on the irregular surface of a sequence of basaltic andesites that form the ledges on the shoreline across the bay. This sedimentary unit is traceable for 11 km along strike to the west-southwest.

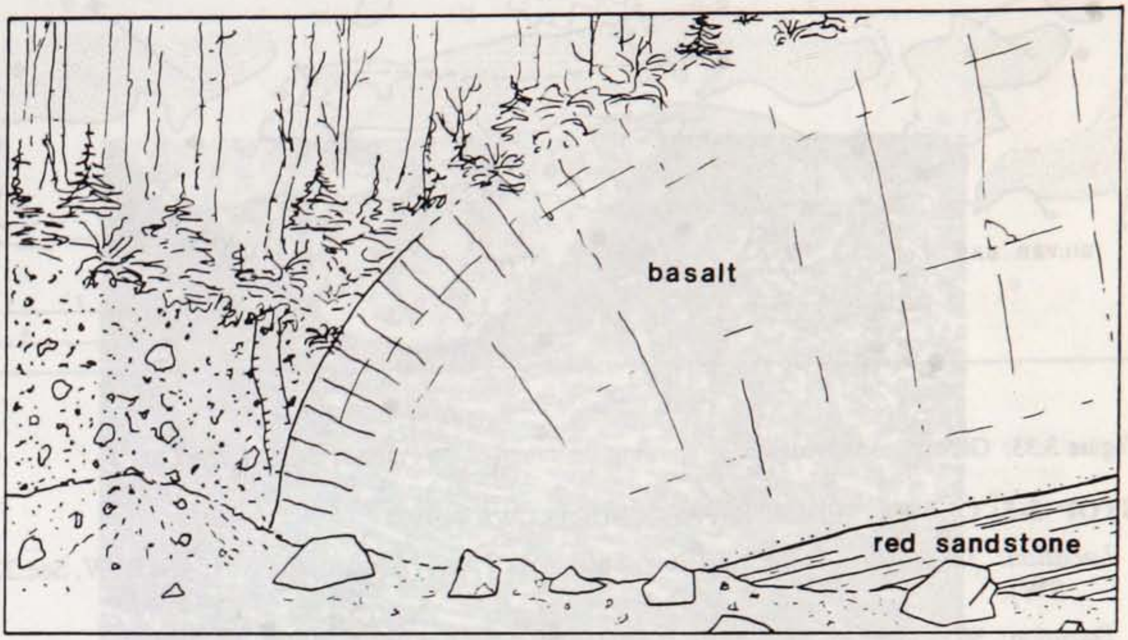


Figure 3.32: Sketch of relationships between Terrace Point basalt flow, lapilli-tuff and breccia, and red sediments in roadcut, Good Harbor Bay. Pyroclastic rocks interpreted as hyaloclastic cone; note small dike above flow surface. From Green (1989).

STOP 5-4: PALISADE RHYOLITE, NSVG

Location: Shovel Point, Tettegouche State Park, Illgen City 7.5' quadrangle (T56N, R7W, Sec. 14).

Duration: 45 minutes.

Description: This long, dip-sloping point is made of a large, porphyritic rhyolite flow characterized by quartz, sanidine, oxidized prismatic Fe-silicate, magnetite, and zircon phenocrysts. It is at least 90 m thick, and can be traced along strike for at least 23 km (Fig. 3.33). The prominent hill to the southwest (Palisade Head) is part of the same flow. It is estimated to have had an original volume of at least 20 and possibly over 100 km^3 (Green and Fitz, 1993).

Although nearly all of this flow looks like a lava flow, critical but unfortunately inaccessible exposures at the base of the Palisade Head cliff show that it is a rheoignimbrite. Its basal few decimeters are devitrified welded tuff with perlitic cracks (Figure 3.34). In less that a meter upward it becomes highly lineated, with the shards smeared out into a fine flow lineation and lamination. Farther upward into the flow the fine lamination becomes degraded as the rock becomes massive and crystalline. This interior, however, in many places shows a flow (pilotaxitic) texture of subparallel tabular quartz paramorphs after tridymite, which show that tridymite was crystallizing from the "reconstituted" rhyolite lava as it underwent rheomorphic flow.

Outcrops of the top facies of the Palisade rhyolite, along Highway 1 north of Illgen City, are flow breccias made of laminated, devitrified welded tuff in which shards are discernible. Thus whereas the top and base were welded and chilled to a glass, the interior retained sufficient heat to flow and to allow tridymite and alkali feldspar to crystallize while flowing, thus eradicating all evidence of its pyroclastic eruption.

A chemical analysis (F-201) is presented in Table 3.5.

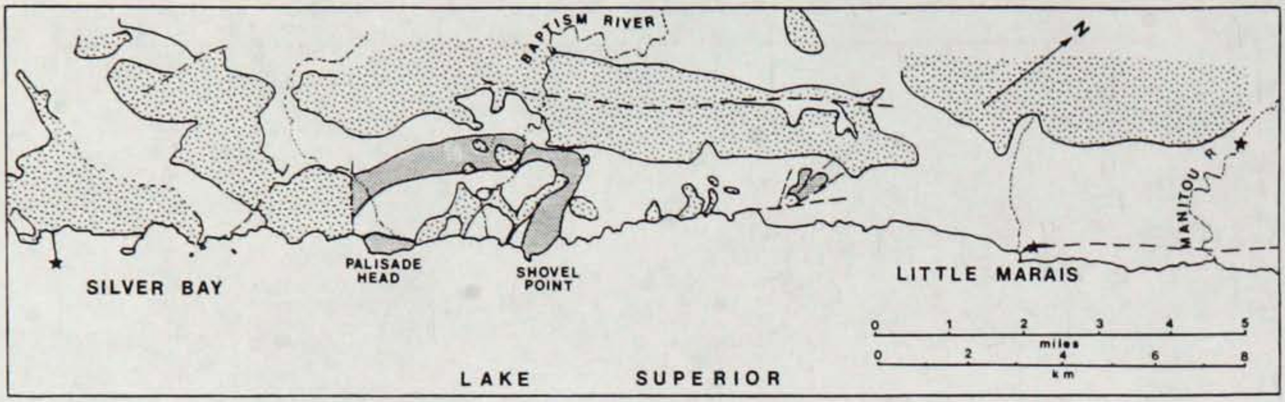


Figure 3.33: Generalized geologic map showing the extent of the Palisade rhyolite flow.

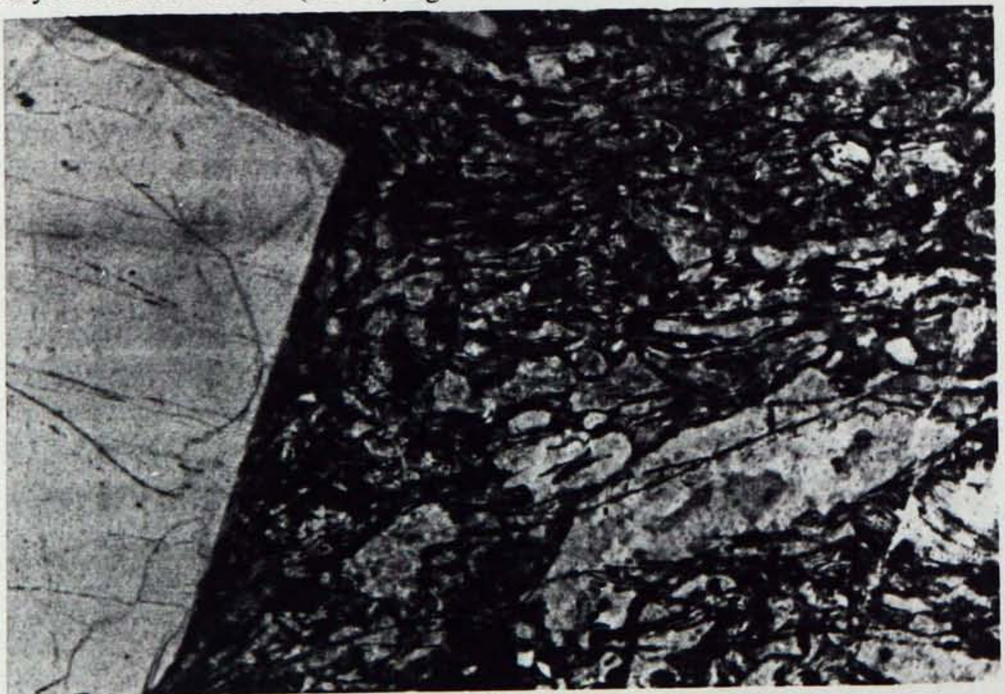
STOP 5-5: OLIVINE THOLEIITE PAHOEHOE FLOWS, NSVG

Location: Gooseberry Falls State Park, Split Rock Point 7.5' quadrangle (T54N, R9W, Sec.22, SW).

Duration: 30 minutes.

Description: Here the Gooseberry River has eroded through a sequence of gently dipping flood basalts which are about 6.4 km above the base of the NSVG (Fig. 3.35). The falls below the highway expose three flows that are characterized by smooth, gently billowing surfaces typical of very fluid lava. The billows have apparent wavelengths of about 2-4 m and relief of a few centimeters. The upper flow has a characteristic columnar, amygdaloidal to massive, ophitic central zone that forms the falls above the highway bridge, as well as the first falls below the bridge, and it overlies the billowy top of the next flow in the middle of this waterfall. The middle flow is about 3 m thick and overlies another smooth, billowed flow top; this contact between the middle and lower flows can be followed along the riverbed downstream (and down dip), where a natural arch has been eroded through it in an island in the center of the channel. The next falls are in the colonnade of this lower flow. These flow surfaces strongly resemble those on some Quaternary Icelandic flood basalts. A chemical analysis of the lower flow (SR-10) is given in Table 3.5.

A



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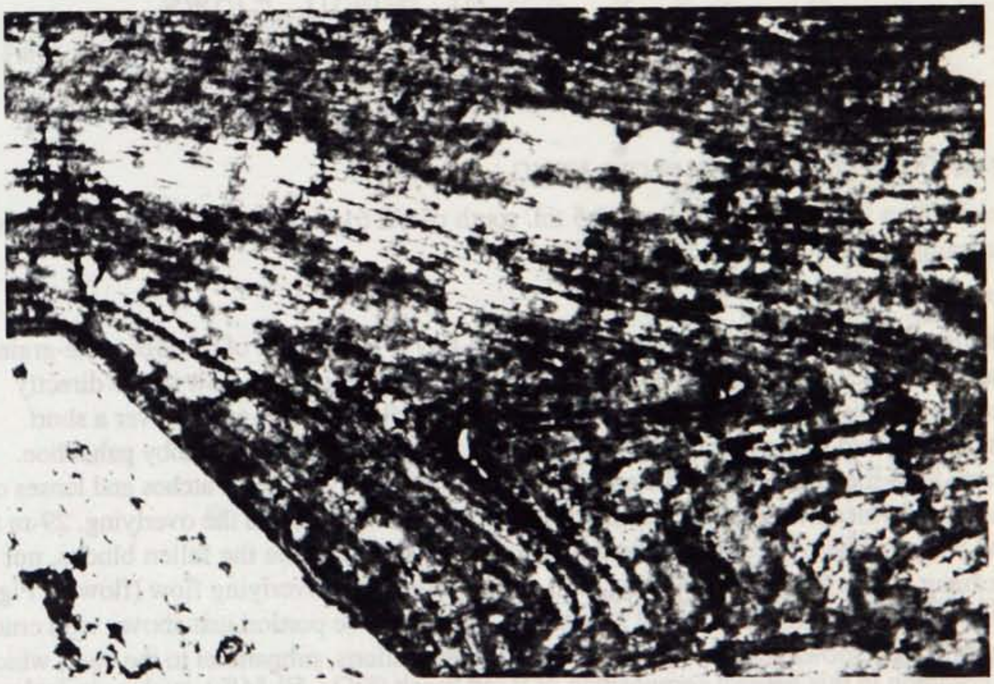

Figure 3.34: Photomicrographs showing welded tuff at base of Palisade rhyolite rheoignimbrite. (a), basal few centimeters; quartz phenocryst on left; most of rock made of compacted, bubble-poor shards. Length of field 2.4 mm. (b) about 60 cm above base; shards strongly deformed next to sanidine phenocryst. Length of field 0.75 mm. (c) about 1 m above base; shards have been smeared out into laminations and folded against sanidine phenocryst. Length of field 2.4 mm.

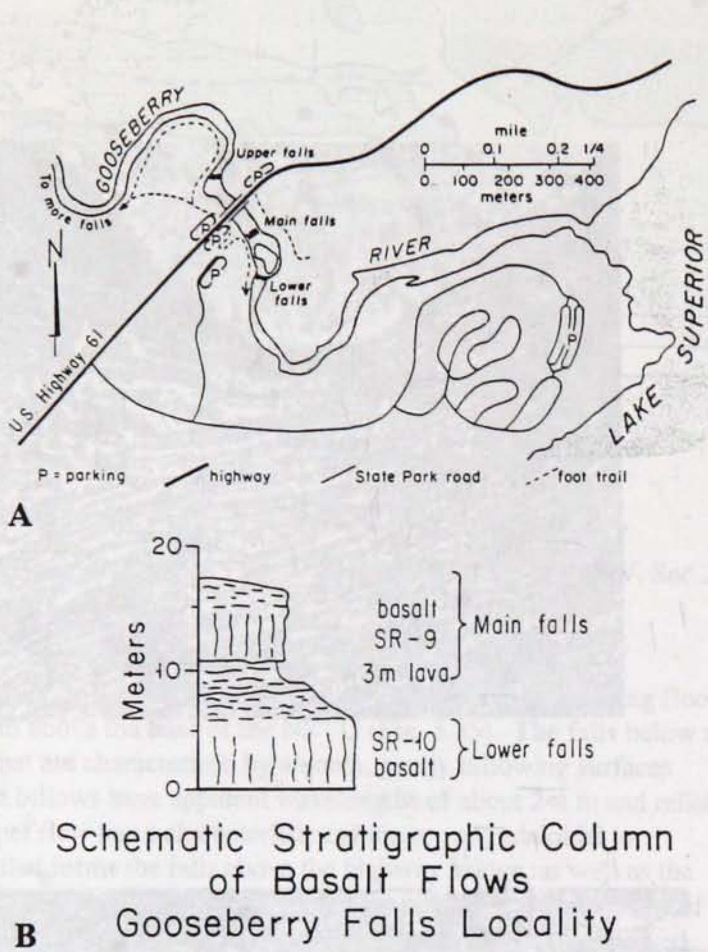

Figure 3.35: Sketch map and stratigraphic column, Gooseberry Falls. From Green (1987).

STOP 5-6: QUARTZ THOLEIITE FLOW, NSVG

Location: Two Harbors Tourist Park, 0.5 mi. south of US 61 on First Street; Two Harbors 7.5' quadrangle (T52N, R10W, Sec.6, NW).

Duration: 20 minutes

Description: Exposed in a seacliff along Burlington Bay is a sequence of thick but fine-grained quartz tholeiite or Fe-tholeiite flows, dipping gently southeast. The flow that is directly accessible (flow A, Fig. 3.36) has a vuggy upper-middle zone that grades over a short distance upward into a highly amygdaloidal top zone, 2-3 m thick, of slabby pahoehoe. Quartz lines the vugs, and calcite and laumontite fill the amygdules. Patches and lenses of friable, red volcanic sandstone occur between the top of flow A and the overlying, 29-m thick flow B in the wave-cut cliff. **Danger: unstable rock!** Examine the fallen blocks, not the overhang. Note the fluid-appearing, non-rubbly base of the overlying flow (flow B, Fig. 3.36), which contains a few small amygdules. The massive portion just above, with crude columns, also shows millimeter-scale oxidation laminations, subparallel to the base, which are common in this general composition of flood basalt (SiO_2 50-54%), such as in Iceland and Northern Ireland (Green, 1989). There are also a few small, lensoid amygdules filled with either black, Fe-rich chlorite or gray agate.

By following a trail along the top of the cliff to the south, one soon comes to a broad, storm-washed ledge made of the upper-middle part of this same flow B. This zone shows hackly fractures and local vugs, and passes up into a thick, rubbly, autobrecciated top zone. Open spaces here are filled with laumontite and calcite. This flow apparently advanced much too fast for rubble to develop at its base (as typical Hawaiian aa flows do), but eventually produced a thick aa top. For a chemical analysis (TH-2) see Table 3.5.

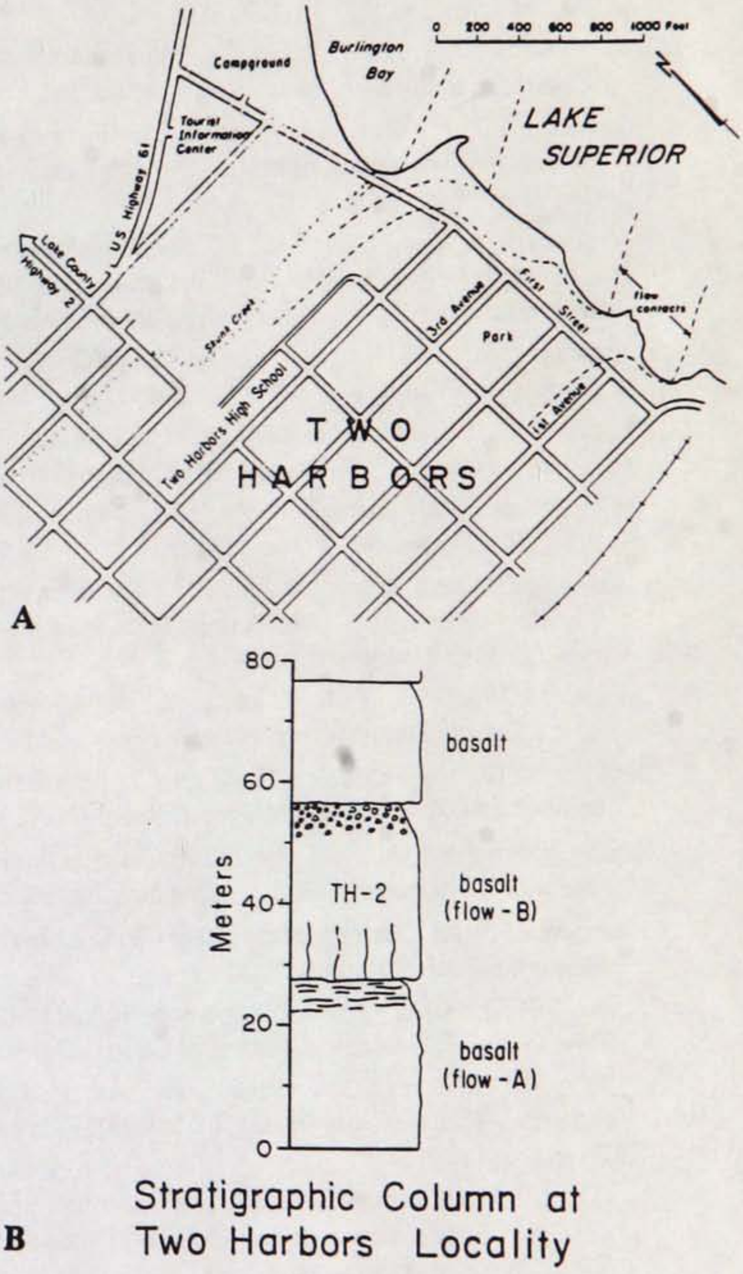

Figure 3.36: Sketch map and column, Two Harbors Tourist Park. From Green, 1987.

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